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Diverse mantle and crustal components in lavas of the NW Cerros del Rio volcanic field, Rio Grande Rift, New Mexico

K.E. Duncker[1,*], **J.A. Wolff**[1], **R.S. Harmon**[2,**], **P.T. Leat**[3,***], **A.P. Dickin**[4], **and R.N. Thompson**[3]

[1] Department of Geology, University of Texas at Arlington, UTA Box 19049, Arlington, TX 76019, USA
[2] Department of Geological Sciences, Southern Methodist University, Dallas, TX 75238, USA
[3] Department of Geological Sciences, University of Durham, South Road, Durham DH1 3LE, UK
[4] Department of Geology, McMaster University, 1280 Main Street West, Hamilton, Ontario L8S 4MI, Canada

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Abstract. Products of Pliocene (2–4 Ma) mafic to intermediate volcanism in the northwestern Cerros del Rio, a dominantly mafic volcanic field in the Española Basin of the Rio Grande Rift (RGR), range from 49% to 63% SiO_2 and exhibit diversity in silica saturation, trace-element patterns, and isotopic compositions. Tholeiites, which are largely confined to west of the Rio Grande, have trace-element abundances that resemble those of oceanic basalts, but with mild depletions in Nb and Ta, and high $^{87}Sr/^{86}Sr$, low $^{143}Nd/^{144}Nd$, and high $\delta^{18}O$ compared to typical OIB. They are regarded as asthenospherically-derived magmas contaminated with continental crust. Alkali basalts and hawaiites erupted from vents east of the Rio Grande are geochemically distinct, having generally higher overall incompatible-element abundances, but with pronounced depletions in K, Rb, Nb and Ta with respect to Th and LREE. Spatially-associated benmoreites, mugearites and latites (collectively termed "evolved" lavas) have similar trace-element characteristics to the mafic mildly-alkaline compositions, but are typically not as depleted in K. Hawaiites and evolved lavas exhibit a good negative correlation of $^{143}Nd/^{144}Nd$ with SiO_2, due to interaction with lower continental crust. The most silicic "evolved" lavas carry the highest proportions of crustal material, and consequently have higher K/Th than the related hawaiites. Several (mostly mafic) lavas contain abundant crustally-derived resorbed quartz xenocrysts in O-isotope disequilibrium with the host magma. The $\delta^{18}O$ values of xenocrystic quartz range over 4‰, indicating a variety of quartz-bearing crustal contaminants beneath the Española Basin. The hawaiites, with their unusual combination of trace-element enrichments and depletions, cannot be generated by any process of fractionation or crustal contamination superposed on a common mantle source type (oceanic or arc-source). It is a regional mantle source type, inasmuch as it was also present beneath NW Colorado during the mid-late Cenozoic. We argue that the hawaiite source must have originally existed as arc-source mantle enriched in LILE, generated during Mesozoic to early Cenozoic subduction at the western margin of North America. This arc-source mantle lost K, Rb and Ba, but not Th or LREE, prior to magmagenesis. Selective element loss may have occurred during lithospheric thinning and uprise of hydrated phlogopite-bearing peridotite – possibly as a thermal boundary layer between lithosphere and asthenosphere – to shallow mantle depths, with consequent conversion of phlogopite to amphibole (an inferior host for K, Rb and Ba). We suggest that this occurred during the early extensional phase of the northern RGR. Further extension was accompanied by partial melting and release of magma from this source and the underlying asthenosphere, which by the Pliocene was of oceanic type. The hawaiite source mantle is the product of a long history of subduction succeeded by lithospheric extension of the formerly overriding plate. Similar chemical signatures may have developed in the mantle beneath other regions with comparable histories.

Introduction

Many continental mafic magmas exhibit close geochemical and isotopic affinities with ocean-island basalts (OIB), modified to varying degrees by interaction with the continental crust through which they pass on their way to the surface. This has led to the hypothesis that the asthenospheric OIB source is continuous beneath oceanic and continental crust (Thompson et al. 1984; Fitton and Dunlop 1985). Compared to OIB, subduction-related magmas (SRM) are characterised by

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[*] Now KE Romanak at: Department of Geological Sciences, University of Texas at Austin, Austin, TA 78713-7909, USA
[**] Now at: NERC Isotope Geology Centre, Keyworth, Nottingham NG12 5GG, UK
[***] Now at: British Antarctic Survey, High Cross, Madingley Road, Cambridge CB3 0ET, UK

Offprint requests to: J.A. Wolff

marked enrichments in Ba, Rb, Th, K and LREE relative to Nb and Ta (Pearce 1983), with magmas of both oceanic and continental arcs having generally similar trace-element and isotopic signatures. These enrichments in SRM are considered to result from chemical modification of convecting asthenosphere in the mantle wedge above the Wadati-Benioff Zone by fluids and/or magma derived from the subducting oceanic slab with entrained pelagic sediments (Arculus and Powell 1986). Beneath the continents, partial melting of lithospheric mantle may supply a further compositionally-distinct component to continental basalts. This is particularly the case in areas of crustal extension and rifting, because upwelling of hot asthenosphere may result in thermal remobilisation and partial melting of overlying lithosphere (Perry et al. 1987). In the western United States, eastward shallow subduction of the Farallon plate beneath the continent during the Mesozoic and early Tertiary was succeeded in the mid-Tertiary by crustal extension. Thus, according to the perhaps rather simplified picture summarised above, Neogene mafic lavas erupted in this region may consist of up to five components, derived from asthenospheric OIB-source-like mantle, mantle-wedge material remaining from early Tertiary subduction, reactivated lithospheric mantle, and upper and lower continental crust.

This paper reports and discusses the major, trace element and isotopic characteristics of lavas from a part of one the larger volcanic fields within the Rio Grande Rift. Very few of the lavas can be directly related to each other by mechanisms such as closed-system differentiation, assimilation-fractional crystallization (AFC), or magma mixing (although these processes have clearly been involved in the genesis of the whole suite), but must be regarded as the products of separate magma batches. For this reason our aim is not to attempt the construction of detailed petrogenetic schemes, but rather to identify the different mantle and crustal components which have contributed to magmagenesis beneath this portion of the Rio Grande Rift.

The Rio Grande Rift

The Rio Grande Rift (RGR) is the only significant manifestation of Tertiary extension in North America to the east of the Colorado Plateau. In the central part of the RGR, comprising the Albuquerque-Belen, Española, and southern San Luis Basins in northern New Mexico, rifting began around 30 Ma, since which time crustal extension has been markedly episodic. Early rifting (at a time when ridge-trench collision of the California coast had barely begun, Atwater 1970) was characterised by low-angle planar and/or listric faulting, with crustal extension locally exceeding 100% (Morgan et al. 1986). In late Oligocene times, the Española Basin, coincident with but broader than the present basin, began to form with a western fault boundary in the Nacimiento Mountains (Aldrich 1986; Morgan et al. 1986). Rift inception was preceeded and accompanied by regional volcanism covering a wide compositional spectrum (Aldrich et al. 1986). The deduced paleogeotherm and structural style suggest relatively low upper mantle strength during this episode (Morgan et al. 1986).

Active crustal spreading ceased for most of the period between 20 and 10 Ma. Spreading was re-initiated along the pre-existing axis between 13 and 9 Ma (Aldrich et al. 1986; Gardner et al. 1986; Morgan et al. 1986; Self et al. 1986), but with a structural style now characterised by half-grabens bounded on one side by widely-spaced high-angle normal faults. Basin development thus involved more vertical than horizontal displacement, which latter may have been as little as 10% (Morgan et al. 1986). Morgan et al. deduce significant lithospheric cooling since the first extensional episode, such that the uppermost upper mantle acquired significant mechanical strength, i.e. took on the physical characteristics of lithosphere.

In the Española Basin and adjacent western rift flank (now covered by the Jemez Mountains Volcanic Field, JMVF), mafic volcanism began during the quiet period at around 16.5 Ma (Gardner et al. 1986). This was followed by extensive mafic volcanism which produced the Paliza Canyon Formation of the JMVF between 13 and 7 Ma (Gardner et al. 1986). Extension rates at this time were at a maximum for the second phase of deformation (Morgan et al. 1986). A further tectonic lull between 7 and 4 Ma ended with the development of the Pajarito fault zone (Fig. 1), marking an eastward shift of the Basin's western boundary (Aldrich 1986; Gardner et al. 1986; Self et al. 1986). The lowest exposed Cerros del Rio (CdR) lavas were erupted at this time. The late Pliocene and early Quaternary saw the formation of the CdR field in the Basin floor, and large-scale rhyolitic volcanism and caldera formation in the rift-flanking JMVF.

