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Earth and Planetary Science Letters xx (2006) xxx–xxx
www.elsevier.com/locate/epsl

Cenozoic intraplate volcanism on New Zealand: Upwelling induced by lithospheric removal

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Received 21 December 2005; received in revised form 31 May 2006; accepted 1 June 2006

Editor: R.W. Carlson

Abstract

Diffuse intraplate volcanism spanning the Cenozoic on the North, South, Chatham, Auckland, Campbell and Antipodes Islands of New Zealand has produced quartz tholeiitic to basanitic/nephelinitic (including their differentiates) monogenetic volcanic fields and large shield volcanoes. New $^{40}Ar/^{39}Ar$ ages, combined with published age data, show no correlations among age, location or composition of the volcanoes. Continuous volcanism in restricted areas over long time periods, and a lack of volcanic age progressions in the direction and at the rate of plate motion, are inconsistent with a plume origin for the intraplate volcanism. Although localized extension took place during some episodes of volcanic activity, the degree of extension does not correlate with erupted volumes or compositions. Major and trace element data suggest that the silica-poor volcanic rocks (primarily basanites) were derived through low degrees of partial melting at deeper depths than the more silica-rich volcanic rocks (alkali basalts and tholeiites) and that all melts were produced from ocean island basalt (OIB)-type sources, containing garnet pyroxenite or eclogite. The Sr–Nd–Pb isotope data indicate that the silica-poor rocks were derived from high time-integrated U/Pb (HIMU)-type sources and the silica-rich rocks from more enriched mantle (EM)-type sources, reflecting greater interaction with lithosphere modified by subduction beneath Gondwana. The first-order cause of melting is inferred to be decompression melting in the garnet stability field of upwelling asthenosphere, triggered by removal (detachment) of different parts of the subcontinental lithospheric keel throughout the Cenozoic. In some cases, large thicknesses of keel were removed and magmatism extended over many millions of years. Decompression melting beneath a thick craton generates melts that are likely to be similar to those from the base of the mid-ocean-ridge melting column. At mid-ocean ridges, however, these melts never reach the surface in their pure form due to the swamping effect of larger-degree melts formed at shallower depths. Different volcanic styles in part reflect the mode of removal, and size and shape of detached parts of the lithospheric keel. Removal of continental lithospheric mantle could be an important process for explaining the origin of diffuse igneous provinces on continental lithosphere.

Keywords: intraplate volcanism; continental diffuse igneous province; New Zealand; $^{40}Ar/^{39}Ar$ ages; geochemistry; lithospheric removal/detachment

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0012-821X/$ - see front matter © 2006 Published by Elsevier B.V.
doi:10.1016/j.epsl.2006.06.001

1. Introduction

Since the plate tectonic model achieved wide acceptance, intraplate volcanism has been primarily attributed to mantle plumes [1], which now are generally presumed to represent cylindrical regions of mantle upwelling (~100–300 km in diameter) from a thermal boundary layer such as the core/mantle boundary. The mantle plume model is, however, being increasingly questioned, leading to the global "Great Plume Debate" (e.g. http://www.mantleplumes.org). The major alternative to the plume model for intraplate volcanism in continental areas is decompression melting of upwelling shallow mantle that results from tectonic thinning of the lithosphere (e.g. [2]). Neither of these models, however, can adequately explain diffuse volcanism known from many continental areas globally.

New Zealand's South Island has been the site of numerous temporally and spatially dispersed episodes of intraplate volcanism throughout the Cenozoic (Fig. 1). This intraplate volcanism has produced: 1) scattered, low-volume alkalic dikes (e.g. Alpine Dikes) and monogenetic volcanic fields (e.g. Waipiata volcanics), 2) several cubic kilometers of tholeiitic volcanic rocks erupted from

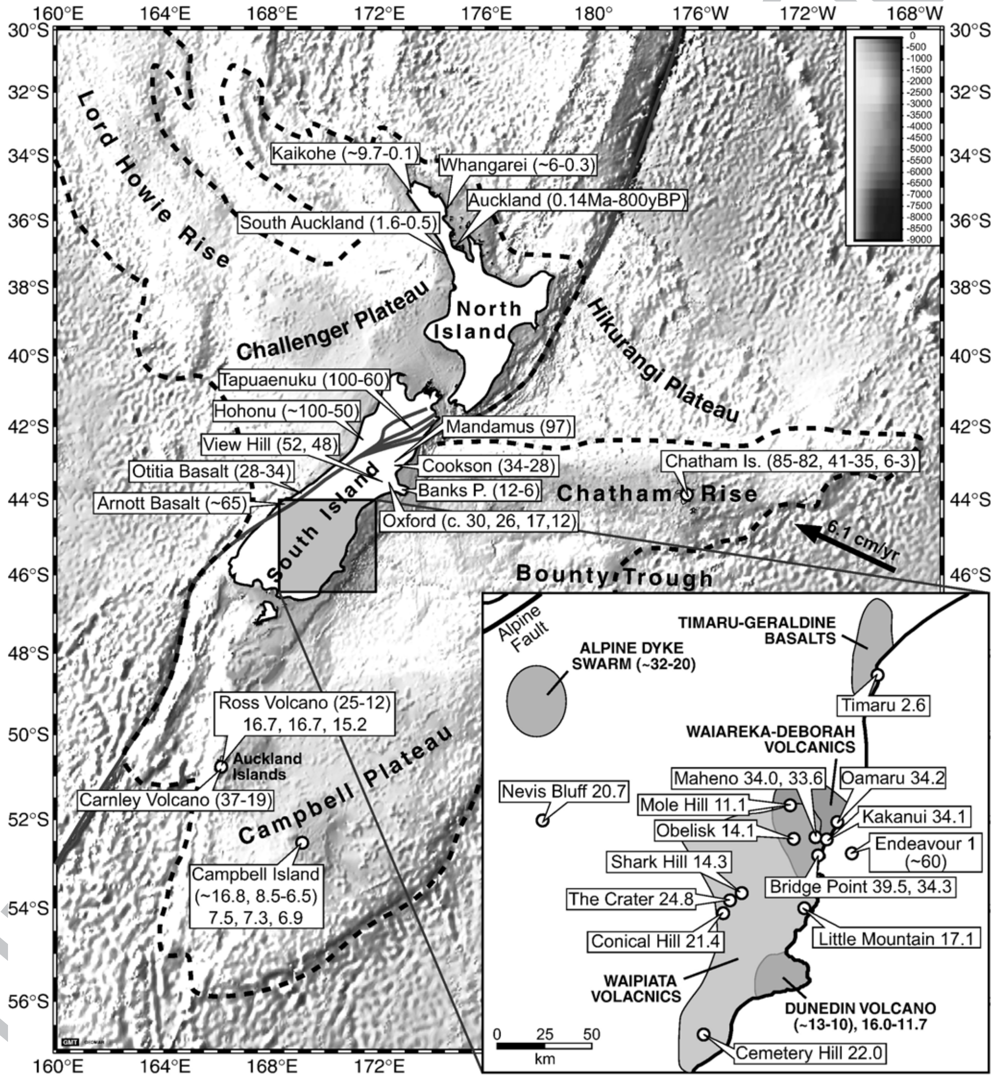

Fig. 1. Overview map of the Zealandia micro-continent, summarizing the published age data. Bathymetry is based on satellite altimetry and ship depth soundings [71]. Shading (see key in upper right hand corner) refers to water depth in meters. Inset: Blow-up map of Otago and southern Canterbury, South Island, showing the $^{40}Ar/^{39}Ar$ age data for intraplate volcanism from this study. Numbers beside the named localities give ages in Ma. Those without parentheses are Ar/Ar ages from this study (Table 1); those within parentheses are K/Ar and other dates from publications referred to in the text. Plate motion vector is from [43].

Table 1
$^{40}Ar/^{39}Ar$ step-heating and single-crystal/particle ages

Sample no.	Dated material	Dating method	Age±2 σ (Ma)	MSWD[a]	% ^{39}Ar in plateau
Otago: Waiareka–Deborah					
LSI7	Matrix	Step heating	39.5±1.8	0.83	69
OU47055	Glass	Step heating	34.2±0.4	0.84	100
	Glass	Single fusions	34.3±0.9	3.31	(N^b=13)
OU54929	Matrix	Step heating	34.0±0.6	0.86	55
KK1	Hornblende	Step heating	33.7±0.3	1.40	87
	Hornblende	Single fusions	34.1±0.1	0.87	(N^b=14)
OU55008	Matrix	Single fusions	33.6±1.8	–	(N^b=2)
MSI99A	Matrix	Step heating	34.3±0.5	1.10	47
Otago: Alpine Dike Outlier					
OU42071	Matrix	Single fusions	20.7±0.4	1.09	(N^b=12)
Otago: Waipiata					
LSI4	Matrix	Step heating	24.8±0.6	0.89	85
OU54935	Matrix	Step heating	22.0±0.3	1.15	67
LSI1	Matrix	Step heating	21.4±0.4	0.92	87
OU66573	Matrix	Step heating	17.1±0.5	0.57	67
	Matrix duplicate	Step heating	17.0±0.5	0.93	84
LSI3	Matrix	Step heating	14.4±0.2	0.53	64
	Matrix duplicate	Step heating	14.2±0.2	0.55	81
OU55018	Feldspar	Step heating	14.1±0.6	1.50	61
MSI76	Matrix	Step heating	11.1±0.6	1.60	72
Otago: Dunedin Volcano					
LSI23	Matrix	Step heating	16.0±0.4	0.66	63
LSI22	Matrix	Step heating	14.6±0.4	0.53	78
OU22855	Matrix	Step heating	13.0±0.6	0.25	67
	Matrix	Single fusions	13.2±0.5	0.95	(N^b=13)
MSI184	Hornblende	Step heating	13.0±0.2	1.04	84
UM peperite	Glass	Step heating	11.7±0.1	0.85	71
Canterbury: Timaru					
OU33298	Glass	Single fusions	2.6±0.4	1.14	(N^b=35)
Campbell Plateau: Auckland Island					
OU19560	Matrix	Step heating	16.7±1.2	0.65	52
OU19577	Matrix	Step heating	16.7±0.1	0.53	99
OU19606	Matrix	Step heating	15.2±0.3	1.30	46
Campbell Plateau: Campbell Island					
OU36173	Matrix	Step heating	7.3±0.3	1.30	70
	Feldspar	Step heating	6.6±0.6	1.50	100
OU36174	Matrix	Step heating	6.94±0.05	1.40	79
OU36176	Feldspar	Single fusion	7.5±0.3	0.67	(N^b=15)