Morgan et al. (1986) note that the second faulting episode was accompanied by regional uplift along the whole length of the RGR, and that the whole evolutionary sequence – two-phase extension accompanied and followed by regional uplift – applies generally to much of the Basin and Range Province and the whole western US. Uplift on this broad scale was ascribed by Angevine and Flanagan (1987) to removal of subcontinental lithospheric mantle.

The Jemez Lineament is a deep crustal feature marked by recognisable faults (Aldrich and Laughlin 1984) and a NE-trending alignment of Neogene volcanic centres stretching some 600 km from eastern Arizona to southeastern Colorado (Laughlin et al. 1976); it forms the southeastern boundary of the Colorado Plateau. Its intersection with the western margin of the Española Basin at the pre-4 Ma Cochiti bounding fault zone is now marked by the JMVF. Northeast of the JMVF, the lineament crosses the RGR as the Embudo fault zone, which effectively forms the northern margin of the Española Basin (Aldrich 1986).

Field relationships

The CdR volcanic field lies in the southwestern part of the Española Basin. Eastward, the lavas and associated tuffs overlie and inter-

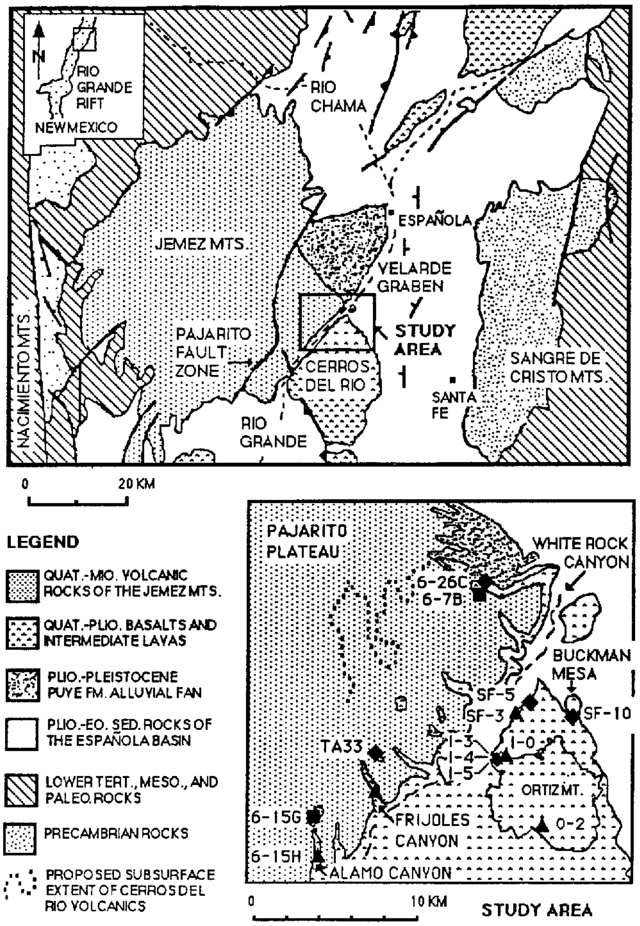

Fig. 1. Geologic map showing the position of the Cerros del Rio volcanic field within the Española Basin, which is bounded on the west by the Jemez Mountains and on the east by the Sangre de Cristo Mountains. Smaller map is of the study area, modified after Smith et al. (1970) and Dransfield and Gardner (1984). Locations for the representative samples given in Table 1 are shown. All other samples were taken from within the map area

stratify with Santa Fe Group and younger basin-filling sediments. North and northwestwards the lavas are interbedded with volcanogenic sediments in the Puye alluvial fan, shed off the developing Tschicoma volcano in the Jemez Mountains (Waresback and Turbeville 1990). Dransfield and Gardner (1985) have identified a northwestward extension of the CdR on the Pajarito plateau (Fig. 1), currently buried beneath the two rhyolitic Bandelier Tuffs, erupted from the JMVF at 1.51 and 1.14 Ma (Spell et al. 1990).

The original volume of the CdR is difficult to estimate; the original extent was probably not less than 600 km^2. The thickness varies from >300 m in White Rock Canyon to <10 m near the southern extremity, and individual flows vary by as much as 2 m to 100 m in response to pre-existing topography. An average thickness of 100 m may be a reasonable estimate for the whole field, in which case the total volume is of the order of 60 km^3.

The northern CdR, which consists of lavas accompanied by phreatomagmatic and strombolian ash and scoria deposits, is divided into eastern and western portions by the Rio Grande drainage in White Rock Canyon. Due to the width of the canyon and extensive slumping on either wall, correlation across the river is difficult and systematic spatial and temporal variations in erupted magma composition have therefore not been fully evaluated. However the following points are clear:

1. Tholeiites are virtually confined to west of the river; in the northern part of the study area (Fig. 1), they rest on gently east-dipping depositional slopes within the Puye alluvial fan, and post-date most of the flows in the main CdR to the south. Vents must lie in the buried portion of the CdR, beneath Bandelier Tuff.
2. There is some tendency toward eruption of more evolved compositions with time east of the Rio Grande, inasmuch as the youngest lavas, which form the small dome-and-lobe edifice of Ortiz Mountain in the northern CdR (Fig. 1), are also the most silicic; however there is no apparent stratigraphic control on magma composition through most of the sequence.

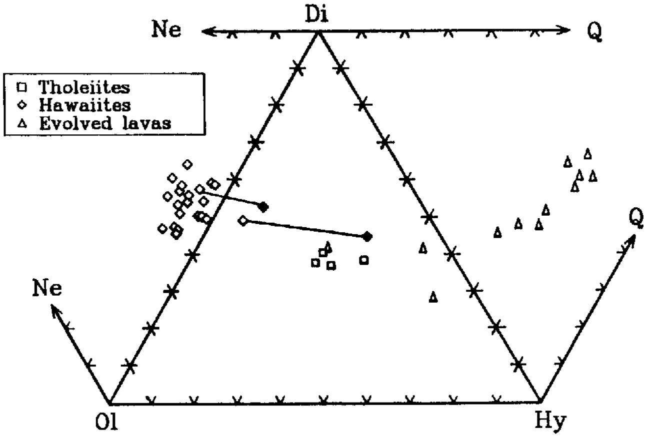

Fig. 2. Normative compositions of CdR samples in the basalt trapezium. Samples with significant modal xenocrystic quartz were corrected for SiO$_2$ content before plotting. The effect on the two most quartz-rich samples is shown by lines connecting analysed compositions (*solid diamonds*) with corrected quartz-free compositions (*open diamonds*)

Classification

When applied to the CdR lavas, formal nomenclatural schemes (e.g. Le Bas et al. 1986) result in a plethora of rock names which obscure compositional similarities and differences. Analysed lavas fall naturally into three groups in terms of their positions in the basalt trapezium (Fig. 2), their trace element contents (Table 1), and their isotopic compositions, and we adopt a convenient classification on this basis. *Hawaiites* refer to nepheline-normative lavas; all plot close to the critical plane of silica undersaturation (Fig. 2). This usage includes samples that should more properly be referred to as alkali basalts, and also one weakly hypersthene-normative composition. All have between 3.9 and 7.1% MgO, and share trace element characteristics that do not correspond to any globally-recognised magma type. Rocks described here as *tholeiites* are olivine-hypersthene normative, have 50.5–51.5% SiO$_2$, and >6.5% MgO (volatile-free). These lavas also differ from both other groups in that they all have chondrite-normalised K/Th ratios (K$_N$/Th$_N$) >1. The third group consists of strongly hypersthene-normative (and usually Q-normative) rocks with >52.5% SiO$_2$ and <6.5% MgO; they have generally evolved compositions with weakly alkaline affinities, and similar trace-element characteristics to hawaiites, although there is a tendency for K/Na to increase with silica content. We refer to these as *evolved* lavas.