[a] MSWD = Mean Squared Weighted Deviates.
[b] N=Number of Single Fusion Analyses.

spatially and temporally restricted centers (e.g. Timaru and Geraldine lava flows), 3) closely related tholeiitic and alkalic volcanism (e.g. Banks Peninsula [3]), 4) large composite shield volcanoes (e.g. Dunedin and Campbell Island volcanoes) having edifice volumes of up to 1200 km^3, which can occur as clusters (e.g. Auckland Islands and Banks Peninsula shield volcanoes). The cause of this Cenozoic melting, which produced magmas ranging from highly SiO_2-undersaturated (e.g. basanitic and nephelinitic) to quartz tholeiitic and their more evolved differentiates, is poorly understood and controversial. In order to explain the widely dispersed Cenozoic intraplate volcanism on the (mostly submerged) New Zealand micro-continent (Fig. 1), here referred to as Zealandia, with the plume model, many dozens of small plumes (plume swarm) or a diffuse megaplume are required. There is, however, no geophysical (e.g. seismic tomographic) evidence for plume-like structures beneath New Zealand (e.g. [4]). In addition, $^3He/^4He$ isotope ratios of fluid inclusions in mantle xenocrysts and basalt phenocrysts from the South Island appear to support derivation of the intraplate magmas from degassed upper mid-ocean-ridge basalt (MORB)-type mantle beneath New Zealand [5].

Many often contradictory models have been proposed to explain the Cenozoic intraplate volcanism on Zealandia. Major continental rifting associated with the

Table 2
Sr–Nd–Pb isotope data for Cenozoic intraplate volcanic rocks from New Zealand

Sample no.	$^{86}Sr/^{87}Sr$	2 σ	$^{143}Nd/^{144}Nd$	2 σ	$^{206}Pb/^{204}Pb$	2 σ	$^{207}Pb/^{204}Pb$	2 σ	$^{208}Pb/^{204}Pb$	2 σ
Otago: Waiareka–Deborah										
LSI7*	0.703380	4	0.512849	3	18.999	1	15.628	1	38.735	2
OU47055	0.703266	7	0.512844	8	19.118	1	15.617	1	38.763	2
OU54929	0.703210	7	0.512861	6	19.246	5	15.627	4	38.854	9
OU54929					19.248	2	15.636	1	38.888	4
OU55008	0.703419	6	0.512839	9	19.099	2	15.637	2	38.792	4
KKI - kaersutite megacryst	0.702856	7	0.512927	6	19.310	5	15.578	4	38.674	9
KKI - kaersutite megacryst	0.702853	7			19.327	14	15.574	11	38.660	28
Otago: Alpine Dike and Outlier										
OU42071	0.703323	7	0.512936	8	19.675	5	15.651	4	38.810	10
OU42071			0.512925	7	19.662	2	15.643	2	38.795	5
Wilkin5B	0.703326	8	0.512883	5	20.030	1	15.660	0	39.611	1
Otago: Waipiata										
LSI4*	0.702988	5	0.512870	3	20.146	1	15.632	1	39.880	2
OU54926	0.703261	7	0.512887	7	19.253	2	15.618	1	38.935	3
OU54926	0.703273	6			19.252	2	15.615	2	38.926	4
OU54935	0.702931	7	0.512853	5	20.449	1	15.653	1	40.120	3
OU54935	0.702927	8			20.450	2	15.646	1	40.103	4
LSI1	0.702895	8	0.512858	9	20.346	5	15.642	4	40.082	9
LSI1					20.349	3	15.644	3	40.101	7
OU66573*	0.702865	4	0.512916	3	19.970	0	15.647	0	39.442	1
Otago: Dunedin Volcano										
LSI22	0.703032	6	0.512871	8	20.189	3	15.664	2	39.802	5
LSI23*	0.703088	5	0.512879	2	19.764	1	15.639	1	39.368	2
OU22855	0.702834	5	0.512913	5	20.074	1	15.658	1	39.598	3
Canterbury: Timaru										
OU33298	0.703564	8	0.512857	7	18.952	1	15.612	1	38.799	2
OU33298	0.703575	8			18.949	1	15.608	1	38.786	2
OU33278	0.703745	6	0.512829	5	18.923	1	15.619	0	38.791	1
OU33278					18.928	2	15.619	2	38.789	5
Campbell Plateau: Auckland Island										
OU 19560	0.703724	7	0.512832	8	18.850	1	15.619	1	39.007	3
OU 19560	0.703720	8	0.512829	8	18.852	2	15.620	1	39.007	3
OU 19560	0.703715	5			18.850	1	15.618	1	39.005	2
OU 19560	0.703723	7								
OU 19577*	0.703066	3	0.512897	3	19.376	2	15.621	2	39.237	5
OU 19606	0.702933	5	0.512935	5	19.642	2	15.596	2	39.161	4
OU 19606	0.702946	7			19.641	2	15.595	1	39.159	3
OU 19564*	0.703458	2	0.512859	3	18.894	2	15.607	2	39.033	5
Campbell Plateau: Campbell Island										
OU36173	0.703415	7	0.512891	12	19.326	1	15.604	1	39.198	1
OU36173	0.703433	7			19.320	2	15.597	1	39.172	4
OU36174*	0.704995	2	0.512793	3	19.070	1	15.625	1	39.114	1
OU36176*	0.703195	2	0.512910	3	19.517	3	15.602	2	39.268	5
OU36163*	0.704019	2	0.512856	3	19.183	1	15.605	1	39.134	3

Errors are 2 σ within run given to the least significant digit.
* Sr–Nd measured on TRITON TIMS, all others MAT262. For analytical details see Appendix 3.

separation of New Zealand from West Antarctica ceased in the mid-Cretaceous, but Weaver and Smith [2] proposed that Cenozoic intraplate volcanism was at least in part related to shallow upwellings related to local rifting events. In contrast, based on an apparent crude WNW-ESE age progression of volcanism, Adams

[6] and Farrar and Dixon [7] proposed that intraplate volcanism resulted from Zealandia over-riding a former spreading center, manifested as a NNE-trending line of asthenospheric upwelling. Coombs et al. [8] noted the similarity in Sr and Nd isotopic composition of Cenozoic basalts on the South Island of New Zealand to basalts from Australia and Marie Byrd Land, Antarctica. These three continental blocks were contiguous prior to 100 Ma but rifted apart at c. 84 Ma, suggesting that the Cenozoic volcanism resulted from melting of common, and therefore lithospheric, source materials (e.g. [8–10]). Lithospheric sources have been proposed for the nephelinitic to tholeiitic volcanism in the South Auckland volcanic field [11] and on the Chatham, Campbell and Antipodes Islands [12]. In contrast, Finn et al. [4] included the Cenozoic intraplate volcanism on New Zealand within a “Diffuse Alkaline Magmatic Province”, encompassing the easternmost part of the Indo-Australian Plate, West Antarctica and the southwest portion of the Pacific Plate, and attributed this magmatism to interaction between the uppermost asthenospheric mantle and subduction-modified subcontinental lithosphere.