Petrography

General

Hawaiites are characterized by euhedral to subhedral olivine phenocrysts, with occasional quench-induced skeletal overgrowths, set

Table 1. Whole-rock major element, trace element, Sr-, Nd- and O-isotope data for representative NW Cerros del Rio lavas. $Fe_2O_3^*$ denotes total iron as Fe_2O_3. See Appendix for analytical details. T = tholeiite, H = hawaiite, E = evolved lava. Samples TA33, 6-26C, I-5 and SF-10 contain significant amounts of xenocrystic quartz. The whole-rock analyses, including quartz, are reported in this table

	T 6-15G	T 6-7B	H TA33	H 6-26C	H I-4	H I-5	H I-6	H SF-5	H SF-10	E 6-15H	E SF-3	E 6-19a	E I-0	E 0-2
SiO_2	50.79	50.53	53.55	52.86	50.31	52.13	49.19	51.98	49.36	54.26	52.31	57.93	63.37	59.91
TiO_2	1.45	1.40	1.25	1.48	1.61	1.48	1.58	1.32	1.48	1.27	1.32	0.86	0.77	0.97
Al_2O_3	16.15	15.75	15.57	15.00	16.75	17.28	15.48	15.58	16.74	17.50	16.10	15.00	16.10	16.89
$Fe_2O_3^*$	11.22	11.25	7.88	8.55	9.64	8.81	12.91	8.52	10.38	8.02	9.75	7.07	5.45	6.23
MnO	0.16	0.17	0.14	0.14	0.15	0.13	0.16	0.15	0.18	0.13	0.16	0.12	0.08	0.10
MgO	6.68	7.48	5.65	6.13	5.60	4.39	5.55	6.76	4.75	4.04	4.69	6.19	2.19	2.89
CaO	9.01	8.72	7.59	7.90	8.80	7.98	9.62	8.02	9.69	6.75	8.50	6.97	5.13	5.71
Na_2O	3.38	3.20	4.29	3.65	4.53	4.69	3.68	4.21	3.79	4.50	3.85	3.85	4.59	4.65
K_2O	0.98	1.00	2.46	2.25	1.71	2.06	0.83	2.04	1.32	1.94	1.19	1.79	2.67	2.44
P_2O_5	0.30	0.30	0.94	0.73	0.79	0.88	0.47	0.78	0.63	0.69	0.62	0.41	0.43	0.52
LOI	−0.06	−0.07	0.07	0.89	0.31	0.41	0.68	0.58	2.10	0.01	1.65	0.44	0.66	0.46
Total	100.06	99.73	99.39	99.58	100.20	100.24	100.15	99.94	100.42	99.11	100.14	100.63	101.44	100.77
Rb	15	16	39	33						26		26		
Be			1.84		1.74	1.81	1.59	1.49	1.6		1.44		1.72	1.76
Sr	447	431	1162	927	1150	1279	528	1018	770	1011	740	790	784	921
Ba	433	416	1734	1222	540	706	287	777	471	1222	500	1084	794	734
Ni	91	108	91	106	33	35	70	119	28	23	51	130	43	9.6
Co			36.9		29.8	22.2	40.3	27.4	35.5		26.8		12.7	17
Cu	67	72	31	37	46	50	73	39	48	29	54	35	25	24
Zn	88	97	55	80	71	69	93	61	68	86	67	74	55	55
V	174	185	140	173	168	156	253	136	199	151	102	120	77	97
Cr	220	240	114	165	86	65	185	128	20	11	127	275	39	19
Sc	25.9	25.5	15.2	17.9	20.5	17.8	25.7	17.2	20.8	13.8	20.5	18.1	10.1	10.7
Y	29	28	27	27						27		24		
La	18	20.1	92.2	56.6	63.5	82.4	25.3	73.3	45.2	69.6	49.8	41.8	59.9	70
Ce	44	44	159	103	134	142	55	125	90	112	87	79	107	104
Nd	23	21	62	43	49	56	27	53	36	45	38	32	39	40
Sm	5.2	4.8	9.6	7.4	8.1	9.1	5.6	8.6	6.7	7.9	6.8	5.7	6.1	6.6
Eu	1.73	1.63	2.72	2.09	2.41	2.61	1.84	2.34	2.11	2.22	2.01	1.57	1.74	1.83
Tb	0.89	0.95	1.11	0.91	1.14	1.09	0.96	1.06	1.04	0.96	0.82	0.67	0.81	0.87
Yb	2.28	2.23	2.31	1.74	2.2	2.56	2.1	2.38	1.95	1.96	2.28	1.56	1.83	1.83
Lu	0.37	0.32	0.28	0.27	0.29	0.33	0.31	0.33	0.29	0.32	0.3	0.3	0.26	0.27
Zr	144	138	222	190	174	200	118	150	135	216	126	176	194	138
Hf	3.6	3.4	5.7	4.6	4.7	5.9	3.4	5.6	4.2	5.4	4.6	4.7	5.5	5.9
Th	2.7	2.7	15.2	8.6	8.4	11.4	3.0	10.5	6.7	12	7.8	6.8	10.7	8.7
U	0.5	0.6	4.0	2.1	1.8	2.3	0.6	2.1	1.3	2.9	1.2	1.4	2.3	1.9
Nb	19	16	47	42						35		28		
Ta	1.06	1.04	2.76	2.71	2.06	2.67	1.12	2.28	1.65	2.25	1.71	1.55	1.51	1.59
$^{87}Sr/^{87}Sr$	0.704769	0.704936	0.704165	0.704127	0.704121	0.704128	0.704245	0.704252	0.704259	0.704201	0.704293	0.704211	0.704338	0.704216
$^{143}Nd/^{144}Nd$	0.512598	0.512617	0.512669	0.512683	0.512690	0.512672	0.512698	0.512664	0.512705	0.512653	0.512686	0.512584	0.512576	0.512568
$\delta^{18}O$	6.3	6.7	6.4	6.6	6.1	6.4	6.5	6.6	6.6	6.6	6.6	6.9	7.0	7.2

Table 2. O- and H-isotope data. * denotes average of two runs. Estimated $\delta^{18}O$ melt refers to the O-isotope composition of silicate liquids calculated (using the fractionation factors of Bottinga and Javoy 1975) to be in equilibrium at 1000° C with the phases in parentheses, for which data are given in the following two columns. Note that estimated melt $\delta^{18}O$ values are in good agreement with measured whole-rock values for the same samples, while those for quartz are in poor agreement

Sample Number	SiO_2 (wt%)	WR $\delta^{18}O$ (‰ SMOW)	Estimated $\delta^{18}O$ Melt (‰SMOW)	Qtz $\delta^{18}O$ (‰SMOW)	Other $\delta^{18}O$ (‰SMOW)	δD (‰SMOW)	H_2O^+ (wt%)
I-6	49.19	+6.5±0.3*	+8.4 (qtz), +6.3 (fsp)	+9.5	+6.5 (fsp)	−107	0.59
SF-10	49.36	+6.6±0.3*				−106	0.68
6-9E	49.37	+6.1					
I-4	50.31	+6.1					
6-7B	50.53	+6.7				−86	0.58
6-15G	50.79	+6.3					
SF-5	51.98	+6.6±0.1*					
I-5	52.13	+6.4					
SF-3	52.31	+6.6				−89	0.60
BAYO	52.75	+6.6					
6-26C	52.86	+6.6	+9.5 (qtz), +6.6 (ol)	+10.6	+5.2 (ol)	−98	0.71
TA33	53.55	+6.4	+7.4 (qtz)	+8.4		−107	0.49
6-15H	54.26	+6.6±0.1*					
6-19A	57.93	+6.9±0.0*	+6.5 (px)				
6-8B	58.03	+6.6					
6-8A	59.63	+7.4±0.1*	+5.6 (qtz)	+6.7			
O-2	59.91	+7.2±0.1*					
SF-2	62.43	+6.7±0.0*	+6.2 (qtz) +6.9 (px)	+7.3	+6.2 (px)		
I-0	63.37	+7.0±0.1*	+6.9 (px)		+6.3 (px)	−103	0.30

in a plagioclase-dominated, sometimes trachytoidal, groundmass with variable amounts of glass. Augite and plagioclase microphenocrysts appear in lavas with relatively low MgO contents. Occasional plagioclase grains appear strongly resorbed.

Tholeiites, likewise, are olivine-phyric and contain occasional plagioclase microphenocrysts, set in a sub-ophitic groundmass.

Evolved lavas very considerably; the most mafic resemble hawaiites. With increasing silica content, hypersthene and hornblende join the phenocryst assemblage and groundmass glass becomes more abundant. Hornblendes in otherwise fresh, glass-bearing rocks are variably to completely replaced by a cryptocrystalline near-opaque mass, a common feature in silicic lavas that is attributed to breakdown of the hydrous mineral during eruptive degassing. Euhedral plagioclase is a more abundant phenocryst; however many samples contain a second population of strongly resorbed plagioclase.

Secondary alteration

Samples vary from being completely fresh to having advanced iddingsitisation of olivines. Other crystalline phases are never affected in the samples which form the basis for this study, as those showing heavier alteration were rejected. Olivine alteration (with probable minor hydration of groundmass glass) is reflected in appreciable measured bulk-rock water contents (Table 2). There is no correlation between degree of alteration and geochemistry; furthermore the H-isotope composition of a range of samples rules out significant post-eruptive, low-temperature modification.

Quartz xenocrysts

Quartz xenocrysts occur in hawaiites and evolved lavas as rounded to subangular, frequently embayed, grains up to 2.5 mm in diameter. Most are optically homogeneous and monocrystalline, although rare composite grains are seen. Some are partly annealed stressed fragments and one was seen to contain oriented needles of rutile. Characteristically, quartz is rimmed by a reaction jacket of radiating clinopyroxene needles, sometimes separated from the xenocryst by a thin coating of very pale brown glass.

Quartz xenocrysts have been previously reported in CdR lavas by Lipman (1969), Baldridge (1979), and Aubele (1978, 1979), and are known in lavas throughout the RGR and adjacent regions (Larsen et al. 1938; Doe et al. 1969; Stormer 1972; Lipman and Mehnert 1975). Two explanations have been offered: (1) that the quartz is a remnant of high-pressure crystallisation, and is therefore cognate; (2) that the quartz grains are crustally-derived xenocrysts. Cognate quartz is a theoretical possibility because of its stability in basalt and basaltic andesite at mantle pressures (Nicholls et al. 1971), and it has been suggested to occur at Lassen Peak in California (Smith and Carmichael 1968) and in low-$^{87}Sr/^{86}Sr$ basaltic lavas of Talasea, New Britain (Lowder and Carmichael 1970; Peterman et al. 1970).

Foreign quartz may be incorporated into relatively mafic magmas by assimilation of granitic or sedimentary crustal material, or by incomplete mixing of mafic and silicic magmas prior to eruption. Although there are no eruptives with quartz phenocrysts present in the CdR, a possible source of silicic magma available for mixing might be the rhyolitic magma chamber which had developed prior to 1.78 Ma (Spell et al. 1990) some 40 km westwards beneath the Jemez Mountains, and was contemporaneous with the later CdR lavas.

The oxygen isotopic compositions of quartz in CdR lavas (Table 2) point to a xenocrystic origin (see Stable isotopes section).

Major and trace element geochemistry

Perhaps the most immediately obvious feature of the chemical data is the general lack of any well-defined compositional trends, despite close spatial and temporal association of the lavas. Plots of compatible elements versus silica (Fig. 3) define broad arrays with roughly linear upper boundaries, such as can be generated by random mixing between magmas lying at all stages along

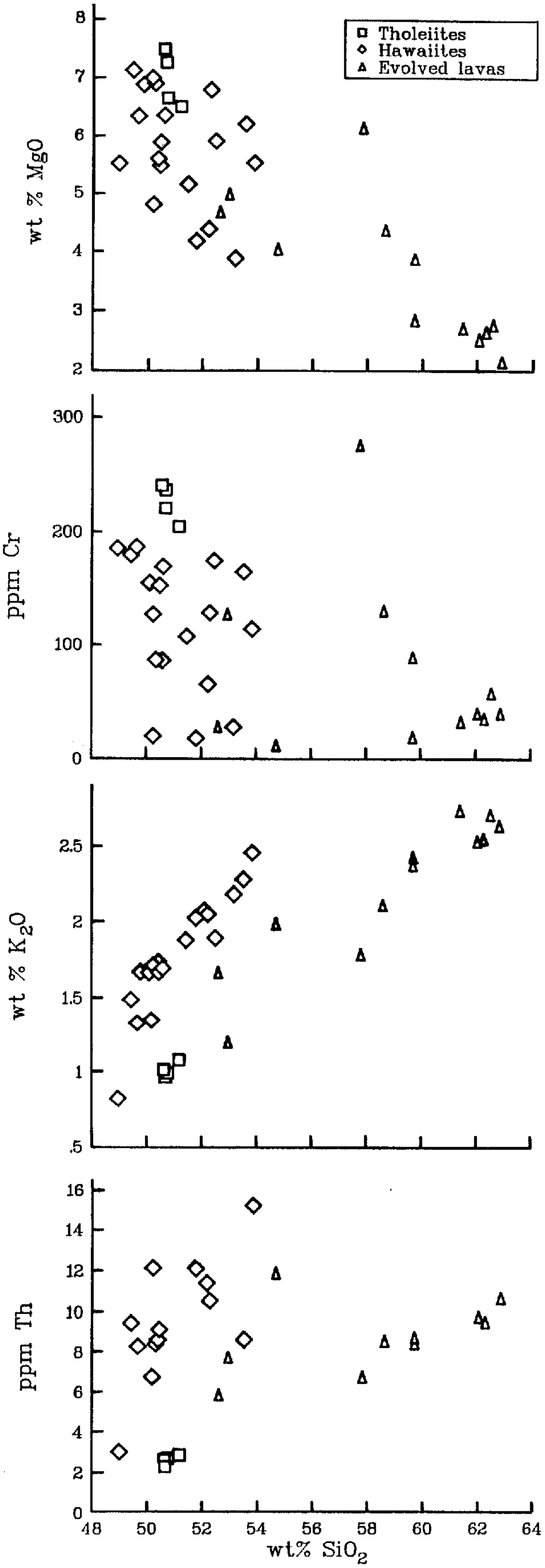

Fig. 3. Variation diagrams vs SiO_2 for MgO, K_2O, Cr and Th, uncorrected for xenocrystic quartz content

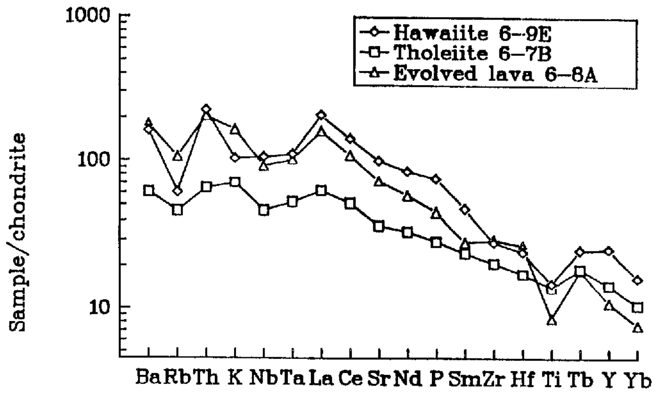

Fig. 4. Chondrite-normalised incompatible trace-element plots (spidergrams) for a representative tholeiite, hawaiite, and evolved lava. Normalisation factors of Thompson et al. (1984) are used

a particular fractionation path from a single ultimate parental composition. While magma mixing may have been an important process in the genesis of individual melt batches, it is clear from the extreme variation in incompatible trace element abundances at constant silica contents (Fig. 3) that more than one "parental" melt was available, reflecting either variable mantle source compositions, variable mantle melt fractions, or different styles and/or degrees of interaction between mantle-derived magmas and crust.

Chondrite-normalised incompatible trace element plots (spidergrams) provide qualitative information on the range and types of magmas present; representative patterns for CdR magmas are shown in Fig. 4. For the range of compositions considered here, fractional crystallisation has little effect on spidergram shapes beyond relative depletion of Sr and Ti by removal of plagioclase and titanomagnetite respectively. Although the general effect of fractionation is to increase slopes of patterns, highly incompatible elements of interest (Ba, Rb, Th, K, Nb, Ta, La) are not significantly separated from each other until the felsic condition is attained, when the appearance of alkali feldspar and accessory rare-metal phases may drastically modify ratios between these elements. Some of these phases can also appear as phenocrysts in very strongly silica-undersaturated mafic magmas such as nephelinites. But the CdR lavas are neither ultra-leucocratic nor strongly undersaturated, and variations in the ratios of strongly incompatible elements should therefore reflect either variable magma source chemistry and/or mineralogy, or assimilation of crust by rising mafic magmas.

Tholeiites have generally smooth patterns with small troughs at Nb and Ta; they otherwise resemble "enriched" mid-ocean ridge basalts (E-MORB) or tholeiitic OIB. The Nb–Ta depletion could indicate a minor arc-source type component in the tholeiite source, but the relatively low concentrations of Ba, Rb and Th with respect to LREE (e.g. $Ba_N/La_N = 1$, Fig. 4) argues against this. A more likely explanation is contamination of E-MORB or OIB-type tholeiitic magmas with continental crust, which is required by the O, Sr and Nd isotopic data.