In order to better understand the origin of Cenozoic intraplate volcanism on New Zealand, we have acquired new age ($^{40}Ar/^{39}Ar$) and geochemical (major element, trace element and Sr–Nd–Pb isotopic) data to assess

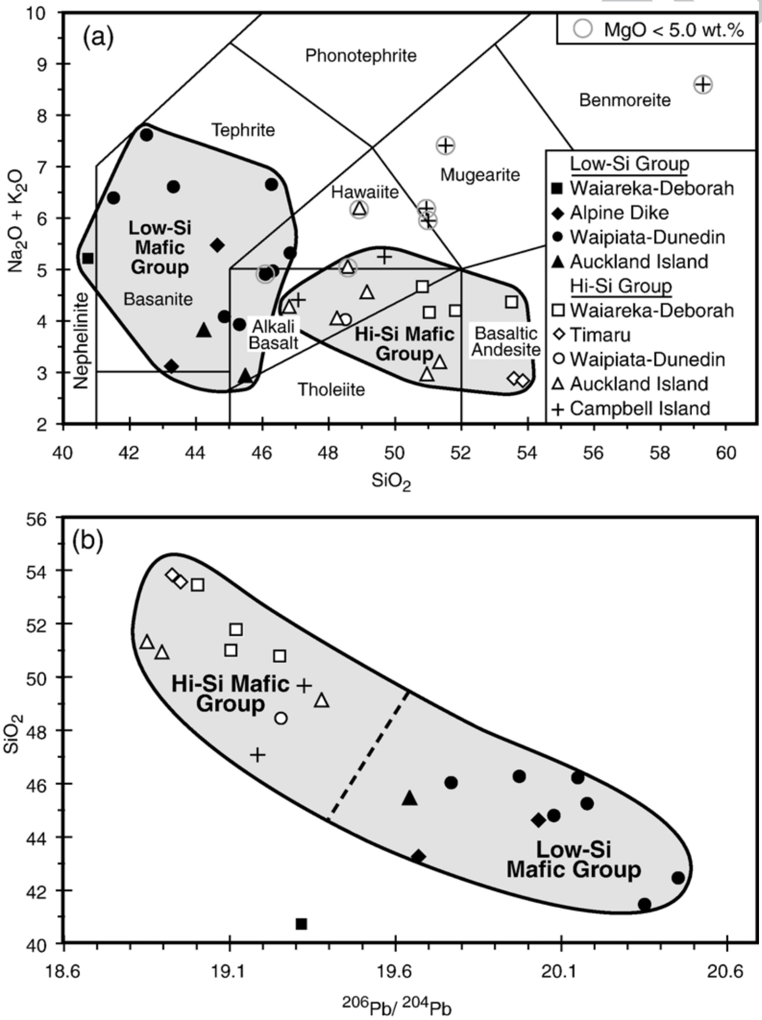

Fig. 2. a) SiO_2 versus alkali (Na_2O+K_2O) diagram after [22] for identification of rock types from the South Island of New Zealand and Campbell Plateau. Samples are divided into hi-Si (SiO_2>46 wt.%; open symbols and +) and low-Si (SiO_2<46 wt.%; filled symbols) groups. Symbols for low MgO (<5 wt.%) samples are circled. b) $^{206}Pb/^{204}Pb$ versus SiO_2 for mafic (MgO>5 wt.%) Zealandia intraplate volcanic rocks forms a negative correlation ($y=-6.5908x+176.71$) with $R^2=0.83$, excluding the Kakanui Mineral Breccia sample (for which major element data for whole rock sample OU53628 was combined with isotope data from kaersutite megacryst KK1). See Supplementary file 4 for geochemical data in this diagram.

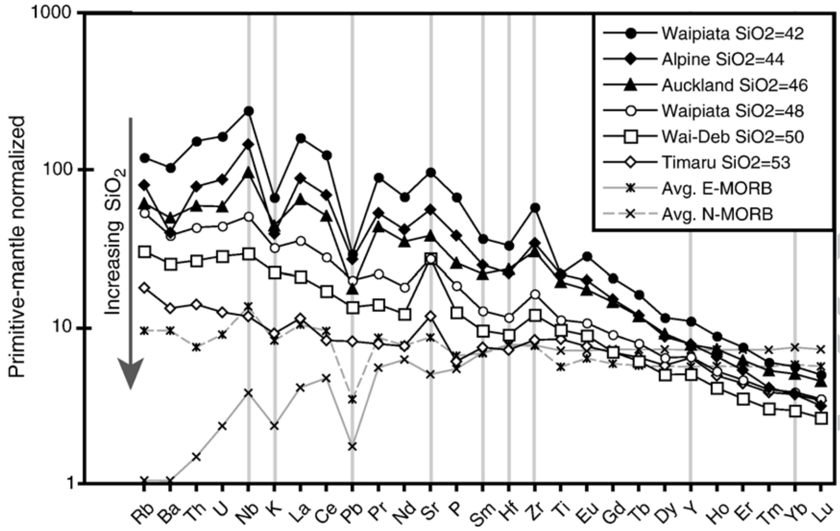

Fig. 3. Multi-element diagram for representative mafic (MgO = 8 ± 2 wt.%) samples from the South Island of New Zealand and the Campbell Plateau. Note the inverse correlation in mafic samples between SiO_2 content and concentrations in intermediate to highly incompatible elements. With increasing SiO_2, negative K and Pb anomalies disappear, positive Zr anomaly becomes less pronounced and positive Sr anomaly becomes more pronounced. Compared to mid-ocean-ridge basalt (MORB), the intermediate and highly incompatible elements are generally enriched, whereas the heavy rare earth elements (HREE) are depleted. These chemical differences are consistent with the Zealandia intraplate volcanic rocks (nephelinite/basanites through tholeiites) being formed through lower degrees of melting of more enriched source material and at greater depths in the presence of residual garnet than MORB. In contrast, MORB is formed at higher degrees of melting of more depleted source material within the spinel stability field. In addition to having HREE depletion, the Zealandia tholeiites are distinct from enriched (E) MORB in that they do not have positive Nb or negative Pb anomalies. These chemical differences to MORB can be explained by interaction of the Zealandia tholeiites with (or in some cases such as Timaru, possibly derivation from) subduction-modified continental lithospheric mantle ± continental crust. Samples in order of decreasing Nb are Waipiata OU54935, Alpine Dike Wilkin 5B, Auckland Islands R7517, Waipiata OU54926, Waiareka–Deborah OU55008 and Timaru OU33278. Average normal (N) and E MORB from [72]. Grey lines underlying the data for certain elements emphasize major incompatible element variations that correlate with variations in Si-saturation of the magmas. See Supplementary File 4 for geochemical data in this diagram.

possible age progressions in volcanism and the sources of magma for volcanic rocks from Otago on the South Island of New Zealand and from the Auckland and Campbell Islands on the Campbell Plateau.

2. Results

2.1. Age determinations

Twenty-six intraplate volcanic samples from the South Island of New Zealand and the Auckland and Campbell Islands on the Campbell Plateau (see Supplementary File 1 for sample descriptions and locations) have been dated with the $^{40}Ar/^{39}Ar$ method (age data are summarized in Table 1; analytical methods, detailed age data and age diagrams are presented in Supplementary File 2).

Six samples from the Waiareka–Deborah Formation [8] near Oamaru northeast of Dunedin (Fig. 1 inset) have ages ranging from 34–40 Ma. Five Waiareka–Deborah samples yielded the same age within error (all errors in this paper are reported at the 2 σ confidence level): 34.2 ± 0.4 Ma and 34.3 ± 0.9 Ma from glass from the margin of the same tholeiitic pillow lava at Boatmans Harbour, 34.0 ± 0.6 Ma from transitional tholeiite, 33.6 ± 1.8 Ma from tholeiite, 34.3 ± 0.5 Ma from tuff at Bridge Point and 34.1 ± 0.1 (single fusions) and 33.7 ± 0.3 Ma (step heating analyses) from kaersutite megacryst KK1 from the nephelinitic Kakanui Mineral Breccia. Dasch et al. [13] determined a slightly younger K–Ar age of 31.6 ± 1.2 Ma for the Kakanui Mineral Breccia. The new age is consistent with the Kakanui Mineral Breccia being located stratigraphically in the uppermost Eocene, just beneath the Oligocene boundary. The age of 39.5 ± 1.8 Ma is from a basaltic andesite dike fragment (with chilled glassy margins) within the tuff sequence at Bridge Point, indicating the presence of older volcanism in the Oamaru area that may not be exposed at the surface.

Seven alkali basalt to basanite samples from monogenetic volcanoes in the Waipiata volcanic field [8] yielded ages of 11–25 Ma (11.1 ± 0.6; 14.1 ± 0.6, 14.4 ± 0.2 and 14.4 ± 0.2 Ma (replicate analyses), 17.1 ± 0.5 and 17.0 ± 0.5 (replicate), 21.4 ± 0.4, 22.0 ± 0.3 and 24.8 ± 0.6 Ma), considerably expanding the published K/Ar age range of c. 13–16 Ma based on two age dates [14]. A single basanitic sample from Nevis Bluff, regarded as a southern outlier of the Alpine–Northwest Otago lamprophyric dike swarm, produced an age of 20.7±0.4, within the Waipiata age range and at the young end of the Alpine–Northwest Otago Dike Swarm range of ages, for which 15 K/Ar whole rock and kaersutite ages (from 12 rock samples) cluster between 23–32 Ma [15]. Cooper et al. [16] reported Rb/Sr and U/Pb ages for four bodies in the same swarm ranging from 20–25 Ma.

Five samples from the Dunedin Volcano (ranging from alkali basalt and basanite to phonolitie and syenite) produced ages of 11.7±0.1 Ma, 13.0±0.2 Ma, 13.2±0.5 Ma (single fusion) and 13.0±0.6 Ma (step heating analyses) from the same sample, 14.6±0.4 Ma and 16.0±0.4 Ma, extending the previously determined K/Ar age range of the Dunedin Volcano of 10–13 Ma [14] by c. 3 million years. The two oldest ages were obtained from the Otago Peninsula for which no age data has been previously published. The Dunedin Volcano appears to have formed during the later part of the Waipiata volcanic cycle near the center of the half-circular Waipiata field as it is exposed on land; Waipiata volcanism also occurred offshore. The age for a quartz tholeiite sample from Timaru of 2.6±0.4 Ma agrees well with the K/Ar age of 2.5±0.7 Ma [17].