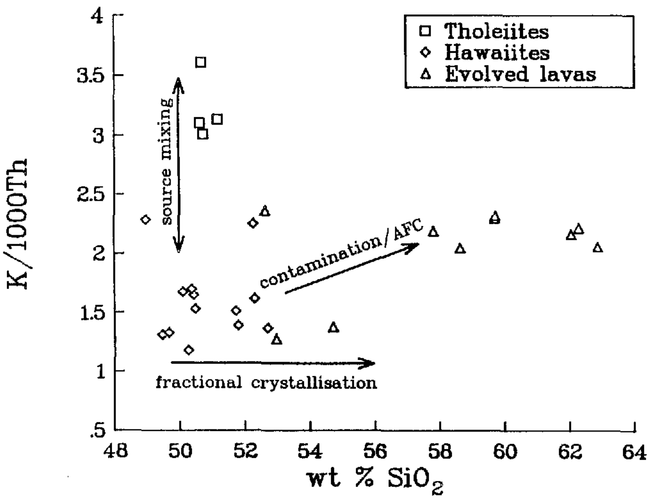

Fig. 5. K/1000Th vs SiO_2 corrected for xenocrystic quartz content, showing the expected effects of crustal contamination of hawaiites, fractional crystallisation, and mixing of mantle sources

Hawaiites have very distinctive spidergrams with pronounced peaks at Th and marked relative depletions in Nb, Ta and K and Rb (Fig. 4). This is in marked contrast to the behaviour of these elements in common magma types, which exhibit either depletion in Nb and Ta with enrichment in K and Rb (arc magmas and/or crustally contaminated magmas), or Nb – Ta enrichment with depletion in K and Rb (OIB magmas, especially silica-undersaturated varieties); recent chemical geodynamic models (Hofmann 1988; Saunders et al. 1988) have emphasised the complementary nature of these common signatures. Crustal contamination of CdR hawaiites (see discussion) cannot explain their K-depleted nature, since the general effect of contamination would be to increase K/Ta compared to common mantle-derived magma types. Leat et al. (1988) recognise the same signature in mafic magmas erupted from the Colorado section of the RGR, suggesting a regional mantle source of unusual composition (or residual mineralogy during melt extraction).

Evolved lavas have similar patterns to the hawaiites but are less K-depleted, although K_N/Th_N and K_N/La_N are still typically <1. This similarity and their Sr – Nd isotope characteristics suggest that they represent batches of hawaiite magma contaminated with continental crust (see discussion). Two samples with comparable K/Th to the hawaiites have the lowest SiO_2 contents and $^{87}Sr/^{86}Sr$ ratios, and the highest $^{143}Nd/^{144}Nd$ ratios, among the evolved lavas (Table 1, Figs. 5, 6).

Radiogenic isotopes

Sr and Nd isotope ratios vary systematically among the three compositional groups, although the overall range in values is small (Table 1, Fig. 6). The hawaiites and evolved lavas exhibit a striking negative correlation between $^{143}Nd/^{144}Nd$ and SiO_2; this trend is also apparent among the hawaiites alone (Fig. 7). This is clear evi-

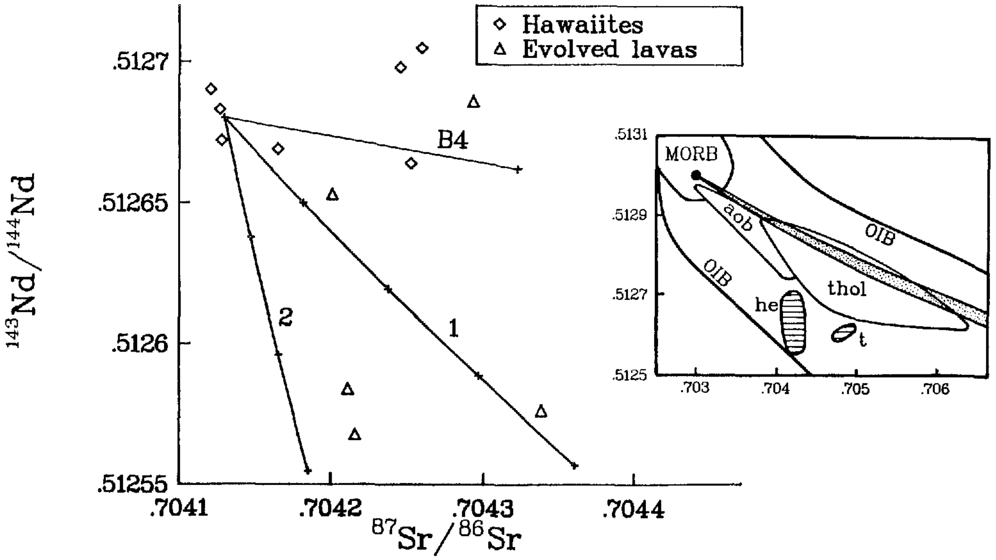

Fig. 6. Sr – Nd isotope ratios of CdR lavas, with mixing lines between typical hawaiite (Sr = 1100 ppm, $^{87}Sr/^{86}Sr = 0.70413$, Nd = 50 ppm, $^{143}Nd/^{144}Nd = 0.51268$) and crust. *Curve B4* is for contamination with basement granite B4 (Table 3), tick at 3% contamination. *Curves 1* and *2* are for contamination with possible felsic lower crustal equivalents of the granites, having the same compositions as B4 and B8 respectively, but with $^{87}Sr/^{86}Sr = 0.7070$. *Curve 1*: ticks at 5%, 10%, 15% and 20% contamination; *curve 2*: ticks at 4%, 8% and 12% contamination. *Inset*: CdR lavas (horizontal rule; *he* = hawaiites and evolved lavas, *t* = tholeiites) compared to alkali olivine basalts (field aob) and tholeiites (field thol) from the RGR and adjacent regions analysed by Perry et al. (1987), whose $^{143}Nd/^{144}Nd$ values have been recalculated to a bulk earth value of 0.512638. Generalised limits of the fields occupied by MORB and OIB are also shown. *Stippled field* bounded by curves is for contamination of a hypothetical mantle-derived magma similar to E-MORB (*filled circle*: Sr = 400 ppm, $^{87}Sr/^{86}Sr = 0.7030$, Nd = 20 ppm, $^{143}Nd/^{144}Nd = 0.5130$), and to Perry et al.'s isotopically most primitive composition, with up to 17% of granites B4 and B8 (Table 3)

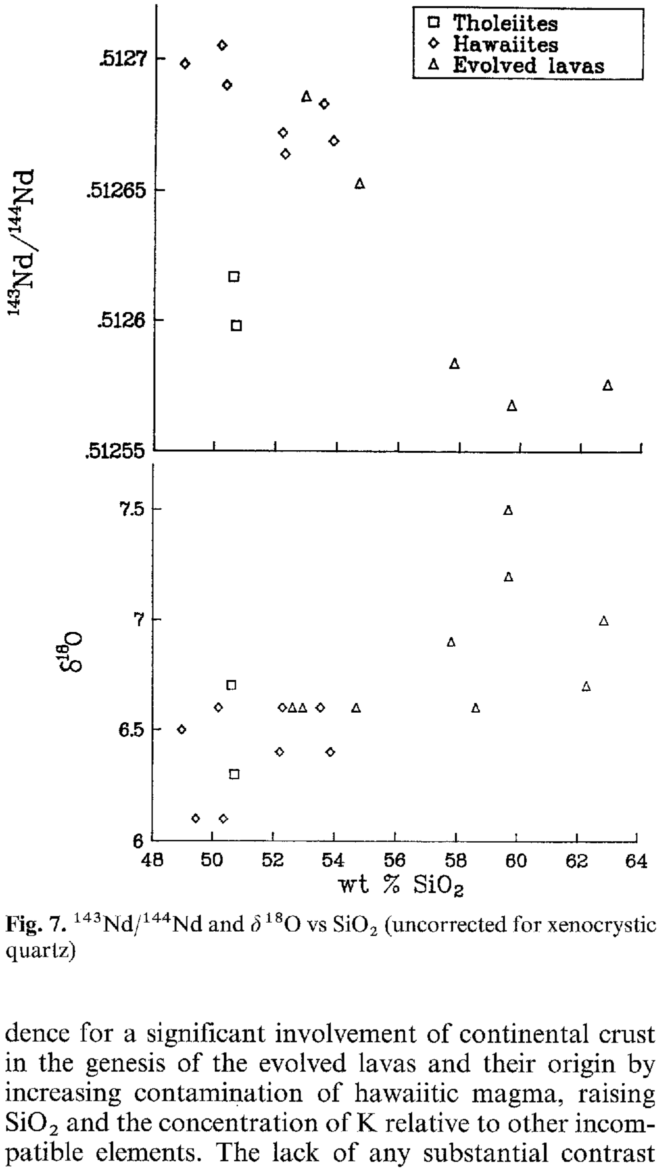

Fig. 7. $^{143}Nd/^{144}Nd$ and $\delta^{18}O$ vs SiO_2 (uncorrected for xenocrystic quartz)

dence for a significant involvement of continental crust in the genesis of the evolved lavas and their origin by increasing contamination of hawaiitic magma, raising SiO_2 and the concentration of K relative to other incompatible elements. The lack of any substantial contrast in $^{87}Sr/^{86}Sr$ between the two groups (Fig. 6) requires that the contaminant have low time-integrated Rb/Sr despite LREE enrichment; Precambrian felsic lower crust is the most likely candidate.

$^{87}Sr/^{86}Sr$ is highest in the tholeiites (Table 1, Fig. 6), reflecting a regional distinction between lavas of tholeiitic and alkaline affinity (Perry et al. 1987). This regional variation may simply reflect the relative sensitivities of the Sr, Nd-poor tholeiites and Sr, Nd-rich alkaline magmas to contamination, particularly with upper crust.

Stable isotopes

Many previous studies (e.g. Taylor 1968; Johnson and O'Neil 1984; Cerling et al. 1985; Harmon et al. 1987) have documented that the O-isotope compositions of originally glassy volcanic rocks more than a few thousand years old are frequently modified by secondary processes such as sub-solidus hydrothermal alteration or low-temperature hydration. Therefore, $\delta^{18}O$ values were measured for both whole rocks and, where possible, for separated minerals. Generally, D/H ratios are expected to be be more readily affected by alteration than $^{18}O/^{16}O$ ratios because the external fluid phase is the predominant source for hydrogen, whereas the silicate rock is the primary oxygen reservoir except at very high water/rock ratios. Therefore, δD values and high-precision H_2O^+ contents were also determined for a subset of representative samples spanning the compositional range of the CdR lavas.

The range of measured $\delta^{18}O$ values is just over 1‰, ranging from +6.1‰ among the hawaiites to +7.4‰ in the evolved lavas. The CdR lavas have modest water contents of 0.3 to 0.7 wt.%, a limited range of H-isotope variation with $\delta D = -107$ to -86‰, and no covariation of δD or $\delta^{18}O$ with H_2O^+ content (Table 2). The $\delta^{18}O$ values for the olivine, pyroxene and feldspar mineral separates (Table 2) fall within the general range expected for these phases crystallising from mafic to intermediate magmas at ~1,000° C and calculated magmatic $\delta^{18}O$ values agree, within a few tenths of a part per thousand, with measured whole rock $\delta^{18}O$ values (Table 2). Together these features are interpreted as confirming the pristine nature of the CdR lavas and thus permit ^{18}O variations within the suite to be interpreted in terms of magmatic processes.

MORB tholeiites have a very restricted range of $\delta^{18}O$ variation. For example, the 52 pristine MORB glasses analyzed by Ito et al. (1987) have a $\delta^{18}O$ range of +5.35‰ to +6.05‰ with an average of $+5.7 \pm 0.2$‰. OIB tholeiites are slightly ^{18}O depleted at $+5.4 \pm 0.2$‰ (Kyser et al. 1982), whereas OIB and continental alkalic basalts have more variable O-isotope compositions but are usually ^{18}O-enriched relative to MORB (Harmon and Hoefs 1984). All analyzed CdR lavas are are more ^{18}O rich than MORB and most OIB tholeiites (Table 2). The two analyzed tholeiites are enriched in ^{18}O by some 0.8‰ relative to MORB, whereas the hawaiites ($\delta^{18}O = +6.1$ to $+6.6$‰) exhibit the same general range as alkalic OIBs. The evolved lavas tend to be more ^{18}O rich than the hawaiites, with $\delta^{18}O$ exhibiting a sympathetic variation with increasing SiO_2 content (Table 2, Fig. 7). This feature and the clear negative correlation of $\delta^{18}O$ values with $^{143}Nd/^{144}Nd$ for the alkalic lavas is consistent with the presence of a substantial crustal component in these rocks.

Origin of quartz

If the quartz is of cognate magmatic origin, its $\delta^{18}O$ values should be relatively low (~ +7.0‰ to +8.0‰) and should exhibit an equilibrium $^{18}O/^{16}O$ fractionation of O isotopes with coexisting mineral phases. Crustally-derived xenocrystic quartz, however, should be characterised by relatively high $\delta^{18}O$ values (because crustal rocks and minerals which have experienced a cycle of surface residence tend to be relatively enriched in ^{18}O), and be out of O-isotope equilibrium with coexisting minerals in the same rock.

Quartz samples separated from five CdR lavas (hawaiites and evolved lavas) have $\delta^{18}O$ values between

+6.7‰ and +10.6‰ (Table 2), a range almost four times that exhibited by the quartz-bearing lavas themselves. Also, plag-qtz, px-qz, and ol-qz values are distinctly different from mineral fractionations observed experimentally at ~1000° C (Chiba et al. 1989). This O-isotope disequilibrium is compelling evidence that the quartz grains in the CdR lavas are of extraneous origin. The three highest $\delta^{18}O$ values of +8.7‰ to +10.6‰ are typical of ^{18}O-rich upper crust and suggest derivation from high-grade metasedimentary or granitic crustal rock. The contaminant is not, however, a single lithology of homogeneous O-isotope composition. The quartz is unlikely to have been introduced via mixing with rhyolitic magmas similar to those present in the Jemez Mountains, which have quartz $\delta^{18}O$ values in a narrow range between +7.5‰ and +8.1‰ (R.S. Harmon and J.A. Wolff, unpublished data). Wavy extinction, recrystallisation and oriented rutile needles in some grains indicate introduction via assimilation of metamorphic rock, or of a first-cycle sediment derived from a metamorphic source, such as the rift-filling Santa Fe Group in the Española Basin. It is probable, however, that the two quartz samples with the lowest $\delta^{18}O$ values of +6.4‰ and +7.3‰ (both from evolved lavas) were derived directly from a high-grade crustal protolith.

Discussion

Crustal contamination: hawaiites and evolved lavas

Several lines of evidence indicate the importance of crustal contamination in the genesis of CdR magmas. Although there is no substantial body of trace element and isotopic data for the local crust (especially lower crust), which precludes any meaningful quantitative modelling of assimilation processes, some general conclusions can be reached by making assumptions about crustal geochemistry, and/or through modelling based on "typical" crustal compositions from the literature. We consider the assumptions made basic and very reasonable. They are: (1) that both the lower and upper crust beneath the RGR have $^{143}Nd/^{144}Nd$ considerably less than the bulk earth value; (2) that both upper crust, with radiogenic Sr, and old Rb-, Th-depleted lower crust with relatively nonradiogenic Sr may be present (cf. Dungan et al. 1986); (3) that the continental crust is enriched in ^{18}O relative to the mantle; (4) that all crust has $K_N/Th_N > 0.7$ (the maximum value found in typical hawaiites); (5) that all crust, especially the upper crust, is depleted in Nb and Ta relative to LILE with respect to bulk earth.