The $^{40}Ar/^{39}Ar$ ages for three samples collected by Wright [18,19] from the upper part of the northern or Ross Volcano in the Auckland Islands are 15.2±0.3 Ma, 16.7 ±0.1 Ma and 16.7±1.2 Ma. Our ages overlap two ages of 14 and 16 Ma reported by Adams [20] from similar localities, Ewing Island and Williamson Point. The overall range of Adams' K/Ar ages for the deeply eroded and complex Ross Volcano is 12–25 Ma. Available K/Ar dates for the southern or Carnley Volcano, 19–37 Ma, are from the Carnley Harbour area within the core of the

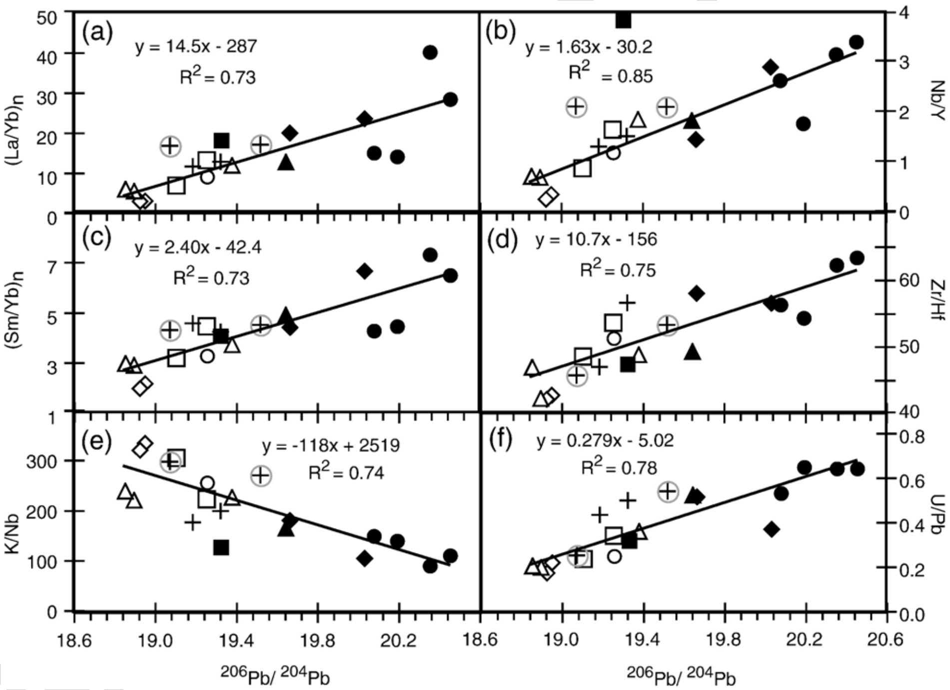

Fig. 4. $^{206}Pb/^{204}Pb$ versus a) $(La/Yb)_n$(n = normalized to primitive mantle after [73]), b) Nb/Y, c) $(Sm/Yb)_n$, d) Zr/Hf and f) U/Pb correlate (R^2 = 0.73–0.85) positively, whereas $^{206}Pb/^{204}Pb$ versus e) K/Nb correlates (R^2 = 0.74) negatively. Positive correlation of Pb isotope ratios with $(La/Yb)_n$ and Nb/Y is consistent with low-Si rocks being derived from more enriched sources through lower degrees of melting than hi-Si rocks. $(Sm/Yb)_n$ in the low-Si rocks indicates residual garnet in the sources of all the Zealandia intraplate volcanic rocks but more residual garnet in the low-Si than hi-Si rocks. High Zr/Hf in all the rocks is consistent with eclogite (with modal clinopyroxene/garnet ≥70%), whereas lower Zr/Hf in the more silicic rocks suggests greater degrees of eclogite melting up to ~50% [53]. The high U/Pb and $^{206}Pb/^{204}Pb$ ratios in the low-Si rocks indicate derivation from a source with high time-integrated $^{238}U/^{204}Pb$ (μ) or a HIMU mantle source. The high K/Nb (and low U/Pb) in the hi-Si rocks are consistent with interaction of asthenospheric melts from HIMU sources with enriched mantle (EM) melts from the overlying subduction-modified (lithospheric) mantle and/or continental crust. Best-fit linear correlations calculated for mafic volcanics with MgO > 5, excluding the Kakanui Mineral Breccia. Kakanui Mineral Breccia incompatible element data are from matrix separate from sample OU53628 and the isotope data are from kaersutite megacryst KK1 in the Kakanui Mineral Breccia. Symbols are the same as in Fig. 2. See Table 2 and Supplementary File 4 for geochemical data in this diagram.

volcano [20], and probably do not represent the younger, upper members of this shield volcano. Additional age dating is necessary to confirm the long history of volcanism on these islands.

The four Campbell Island ages (6.94±0.05 Ma, 6.6±0.6 Ma and 7.3±0.3 Ma from feldspar and matrix step heating analyses of the same sample and 7.5±0.3 Ma) fall within the age range determined by the K/Ar method, 6.5–8.5 Ma, together with a discrepant date of 11.1 Ma for a sample with low radiogenic argon and an imprecise 16 Ma age for an alkali gabbro [21]. The alkali gabbro intrudes a unit consisting of marine sediments of mid- to late Miocene age interbedded with and overlain by tuffs, volcanic breccias and lignite. There is a rapid conformable

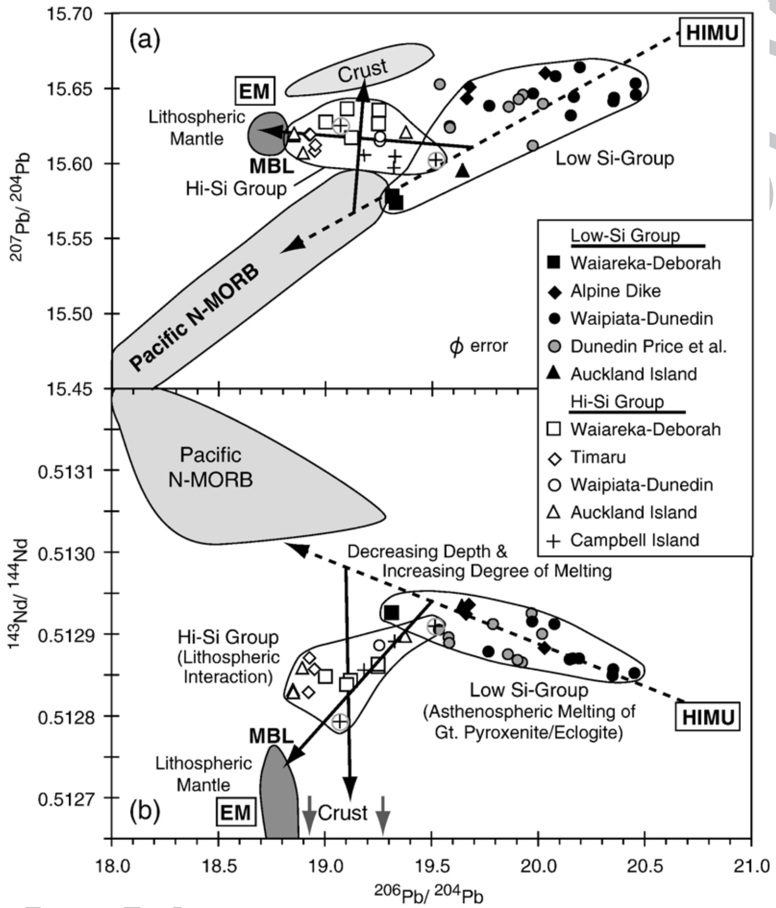

Fig. 5. On $^{206}Pb/^{204}Pb$ versus a) $^{207}Pb/^{204}Pb$ and b) $^{143}Nd/^{144}Nd$ isotope correlation diagrams, the low-Si and hi-Si groups form nearly completely distinct fields, which require mixing of at least three components. The low-Si group rocks have more radiogenic Pb and less radiogenic Nd isotope ratios than normal mid-ocean-ridge basalt (N-MORB) and form an array between Pacific N-MORB and high μ (HIMU)-type mantle, consistent with melting of garnet pyroxenite/eclogite within upwelling asthenosphere at greater depths and lower degrees of melting than N-MORB. The dashed arrow through the low-Si group array indicates increasing degrees of melting of the pyroxenite/eclogite component with decreasing depth of melting. N-MORB is formed by the highest degrees of melting, primarily in the spinel stability field, with the largest contribution from the most depleted (peridotitic) source material. Hi-Si Zealandia rocks form a crude array extending from the low-Si group field towards enriched mantle (EM) or the field of c. 108 Ma Group B and C dikes from Marie Byrd Land (MBL) Antarctica [27]. In contrast to the Group A dikes, which have HIMU-type geochemical characteristics and more radiogenic Pb and Nd isotope ratios, the Group B and C dikes (gray field labeled MBL) are characterized by 1) low $Fe_2O_3^t$ (<13 wt.%) and TiO_2 (<2.2 wt.%), 2) distinct Nb and Ta troughs and distinct peaks at fluid-mobile elements such as Rb, K and Sr on chondrite-normalized multi-element diagrams (subduction-type incompatible element signatures), and 3) the most radiogenic Sr and least radiogenic Nd and Pb isotope ratios. The Group B and C dikes are interpreted to most closely reflect the lithospheric mantle composition beneath Gondwana [27]. At least parts of the subduction-modified lithospheric mantle beneath Gondwana (Marie Byrd Land and Zealandia), however, is expected to have had more extreme EM-type compositions with less radiogenic Pb and Nd isotopic compositions than the Group B and C dikes. Solid arrows indicate interaction of hi-Si melts (alkali basalts and tholeiites) derived through melting of upwelling asthenosphere with the Zealandia continental lithosphere (mantle and crust). Data from Dunedin volcano [74] is also shown. Field for representative New Zealand sediments is from Tappenden [26]. In (b) the errors are smaller than the symbols.

transition from volcaniclastic deposits to the overlying lavas forming the island volcano. In conclusion, the subaerial part of the Campbell Island volcano is likely to have formed within 1–2 million years.