The alkaline affinities (mugearite, benmoreite, latite) of the evolved lavas link them with the hawaiites rather than the tholeiites. Also, both hawaiites and evolved lavas are found throughout the northern CdR, whereas tholeiite vents are areally restricted to west of the river. Hawaiites and evolved lavas have similar trace element patterns (Fig. 4), the increase of K/Th and K/Ta with SiO_2 being the most notable variations that cannot be ascribed to fractional crystallisation. The most silicic lavas generally have the highest K/Th, the lowest $^{143}Nd/^{144}Nd$, and the highest $\delta^{18}O$ values (Figs. 5, 7). On the Nd–Sr isotope diagram, these lavas define a trend of sharply decreasing $^{143}Nd/^{144}Nd$ with little change in $^{87}Sr/^{86}Sr$ (Fig. 6). Although contamination with Sr-poor upper crust cannot be ruled out, it is extremely unlikely because present-day $^{87}Sr/^{86}Sr$ of upper crust in the region ranges up to values in excess of 0.8 (Vuataz et al. 1988), and therefore the presence of even a strongly Sr-depleted potential contaminant should be apparent. Sr–Nd isotope data for two Proterozoic basement granites from the Jemez and Nacimiento Mountains west of the Española Basin are given in Table 3; the samples are both radiogenic in Sr and are not strongly Sr-depleted. Mixing lines between typical hawaiite and granites pass well above evolved lavas on the Sr–Nd diagram (Fig. 6). While isotopic variation among the hawaiites can be accounted for by very small degrees (2–3%) of granite assimilation, the hawaiite-evolved trend can only be modelled by assimilation-fractional crystallisation (AFC) involving the granite compositions if the bulk distribution coefficient for Nd is much greater than that for Sr, which we consider unlikely. It is more probable that the observed trend represents a mixing line between hawaiite and lower crust with relatively nonradiogenic Sr. Two such lines are shown in Fig. 6, assuming that the lower crust contains felsic lithologies similar to the granites, but with lower $^{87}Sr/^{86}Sr$. If the observed hawaiites are representative of magmas that interacted with crust to produce the more silicic compositions, then the generally lower abundances of incompatible elements in the latter (Table 1) require mixing between relatively lithophile element-rich mantle-derived alkaline magmas and lithophile element-poor lower crustal melts, with only a limited role for fractional crystallization in generating the higher K/Th silicic lavas with silica contents of around 58%. Fractional crystallisation and/or AFC may however have operated among the evolved compositions above about 58% SiO_2, since all of these samples fall within a narrow range of K/Th (Fig. 5), and are in any case too silicic to have been generated by simple addition of granitic (75% SiO_2) material in the amounts (max. 15%) suggested by the Sr–Nd isotopic data. The generally small isotopic shifts produced by contamination are a consequence of the trace element-rich nature of the mantle-derived alkaline melts.

Table 3. Sr-Nd isotope data for two Proterozoic basement granite samples. Sample B4 is from the Guadalupe Box, SW Jemez Mountains, and B8 is a leucogranite from Joaquin Canyon, Nacimiento Mountains. The samples were collected 50 km and 60 km respectively west of the NW CdR

Sample	B4	B8
Rb, ppm	112.0	181.7
Sr, ppm	394.3	164.8
Sm, ppm	8.82	10.49
Nd, ppm	42.94	60.56
$^{87}Sr/^{86}Sr_{(2 Ma)}$	0.72216	0.76806
$^{143}Nd/^{144}Nd$	0.511983	0.511797

Two of the evolved lavas have trace element and isotopic characteristics closely similar to the hawaiites (Figs. 5–7), despite slightly higher silica contents and hy-normative (albeit not Q-normative) compositions (Fig. 2). They are among the lavas with the lowest K/Th values (Fig. 5). We regard these as representing the products of near-pure fractional crystallisation of hawaiitic magma, with little or no involvement of continental crust during this fractionation interval.

Without precise information on contaminant lithologies, we consider more detailed modelling of contamination processes to be unwarranted at this stage. Nonetheless some significant points regarding components contributing to the CdR hawaiites and evolved lavas can be made. The well-correlated trend for hawaiites – evolved lavas in $^{143}Nd/^{144}Nd$ vs SiO_2 (Fig. 7) suggests that none of the analysed CdR lavas of alkaline affinity have escaped the effects of crustal contamination; the most silicic lavas are the most contaminated. The hawaiites, having low K/Th values, cannot have been produced by crustal contamination from some more familiar mantle-derived magma with high K/Th; also, the lavas with the lowest K/Th generally have the highest $^{143}Nd/^{144}Nd$, and are therefore the least contaminated. Regardless of what is assumed about K and Th distributions in the continental crust, this alone is compelling evidence that the low-K/Th component source is located in the mantle beneath the Española Basin.

Contamination and origin of CdR tholeiites

Perry et al. (1987) presented Sr–Nd isotope data on basalts from the RGR and adjacent regions to the west, and showed that tholeiitic magmas are displaced to higher $^{87}Sr/^{86}Sr$ and lower $^{143}Nd/^{144}Nd$ compared with associated alkali olivine basalts. This is considered to be a consequence of the generally lower Sr and Nd contents of the tholeiites, rendering the isotopic ratios of these elements more susceptible to contamination. The CdR tholeiites conform to this behaviour; their Sr–Nd isotope characteristics suggest they carry an upper crustal component. They cannot be modelled as contaminated equivalents of magmas related to the hawaiites, because the very low K/Th ratios of the latter preclude the generation of a composition with $K_N/Th_N > 1$ by this process, even if extremely Th-depleted crust is invoked as a contaminant (Fig. 5). Any upper crustal contaminant is very unlikely to be Th-depleted. Tholeiite trace-element spidergrams generally resemble E-MORB or tholeiitic OIB, apart from the slight Nb–Ta depletion (Fig. 4). Mixing lines between enriched oceanic tholeiite compositions and the local granites pass through the field of Sr–Nd isotope compositions of regional tholeiites determined by Perry et al. (1987), but do not satisfactorily model CdR tholeiites (Fig. 6). Again, an AFC process is only viable in the unlikely event of the bulk distribution coefficient for Nd being significantly greater than that for Sr during fractional crystallisation. Either the analysed granites are more radiogenic in Nd than contaminants encountered by the tholeiitic magmas during their ascent, or – which we consider to be more probable – the tholeiites were contaminated by lower crust with nonradiogenic Sr prior to interacting with the upper crust. The lack of chemical variations among the analysed tholeiites precludes identification of mantle-derived components on the basis of trend extrapolation, as is possible for the hawaiites. In a regional context (see below), the simplest explanation of CdR tholeiite geochemistry is that they are asthenospherically-derived liquids, closely resembling common ocean-basin magmas, contaminated with small amounts of upper (and probably also lower) continental crust.

Source mixing

Two of the hawaiites and one evolved sample have trace element abundances and ratios intermediate between tholeiites and typical hawaiites. These samples cannot be modelled by addition of any feasible crustal composition to either mantle component, and one of them (sample I-6) has the highest $^{143}Nd/^{144}Nd$ value of the entire suite, despite rather low abundances of Nd and other trace elements, which would render it particularly susceptible to crustal contamination. More probably, these lavas represent mixtures between hawaiite-source and tholeiite-source components. The mixing mechanism (magma mixing vs partial melting of a heterogeneous mantle source region) cannot be determined, although Leat et al. (1988, 1989) have argued on dynamic grounds for the similar NW Colorado lavas that, if one component is located in the lithosphere, magma mixing is more likely. The evolved lava with similar K/Th (Fig. 5) is considered to be a differentiated equivalent. Note that the most silicic evolved lavas cannot be derived from these compositions despite their similar K/Th; the contrast in $^{143}Nd/^{144}Nd$ requires accompanying crustal assimilation, which would act to raise K/Th.

Origin of the hawaiite K-depleted component

If our inference that none of the CdR lavas have escaped crustal contamination is correct, then the hawaiite source component lies within the Sr–Nd mantle array and does not have a time-integrated Rb/Sr–Sm/Nd history that can be distinguished from OIB or arc-source magmas. Nonetheless the distinctive and unusual depletions in Rb, K, and Nb, Ta require a mantle type or component, or melt generation process, not commonly recognised. The hawaiite signature could be the result of an uncommon residual phase assemblage during partial melting, or of chemical modification to some initially "normal" mantle type prior to magmagenesis. We will evaluate each of these in turn.

Effect of residual phases. Generation of the hawaiite trace-element characteristics from partial melting of some common mantle type would require a residual phase assemblage capable of retaining Nb, Ta, and/or K and Rb (and to a lesser extent Ba), depending on the postulated mantle type.