2.2. Geochemistry

New major element, trace element and Sr–Nd–Pb isotope data and analytical methods are presented in Table 2 and Supplementary Files 3 and 4. Supplementary File 4 contains all geochemical data in a single spreadsheet. Samples range from tholeiite to basaltic andesite (Timaru, Waiareka–Deborah, Auckland Islands), alkali basalt to benmoreite (Dunedin Volcano, Waipiata and Auckland and Campbell Islands) and basanite/nephelinite (Waiareka–Deborah, Alpine Dike Swarm, Auckland Islands, Waipiata and Dunedin Vol cano) based on the SiO_2 versus alkali diagram [TAS after 22] (Fig. 2). Mafic (MgO > 5 wt.%) samples are divided into hi-Si (SiO_2 > 46 wt.%) and low-Si (SiO_2 < 46 wt.%) groups. Auckland Island samples extend to more evolved hawaiitic compositions and Campbell Island samples range from alkali basalt to benmoreite. Basaltic andesites can be derived from tholeiites through removal of olivine, clinopyroxene, plagioclase and minor Fe–Ti oxides, whereas benmoreites can be derived from alkali basalts primarily through removal of these phases plus apatite.

In the mafic samples, SiO_2 is negatively correlated with FeO^t (total iron as FeO) and CaO and incompatible element abundances (e.g. Rb, Ba, Th, U, Nb, Ta, K, LREE, Sr, P and Ti, excluding Nevis Bluff sample for Sr and Ba). On multi-element diagrams (e.g. Fig. 3), the samples display enrichment in highly to moderately incompatible elements with all samples having steep heavy rare earth element (HREE) patterns and $(Sm/Yb)_n$ (n = normalized to primitive mantle), $(Dy/Yb)_n$ and $(Zr/Y)_n > 1$. With increasing SiO_2-saturation in the mafic rocks (basanite to tholeiite), the Nb and Zr peaks and the K and Pb troughs become less pronounced, whereas the Sr peak becomes more accentuated. These variations are reflected in negative correlations of SiO_2 with ratios of more to less incompatible elements (e.g. La/Yb, La/Sm, Nb/Y, Th/Yb, Zr/Y and U/Pb) and with ratios of elements with similar bulk partition coefficients during mantle melting (e.g. Zr/Hf, Nd/Pb and Ce/Pb). SiO_2, however, correlates positively with ratios of fluid-mobile to less fluid-mobile elements (e.g. (K, Rb, Ba, Sr, Pb)/Nb and Pb/(U, Nd, Ce)). Excluding evolved (benmoreite and mugearite) samples, $^{206}Pb/^{204}Pb$ isotope ratios form inverse correlations with SiO_2 (Fig. 2b) and (Ba, Rb, K, Pb, Sr)/Nb ratios and positive correlations with FeO^t, CaO, most incompatible element concentrations and La/Yb, La/Sm, Nb/Y, Th/Yb, Zr/Hf and (Ce, Nd, U, Th)/Pb ratios (e.g. Fig. 4). On isotope correlation diagrams ($^{87}Sr/^{86}Sr$ versus $^{143}Nd/^{144}Nd$ and $^{206}Pb/^{204}Pb$ versus $^{207}Pb/^{204}Pb$, $^{208}Pb/^{204}Pb$ and $^{143}Nd/^{144}Nd$), the more SiO_2-saturated (hi-Si) mafic rocks form fields distinct from those of highly SiO_2-undersaturated (low-Si) rocks, with the low-Si rocks having compositions similar to high time-integrated $^{238}U/^{204}Pb$ (μ) (or HIMU) ocean island basalts (OIBs) and the hi-Si rocks having compositions tending towards the enriched mantle (EM) components observed in OIB (e.g. Fig. 5).

3. Discussion

3.1. Temporal and spatial variations of Cenozoic intraplate volcanism

Late Cretaceous and Cenozoic tholeiitic to nephelinitic intraplate volcanism has been commonplace and widespread in New Zealand and on the Chatham Rise and the Campbell Plateau [2]. The Late Cretaceous intraplate volcanism (~100–65 Ma) forming the Mandamus Igneous Complex (e.g. [23]), the Tapuaenuku Igneous Complex [24], the Hohonu Range [16] and the Southern Volcanics of the Chatham Islands [12,25] is probably associated with the final phase of Gondwana breakup in which the micro-continent Zealandia separated from West Antarctica and may in part reflect the involvement of a mantle plume head in this breakup [2,12,24,26,27].

Cenozoic intraplate volcanism on Zealandia began with Paleocene volcanism offshore of Otago (Fig. 1 inset; 1 km of tuffs in the Endeavour 1 Drillhole [8]), in South Westland (Arnott basalt [28,29]) and possibly in Marlborough (basanitic dikes in the Tapuaenuku Igneous Complex; [24]). The Arnott basalt, outcropping west of the Alpine Fault (on the Australian Plate), was located northwest of the Auckland Islands at the time of its formation, ~500 km south of its present location [16,30,31]. Paleocene to Lower Eocene tholeiitic basalts are also present in Canterbury [32] (e.g. the View Hill Volcanics near Oxford have been dated at 51.7 ± 2.4 Ma and 47.9 ± 3.6 Ma [33]). In the Late Eocene (~34–41 Ma), nephelinitic to tholeiitic volcanic rocks were erupted in eastern Otago (Waiareka–Deborah Volcanics; Table 1) and on the Chatham Islands (Northern Volcanics; [25,34]. Our new results show that nephelinitic through tholeiitic rocks were erupted simultaneously within error of 34.1 Ma in a very restricted area in Otago near Oamaru. The Lower Oligocene Otitia

Basalt outcrops in South Westland [35]. Relatively large volumes (~100 km^3) of Early Oligocene (~34–28 Ma) tholeiitic to nephelinitic volcanic rocks are widespread in northern Canterbury and eastern Marlborough (e.g. Cookson [36] and Oxford [37] volcanic rocks). Some of the lamprophyric Alpine Dikes, ranging from nephelinite, basanite and carbonatite to phonolite and trachyte [38], intruded in the Oligocene and the Early Miocene (Table 1 and [15,16]).

Neogene and Quaternary intraplate volcanism was widespread on Zealandia. Our new data show that dozens of alkali basaltic and basanitic to mugearitic and phonolitic monogenetic centers formed in eastern Otago as part of the Waipiata volcanic field [8] during the Late Oligocene to Late Miocene (25–11 Ma; Table 1), overlapping onto the older (40–34 Ma) Waiareka–Deborah volcanic field from the south (Fig. 1b). The mid-Miocene (c. 16–10 Ma; Table 1 and [14]) Dunedin composite shield volcano, which lies geographically within the diffuse Waipiata volcanic field, has compositions ranging from alkali basalt and basanite to trachyte and phonolite, strongly overlapping the compositional range of the Waipiata volcanics. The Dunedin Volcano also temporally overlaps both the Waipiata volcanism and volcanism on the Auckland Islands as reported here (17–15 Ma; Table 1; [20,39]). The large composite Lyttelton (350 km^3) and Akaroa (1200 km^3) shield volcanoes on Banks Peninsula formed between 12–6 Ma. With decreasing age, volcanism on Banks Peninsula produced first andesites through rhyolites, then tholeiites and alkali basalts through trachytes, and finally tholeiites through nephelinites [3,40]. The Banks Peninsula volcanism overlaps in age with alkali basaltic to benmoreitic volcanism on Campbell Island as reported here (6.6–7.5 Ma; Table 1, Fig. 1) and widespread basanitic through tholeiitic (and their differentiates) volcanism on the northwestern part of the North Island, which contains the Northland (9.7 Ma to Recent) and Auckland (2.7 Ma to Recent) Volcanic Provinces [11]. Pliocene volcanism also took place on Chatham Island (basanitic to phonolite; 6–3 Ma [12,25], and near Geraldine and Timaru in Canterbury on the South Island (tholeiitic; 2.6 Ma; Table 1 and [17]). Pleistocene (≤1 Ma [41]) volcanic rocks (basanitic to phonolitic) are present on the Antipodes Island [34] at the northeastern edge of the Campbell Plateau.

In summary, it has long been recognized that Cenozoic intraplate volcanism on Zealandia, a region of relatively mature continental crust, was nearly continuous and widely dispersed [2]. The new age results from Otago presented here show that 1) compositionally diverse nephelinitic to tholeiitic volcanic rocks of the Waiareka–Deborah Formation were erupted simultaneously (within analytical errors), 2) the alkaline Waipiata field of monogenetic centers formed over ≥14 million years (c. 25–11 Ma) expanding the published age range (13–16 Ma; [14]) by 11 Ma, 3) volcanism associated with the Dunedin Volcano (16–<12 Ma) began 3 million years earlier than previously reported, and 4) the Dunedin volcano formed within the Waipiata field during the final stages of Waipiata volcanism. These results provide important constraints on the timing and ultimately origin of different types of intraplate volcanism on Zealandia.

3.2. Problems with previously proposed models for the origin of intraplate volcanism

Diffuse intraplate volcanism in Zealandia had a wide compositional range including tholeiite through rhyolite, alkali basalt through trachyte and basanite through phonolite (and syenite) with rarer nephelinite and carbonatite. No systematic age progressions (except on small scales of <100 km, e.g. Banks Peninsula) are obvious in the distribution of the volcanic rocks. Multiple episodes of volcanism occurred 1) at the same location (e.g. Chatham Island; Late Cretaceous, Eocene and Pliocene [12,25] or within restricted areas (e.g. Waiareka–Deborah, Waipiata and Dunedin Volcanism occurred within an area of ~100 km across from c. 40–10 Ma), and 2) at widely differing locations at the same time (e.g. Auckland Island, Dunedin Volcano and Waipiata volcanics; Banks Peninsula, Campbell Island and Northland Volcanic Province). Zealandia drifted ~4000 km during the Cenozoic (~6.1 cm/yr on average to the NW [42,43] and ~1800 km between 40 and 10 Ma during which time the Waiareka–Deborah, Waipiata and Dunedin magmas erupted in a restricted area of ~100 km across. Therefore the Cenozoic intraplate volcanism is incompatible with the hotspot (mantle plume) hypothesis, which predicts chains of volcanoes that become progressively older in the direction of plate motion, unless we invoke a swarm of weak pulsating plumes. No obvious plume-like structures, however, have been identified in seismic tomographic images beneath Zealandia [4]. Spatially restricted volcanism occurring over long time periods is also inconsistent with the passage of Zealandia over a linear, NE-oriented region of upper mantle upwelling [6,7]. Major continental rifting ended in the mid-Cretaceous, and later extension is spatially coincident with only a minority of Cenozoic intraplate volcanic events, so continental rifting accompanied by major lithospheric thinning also fails to explain the origin of the widespread Cenozoic intraplate volcanism of Zealandia. Furthermore, it is unlikely that Cenozoic extension caused enough lithospheric thinning to generate the larger degrees and extents of melting required, for example, to form tholeiitic melts and large composite volcanoes, such as the Dunedin (possibly 800 km^3 if possible intrusives are included [44]), Banks Peninsula (~1600 km^3), Auckland Island or Campbell Island volcanoes. It is also difficult to conceive of mechanisms for causing such relatively short (≤6 m.y.) but extensive melting events solely in the lithosphere through the melting of metasomatized lithosphere via heat conduction from the asthenosphere, since conductive heating is an inefficient process for producing large volumes of melt [45]. Finally, although 30% assimilation of continental crust by a basanitic melt could explain much of the major element chemistry of the tholeiites, the lower abundances of most incompatible elements in the tholeiites compared to basanites and common crustal rocks in New Zealand are inconsistent with such a model.

3.3. Generation and sources of Cenozoic intraplate volcanism

Systematic differences that correlate with degree of SiO_2-saturation of the parental melts exist in the major element, trace element and Sr–Nd–Pb isotopic compositions of the Cenozoic Zealandia intraplate volcanic rocks. As noted above, the mafic (MgO>5 wt.%) intraplate volcanic rocks can be divided into two groups based on their SiO_2 contents: 1) low-Si group with SiO_2<46% consisting primarily of basanites but also alkali basalt, nephelinite and tephrites, and 2) hi-Si group with SiO_2>46 wt.% consisting of alkali basalt through hawaiite and tholeiite through basaltic andesite. In contrast to the hi-Si group, the low-Si group has 1) high FeO^t and CaO, 2) high incompatible element abundances (e.g. Rb, Ba, Th, U, Nb, Ta, K, LREE, Sr, P and Ti), 3) high ratios of more to less incompatible elements (e.g. La/Yb, La/Sm, Nb/Y, Th/Yb, Zr/Y, U/Pb, including HREE ratios such as Sm/Yb, Dy/Yb and Tb/Yb), 4) high ratios of elements with similar bulk partition coefficients during melting of mantle rocks (e.g. Zr/Hf, Nd/Pb and Ce/Pb), and 5) low ratios of fluid-mobile to less fluid-mobile elements (e.g. (K, Rb, Ba, Sr, Pb)/Nb and Pb/(U, Nd, Ce)) (e.g. Figs. 3 and 4). Finally the low-Si group extends to less radiogenic Sr and more radiogenic Nd and Pb isotopic compositions than the hi-Si group and falls on a mixing array between the ocean island basalt HIMU and the Pacific MORB components (Fig. 5). The hi-Si group extends from the low-Si array towards enriched mantle (EM) and specifically towards mid-Cretaceous volcanic rocks from Marie Byrd Land Antarctica [27] (Fig. 5), which provide insights as to the composition of the lithospheric mantle beneath western Gondwana of which Zealandia was a part until c. 84 Ma. The wide range of geochemical differences between the low-Si and hi-Si volcanic rocks indicates that the melts were formed under fundamentally different conditions, to be discussed below.

Comparison of the major element compositions of the Cenozoic intraplate Zealandia basalts with melts formed experimentally at high pressures can provide important insights about the source(s) of the basalts. Such high-pressure (1.5–7.0 GPa) experiments have been carried out on a wide array of possible mantle compositions including volatile-free peridotite (e.g. [46]), peridotite+CO_2 [47], peridotite–basalt mixtures [48], garnet pyroxenite [49,50] and eclogite+CO_2 [51]. Comparison of the melts formed at a given MgO (e.g. 10–12 wt.%) in the high-pressure melting experiments in the sequence eclogite+CO_2, garnet pyroxenite, peridotite–basalt mixtures to volatile-free peridotite shows that there is a general increase in SiO_2 (from <40 to ~52 wt.%) and Al_2O_3 and a general decrease in CaO, FeO^t and TiO_2 in the melts (see compilation in [51], their Fig. 12). The mafic Zealandia intraplate rocks fall within the range defined by these experimental results. The major element contents and inverse correlations of SiO_2 with CaO, FeO^t and TiO_2 in the Zealandia rocks suggest that the low-Si melts formed from sources containing carbonated eclogite, whereas the hi-Si melts formed through greater degrees of melting of a volatile-free peridotite±eclogite/pyroxenite source.

The negative correlations of SiO_2 with incompatible element abundances and ratios of more to less incompatible elements are consistent with derivation of more Si-rich melts through increasing degrees of partial melting of more depleted sources with decreasing depth (e.g. [46,52]). The steep HREE patterns (e.g. $(Sm/Yb)_n$ and $(Dy/Yb)_n>1$) indicate the presence of residual garnet in the source of all parental intraplate melts. The high (super chondritic) Zr/Hf ratios of all melts could result from melting of eclogite (or garnet pyroxenite) with high clinopyroxene/garnet modes of ≥70 in their source(s) [53,54]. Based on the calculations of Pertermann et al. ([53]; see their Fig. 10), the range in Zr/Hf and Sm/Yb in the New Zealand rocks can reflect eclogite melt fractions of several percent for the basanites to about 50% for the tholeiites.

The trace element abundances of the New Zealand intraplate volcanic rocks (Fig. 3) are characteristic of ocean island basalts (OIBs). They are enriched compared to MORB and have different trace element characteristics (patterns on multi-element diagrams) than subduction zone basalts that display highly spiked patterns with relative enrichments in fluid-mobile elements (e.g. Rb, Ba, U, K, Pb and Sr) but relative depletion in fluid-immobile elements (e.g. Nb), resulting in high (Rb, Ba, U, K, Pb, Sr)/Nb and low (Ce, Nd, U)/Pb ratios. The trace element patterns of the low-Si rocks, characterized by peaks at Nb and troughs at K and Pb, and their high Nd/Pb, Ce/Pb and U/Pb and radiogenic Pb isotope ratios indicate derivation from a HIMU-type mantle source. HIMU trace element and isotopic compositions are characteristic of volcanic rocks from ocean islands such as Mangaia, Tubuaii and St. Helena, interpreted to result from the presence of ancient (c. 1–2 Ga) recycled ocean crust in their sources (e.g. [55]). Although the low-Si Zealandia samples have elevated Pb isotope ratios (e.g. $^{206}Pb/^{204}Pb = 19.3–20.5$), they are mostly not as high as those of the endmember HIMU-type ocean island localities mentioned above (with $^{206}Pb/^{204}Pb$ of 20.5–22.0). On Sr–Nd–Pb isotope correlation diagrams, the low-Si rocks form an array between Pacific N-MORB and the HIMU endmember ocean islands. The less radiogenic Pb isotopic composition of the low-Si rocks from Zealandia compared to endmember HIMU could either reflect lower time-integrated parent/daughter (U/Pb and Th/Pb) ratios in the Zealandia sources for similar time periods, or similar (or higher) parent/daughter ratios in their sources for shorter periods of time. The similarity in U/Pb and Th/Pb ratios in the Zealandia intraplate volcanism and ocean island basalts, and location of much of the low-Si field to the right of the extended MORB basalt array on both Pb isotope diagrams, in particular on the thorogenic Pb isotope diagram (not shown), favor the latter possibility. Therefore, relatively young recycled ocean crust (eclogite) or young pyroxenite layers of igneous origin are likely to be present in the sources of the Zealandia intraplate volcanic rocks.

In accordance with the lack of evidence for mantle plumes beneath Zealandia and the problem of generating large volumes of entirely lithospheric melting (e.g. on the scale necessary to form a shield volcano), we favor generation of most of the Zealandia intraplate melts from asthenospheric sources. Finn et al. [4] have shown that the south Pacific asthenosphere has unusually low seismic velocities and propose that volcanism results from decompression melting of upwelling warm (~1300–1400 °C) Pacific mantle and partial melting of the base of the metasomatized subcontinental lithosphere. Compared to MORB, the intermediate and highly incompatible elements of all the Cenozoic intraplate volcanic rocks are generally enriched, whereas the heavy rare earth elements (HREE) are depleted. These chemical differences are consistent with the Zealandia intraplate volcanic rocks being formed through lower degrees of melting from more enriched (eclogitic/pyroxenitic) source material and at greater depths in the presence of residual garnet, than MORB. In contrast, MORB is formed at higher degrees of melting of more depleted source material predominantly within the spinel stability field.

The isotope data for the low-Si group forms an array between the MORB and HIMU sources (Fig. 5), suggesting that the source of these rocks is the upper mantle, which contains some HIMU component. As discussed above, the HIMU component is likely to reflect younger eclogite/garnet pyroxenite than commonly sampled by mantle plumes. The MORB source upper mantle may contain 5% pyroxenitic/eclogitic layers in a peridotitic matrix [56]. Since altered oceanic crust contains a significant amount of calcium carbonate in vugs and veins, at least some of the eclogite in the mantle is likely to be carbonated. Carbonated eclogite could also be created by other processes within the mantle, for example crystallization of carbonate-rich melts [57]. Dasgupta et al. [51] present a model for melting and metasomatism in upwelling carbonated oceanic mantle (containing carbonated peridotite and eclogite or pyroxenite) that can explain the geochemistry of the Cenozoic Zealandia mafic volcanic rocks. Carbonated low-silicate melts of eclogite/pyroxenite and peridotite, such as melilitites, nephelinites and basanites, will form in the deeper parts of the upwelling mantle column, whereas CO_2 absent melting of eclogite/pyroxenite and peridotite will commence at higher levels in the upwelling column, producing high-silica alkali basalts and tholeiites. In contrast to the mid-ocean-ridge setting, the continental lithosphere (thicker than oceanic lithosphere) precludes extensive melting in the spinel stability field (Fig. 6) and therefore the Zealandia intraplate melts may be similar to melts from the deeper portions of the MORB melting column. Melts in equilibrium with garnet, however, are rarely preserved at mid-ocean ridges, due to the swamping effect of large volumes of melts formed at shallower depths. In conclusion, because of its greater thickness, continental lithosphere may allow sampling of the lower portions of the MORB (upwelling upper mantle) melting column. Variations in the thickness of the continental lithosphere may ultimately control the type of melting and the style and composition of the ensuing volcanism. A fundamental question, however, remains: What triggers the upwelling of upper mantle beneath Zealandia and what controls its location and duration? This will be discussed in the next section.

First we need to discuss some additional aspects of the geochemistry of the hi-Si rocks that are not consistent with asthenospheric melting. Although the trace element abundances of the hi-Si rocks are similar to those of ocean island alkali basalts and tholeiites, the almost smooth incompatible trace element patterns (except for positive Sr anomaly) without a pronounced Nb peak and Pb trough of the samples with the highest silica on the multi-element diagram (Fig. 3), as reflected for example

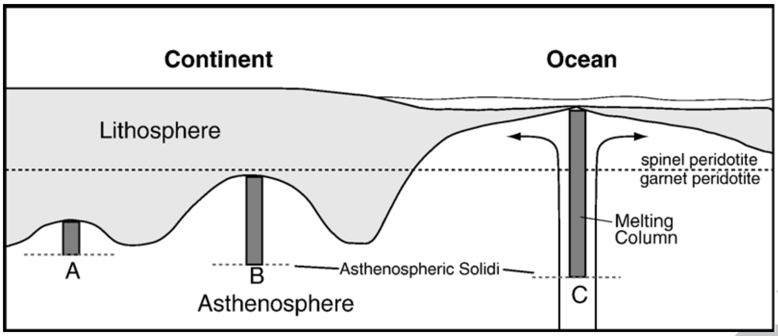

Fig. 6. Melting model illustrating the difference lengths of the Zealandia (A and B) and MORB (C) melting columns, i.e. depth intervals over which melting occurs, which extend from the asthenospheric solidus (based on temperature of the upwelling mantle) to the base of the lithosphere. The deeper the mantle that is sampled by the upwelling, the hotter the upwelling mantle will be and thus the solidus will be crossed at greater depth, resulting in greater overall degrees of melting. Beneath continental lithosphere, basanites and possibly nephelinites (melting column A) can be formed by low degrees of melting of garnet pyroxenite/eclogite within the shortest melting columns, resulting from the least amount of upwelling asthenosphere. Greater asthenospheric upwelling and melting of the garnet pyroxenite/eclogite component occur where greater amounts of continental lithosphere have been removed (melting column B), producing alkali basalts and tholeiites with garnet in the residuum. Melting column B is longer, since the hotter mantle from greater depth will cross the solidus earlier and melting will continue to shallower depths. The longest melting columns exist beneath mid-ocean ridges (C), where due to the thin oceanic lithosphere extensive melting takes place in the spinel peridotite stability field, forming MORB tholeiites.

by the high (Ba, K, Pb, Sr)/Nb and the low (Ce, Nd, U, Th)/Pb ratios, are not characteristic of either MORB or most OIB. The trace element characteristics of the hi-Si rocks, however, can be explained through higher degrees (than required to generate the low-Si rocks) of melting of a garnet peridotite + eclogite (or garnet pyroxenite) source and subsequent mixing with melts having incompatible element compositions similar to subduction zone or crustal melts.

Several considerations favor a continental lithospheric source for the EM-type component. First, the more saturated melts are likely to have formed at shallower depths, as also proposed by Cook et al. [11] for the alkali basaltic and tholeiitic South Auckland volcanic rocks. Second, Zealandia was located at the edge of the Gondwana margin above a subducting slab for most of the Mesozoic. Hydrous subduction zone melts have the appropriate trace element and isotopic composition for the EM endmember and are likely to have overprinted (metasomatized) the Zealandia lithospheric mantle with this signature while it was still part of Gondwana [4,11]. The trend of the hi-Si group towards the field for mid-Cretaceous Marie Byrd Land volcanic rocks, believed to have compositions most closely resembling those of the lithospheric mantle beneath Gondwana [27], further supports lithospheric interaction/derivation of the hi-Si melts. Melting of a hydrous Zealandia lithospheric mantle could have possibly even generated some of the olivine tholeiitic melts [58]. It has been shown that historic tholeiites in the Canary Islands – only present on the easternmost and thus presumably oldest island of Lanzarote – can be derived through mixing between basanites (~50–60%) derived from sublithospheric sources and a high silica melt component (~40–50%), produced either by incongruent dissolution of orthopyroxene during direct reaction of basanitic melt with the lithosphere or by diffusive infiltration of alkalis (DIA) from the basanites into the surrounding lithosphere [59]. Similar processes can also explain the generation of some of the Zealandia tholeiitic melts, such as those erupted at Timaru and Geraldine, that produced similar amounts of tholeiite as the six year Timanfaya eruption on Lanzarote (2–3 km^3 [60]). Third, the crust along the Gondwana margin, formed primarily through subduction-zone volcanism and erosion of these volcanic rocks, also has an appropriate composition to serve as the EM end-member. More SiO_2-saturated melts, due to their higher viscosity and larger volumes, are more likely to stagnate and differentiate in the crust (and lithospheric mantle), leading to assimilation during fractional crystallization (AFC). Because they have lower contents of incompatible elements, the more saturated melts are also more sensitive to contamination in the lithosphere. In conclusion, the composition of the hi-Si group can be explained by interaction between asthenospheric melts (with compositions intermediate between MORB and HIMU) and a (EM-type) lithosphere, with variable amounts of lithospheric melting being likely.

3.4. Lithospheric removal model to explain Cenozoic intraplate volcanism on Zealandia

A fundamental question concerning the origin of Cenozoic intraplate volcanism on Zealandia is what causes the melting. As noted above, the petrology and geochemistry of the volcanic rocks are consistent with generation of the parental melts through decompression melting of upwelling asthenosphere combined with lithospheric melting to produce the hi-Si melts. The problem is how to affix an asthenospheric upwelling to a restricted region of the lithosphere, for example Banks Peninsula (~50 km across) where volcanism persisted for ~6 m.y. or the Waiareka–Deborah, Waipiata and Dunedin volcanic region (~100 km across) where volcanism lasted ~30 m.y. The plate would have moved ≥300 km and ≥1800 km respectively in the aforementioned time intervals. If the upwelling regions were ≥300 km and ≥1800 km respectively, then why is the volcanism not regionally more extensive? On the other hand, if the upwelling regions had diameters similar to the volcanic areas and were fixed in the sublithospheric mantle (similar to a plume), why wasn't the volcanism more extensive over time and why don't we see any age progressions in the volcanism in the direction and at the same rate as plate motion?

Removal/detachment of parts of the lithospheric keel beneath Zealandia can explain how asthenospheric upwelling can be localized to specific regions of the plate over long periods of time (Fig. 7). In the Dunedin region, flexural modeling suggests a thermally weakened plate, which is interpreted to have resulted from a buoyant load, probably hot asthenosphere, emplaced beneath the crust during the mid Miocene [61]. These results are consistent with detached lithospheric mantle allowing hot asthenosphere to well up to shallow depths beneath the Dunedin Volcano when it formed in the mid Miocene.

Regions of Zealandia's lithospheric keel are inferred to contain large amounts of garnet pyroxenite/eclogite (e.g. frozen intruded melts and cumulates), as a result of magmatic underplating associated with 1) subduction or plume volcanism beneath the former Gondwana margin in the Mesozoic [27,62] and 2) rifting during separation of Zealandia from West Antarctica and Australia (e.g. [2]). It has been found for example that lower crustal and mantle compositions resulting from arc magmatism can create a 1–5% density contrast within the normal upper mantle [63]. Such density contrasts are sufficient to cause gravitational or Rayleigh–Taylor instabilities, resulting in the removal/detachment of the lower lithosphere in a ductile manner (e.g. [45,64,65]). This mechanism only requires a dense region in the lithosphere that is gravitationally unstable and that possesses a rheology conducive to flow [66,67]. Modeling results suggest that the latter requirement is easily met at normal lithospheric-mantle rheologies and temperatures [65]. Instabilities initiate in response to boundary perturbations, and growth rates increase with strain rate. Since younger, thinned continental lithosphere

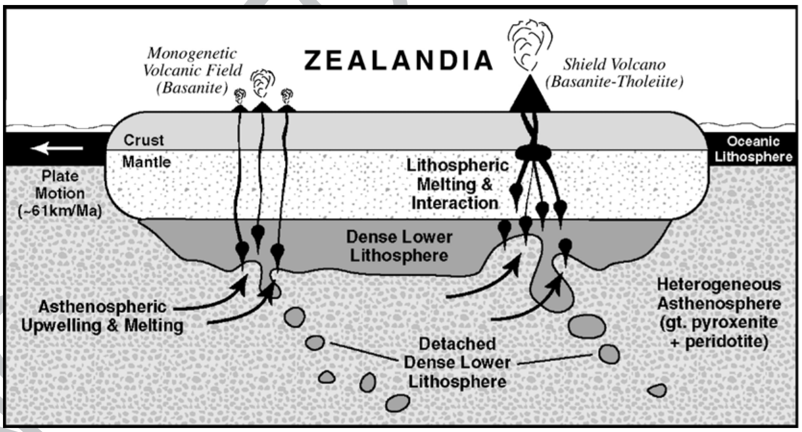

Fig. 7. Lithospheric removal/detachment model for explaining Cenozoic intraplate volcanism on Zealandia. Removal of the cooler, dense lower lithosphere (enriched in eclogite or garnet pyroxenite, e.g. intruded, frozen melts, and/or garnet-rich cumulates) is the result of Rayleigh–Taylor or gravitational instabilities developed along the lithospheric–asthenospheric mantle boundary. Lithospheric stripping is asymmetrical due to the movement of Zealandia lithosphere relative to the asthenospheric mantle. Zealandia is moving west relative to sinking lithospheric instabilities. Clusters of descending dense lower lithosphere may occur beneath areas with long histories of volcanic activity. When relatively small amounts of the lithosphere are removed, garnet pyroxenite/eclogite in hotter upwelling asthenosphere melts to small degrees to form basanitic (and possibly nephelinitic) melts, which form monogenetic volcanic fields (predominantly cinder cones and lava flows). When larger areas of lithospheric mantle are removed, possibly through progressive growth of an instability, larger volumes of asthenosphere upwell to shallower depths, resulting in greater degrees of melting to form alkali basaltic and possibly tholeiitic melts. These melts interact extensively with the lithosphere (possibly both crust and mantle) giving them more enriched trace element and isotopic signatures. Some tholeiites may result directly from lithospheric mantle melting. Areas of greater lithospheric detachment allow greater degrees of melting and the formation shield volcanoes.

is most likely to be gravitationally unstable and sink with respect to the adjacent asthenosphere [65,68], the young Zealandia lithosphere, thinned and heated in the Late Cretaceous as a result of continental breakup, is likely to have been particularly susceptible to the development of gravitational instabilities. Movement of the low viscosity Zealandia lithosphere through an anomalously hot upper mantle [4] at high rates of plate motion during the Cenozoic would have further enhanced the formation of gravitational instabilities.

The distribution and timing of volcanism can provide important clues as to the shape and amount of lithosphere detached by such instabilities, which should detach from regions of weakness or the top of the region of negative buoyancy. Since the height of the region that detaches should be similar to the width of the instability as it sinks, the surface distribution of volcanism should reflect the width of the instability and height of the removed lithosphere (e.g. [69]). In addition, rapid instability growth and detachment yield more focused detachment zones (scars) with greater relief than does slower instability development, which favors broad detachment zones with limited relief along the lithosphere–asthenosphere boundary [45]. For Zealandia, this implies multiple small detachments to explain scattered low-volumes of low-silica volcanism such as the Waipiata monogenetic volcanic field. We suggest that areas of more persistent and larger volume volcanism (e.g. Dunedin and Banks Peninsula shield volcanoes) resulted from rapid development of thicker instabilities. Development of the large Dunedin instability within the broader Waipiata field may reflect higher intensities of lithospheric mantle modification (i.e. addition of dense material) affecting greater lithospheric thickness beneath the Dunedin volcano than in the surrounding base of the lithosphere beneath the Waipiata field. Judging from the volumes and distribution of Cenozoic magmatism, we infer that lithospheric modification was highly heterogenous beneath Zealandia.

As a portion of the lithosphere is stripped off and sinks, asthenosphere can well up into the resulting cavity, melting by decompression. New volumes of asthenosphere must have continually upwelled into cavities in the base of the Zealandia lithosphere as the plate moved over the asthenosphere at a rate of ~60 km/m.y. Therefore decompression melting and volcanism could have continued for millions to tens of millions of years until the lithosphere thermally healed itself by conductive cooling. Sequential detachment of adjacent lithospheric material from an area, such as Banks Peninsula where Akaroa volcano began forming as activity at Lyttleton volcano was ending, slows thermal healing, promotes prolonged volcanism in the area, and contributes to increasing degrees of melting with time. After a final detachment event, the lithosphere will slowly anneal, resulting in decreasing degrees of melting and melt production and increasing degree of Si-undersaturation with time, as is observed on Banks Peninsula [3].

Removal of larger thicknesses of lithospheric mantle allows asthenosphere to upwell to shallower depths and over longer vertical distances (reflected in a longer melting column), to produce larger degrees of partial mantle melting and thus increased volumes of more SiO_2-saturated melts. Increased heat transfer from the upwelling asthenosphere to newly exposed parts of the Zealandia lithosphere will enhance lithospheric melting, resulting in greater lithospheric contributions to the erupted magmas, in particular to the largest degree melts (tholeiites) formed at the top of the melting column. Release of hydrous fluids through devolatilization and alkali transfer from the detached metasomatized lithosphere could increase melting and the production of high silica melts within the overlying upwelling asthenosphere and within the non-detached lithosphere. Detached hydrous lithosphere could also melt as it sinks, providing yet another source for melts [45]. Devolatilization of detached carbonated lithosphere could also serve as a source of CO_2 causing the carbonation of peridotite and eclogite/pyroxenite in the upwelling asthenosphere above the detaching lithosphere, providing an additional mechanism for the formation of Si-undersaturated melts in regions that have undergone a detachment event.

In contrast, removal of small thicknesses of lithosphere promotes only limited upwelling and thus short-lived production of small volumes of asthenospheric melt from a short melting column. Heat transfer from the upwelling asthenosphere to the base of the overlying lithosphere and volatile and alkali (+other element) transfer from the detaching lithosphere to the overlying asthenosphere and lithosphere will be at a minimum and thus melts formed from decompression melting of asthenosphere will have minimal interaction with lithospheric components. For volcanic fields such as the Waipiata, which erupted small volumes of generally primitive HIMU-type magmas over large areas and long periods of time, we envisage repeated thin detachments. It is probable that not all of these thin detachments produced surface volcanism, with many small-volume melts instead becoming trapped in the crust [70].

In conclusion, the lateral extent of the cavity or cavities formed by instability growth and detachment will govern the area of a given volcanic province, whereas the vertical extent and volume of the cavity will control the extents and amounts of melting, thereby influencing the volcanic style: monogenetic volcanic fields versus composite shield volcanoes. Interruptions to cavity annealing (repeated detachments) may result in repeated cycles of volcanism at individual shields and complicated temporal geochemical trends.

Acknowledgements

We thank D. Rau, J. Sticklus, S. Hauff for analytical assistance, M. Portnyagin for comments on this manuscript, and the "Great Plume Debate" Chapman Conference in Fort William (2005) for talks and discussions that helped further develop ideas in this paper. We are especially grateful to Lindy Elkins-Tanton and Kurt Panter for constructive reviews which helped to improve this manuscript significantly. We acknowledge Otago University for providing KH with the William Evans Fellowship, during which this project was initiated, and the German Research Foundation (DFG Project HO1833/12-1, New Zealand Volcanism) for funding part of this project. JW acknowledges support from the New Zealand FRST via contract C05X0006.

Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.epsl.2006.06.001.

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