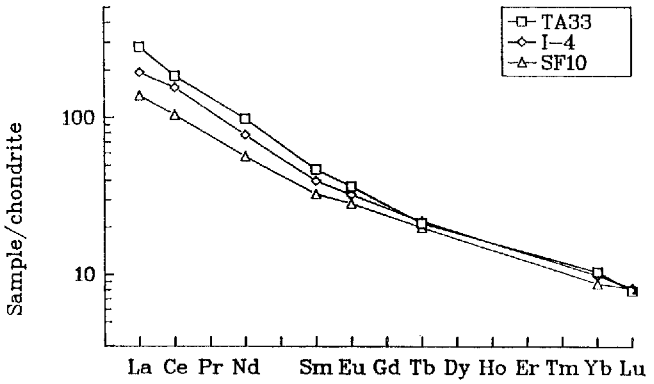

Fig. 8. Chondrite-normalised REE patterns for three hawaiite samples

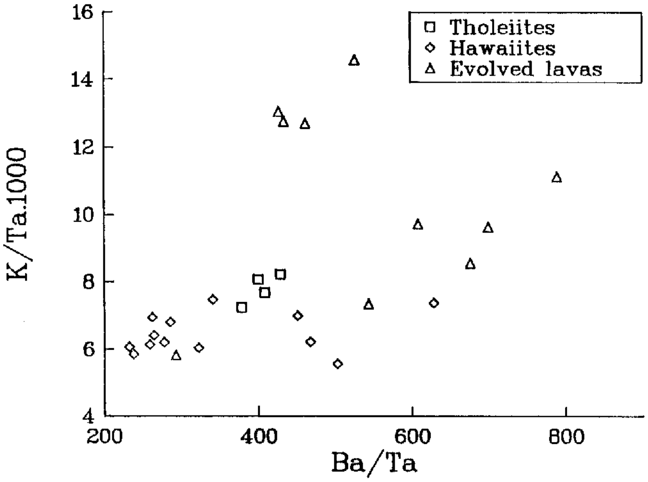

Fig. 9. K/Ta.1000 vs Ba/Ta. Note overall weak positive correlation, ascribed to crustal contamination of hawaiite to produce evolved compositions. There is however no correlation among the hawaiites

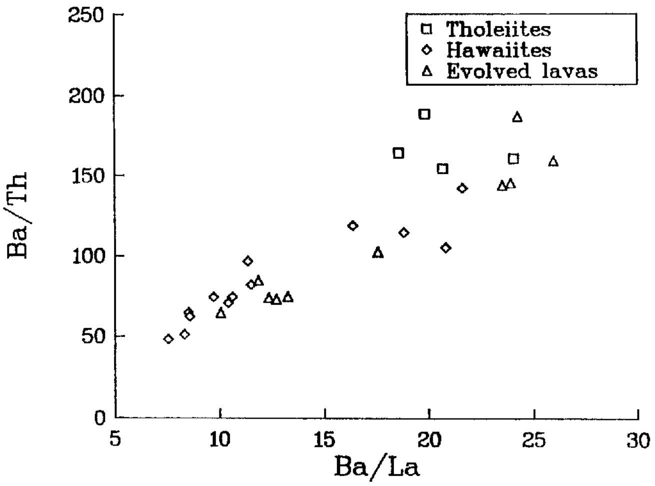

Fig. 10. Ba/Ta vs Ba/La

Highly alkaline oceanic magmas are often K and Rb depleted with respect to Th and (Nb, Ta); if these are the most representative of the asthenospheric OIB source end-member, which is generally supposed to extend beneath the continents, then K, Rb-depleted mantle could exist beneath the western United States. A residual phase with a strong affinity for Nb and Ta is then required. While the possibility exists that such a phase could be indirectly involved in the genesis of arc magmas and be responsible for their characteristic Nb, Ta depletions, it is generally believed that typical arc chemical signatures are imposed on sub-arc mantle by fluxing of a component from the descending slab, and are not generated during partial melting. If such a phase exists at all, it must therefore reside not in the mantle wedge beneath volcanic arcs but in the slab, from which a LILE-rich, Nb, Ta-poor component is added to the wedge. In any case, the three most commonly invoked culprit phases, sphene, perovskite and rutile, cannot be residual in the mantle source during CdR hawaiite genesis. Sphene strongly sequesters not only Nb and Ta but also the middle REE; hawaiite REE patterns do not display any marked upward concavity (Fig. 8). Perovskite is unstable in opx-bearing mantle assemblages (Arculus and Powell 1986), while mantle-derived melts are strongly undersaturated in rutile (Ryerson and Watson 1987).

Alternatively, given the long history of subduction off of western North America prior to the mid-Tertiary, the hawaiite source mantle may be essentially arc-like (i.e. enriched in Ba, Rb, Th, K and LREE relative to Nb and Ta), with K retained during partial melting in a residual phase such as phlogopite. However some hawaiites have high Ba/Rb ratios (Fig. 4), and Ba generally partitions more strongly into potassic phases than does Rb. More significantly, there is no correlation between K/Ta and Ba/Ta among the hawaiites (Fig. 9), such as would be expected were a phase which sequesters Ba as efficiently as phlogopite (Irving and Frey 1984) responsible for holding back K from escaping melts.

We conclude that the distinctive trace element characteristics of the hawaiites are inherited from their mantle source, and were not imparted during melting.

Mantle metasomatism. If a component has been added to the local mantle to produce the hawaiite signature, it must have been rich in Th and LREE, and poor in K, Rb, Nb and Ta. This is unlikely to be the case for any silicate or aqueous fluid capable of moving through the mantle. Recent experimental work by Meen et al. (1989) has shown that CO_2-dominated fluids do not accomodate significant amounts of REE under mantle conditions; furthermore, CO_2-rich fluids are probably incapable of infiltrating grain boundaries in mantle assemblages (Watson and Brenan 1987; Brenan and Watson 1988). Carbonatite liquids however may have some of the required physical and chemical properties, and may be significant mantle metasomatising agents (Green and Wallace 1988). The principal objection to invoking carbonate metasomatism of the hawaiite source is that – by analogy with carbonatites emplaced in or on the Earth's crust – the metasomatising agent should be very rich in Ba, yet the hawaiites are somewhat Ba-depleted; they typically have lower Ba/Th and Ba/La than CdR tholeiites (Fig. 10).

Removal of a component prior to magmagenesis. As a final possibility, we shall consider selective removal of incompatible elements from the mantle source prior to genesis of hawaiite (or unseen magmas parental to the hawaiites). Nb and Ta are unlikely to be mobilised by any process save partial melting, which will not serve to separate them significantly from La (see arguments above regarding possible residual phases). LILE are generally more mobile, and thus removal of K and Rb from a subduction-related mantle – the presence of which is tectonically feasible – already depleted in HFSE is more probable. The most obvious way of removing hygromagmatophile elements from a volume of mantle is through a very small degree of partial melting. Prior melt extraction is unlikely to produce marked separation of K and Rb from Th, however. Alternatively, volatile loss may exert some control on K and Rb within the mantle source (cf. Saunders et al. 1987); this is most likely to cause fractionation among LILE if accompanied by a phase change, resulting in selective loss of cations which cannot be accomodated in the new assemblage. The phase change from phlogopite to amphibole at around 20 kbar is of relevance here. Olafsson and Eggler (1983) have shown that the stable hydrous phase in mantle peridotite co-existing with small amounts of H_2O and CO_2 is phlogopite at pressures >22 kbar, and amphibole below this pressure. LREE exhibit greater affinity for amphibole than phlogopite, Th behaves in a similar fashion toward both phases, while amphibole tends to reject K, Ba and Rb compared to phlogopite (Wörner et al. 1983; Irving and Frey 1984). Therefore, conversion of phlogopite to amphibole in the mantle may act to expel K, Ba and Rb, but not Th or LREE, into a coexisting fluid or melt. Although K and Rb exhibit the greatest depletions (Fig. 4), Ba is also significantly depleted relative to Th and La in the hawaiites compared to the tholeiites (Fig. 10).

The phlogopite-amphibole conversion followed by partial melting is plausible in an extensional environment where lithospheric attenuation and mantle shallowing is actively taking place. Prior to extension, sub-CdR mantle may have been hydrated by slab-derived fluids yielding a phlogopite-bearing assemblage during early Tertiary shallow-angle subduction beneath the western United States to produce "normal" arc-source mantle wedge material. Subsequent lithospheric thinning during the mid-Tertiary allowed this material to rise to depths <70 km, and to lose K, Rb and Ba due to the phlogopite-amphibole conversion. Further extension caused melting and magma production from the selectively-depleted source. The geochemistry of the CdR hawaiites is thus a direct product of their specific tectonic environment, in a region with a long history of subduction, which has recently undergone extension.

CdR magmatism in a regional context

Leat et al. (1988, 1989, 1990) have described the Miocene to Recent mafic volcanism of the Colorado section of the RGR, where they note the presence of magmas representing three distinct mantle endmember components designated Groups 1–3. Group 1 magmas are derived from OIB-source mantle, Group 2 are potassic, high La/Ta magmas thought to represent partial melts of lithospheric mantle, while Group 3 are similar to our hawaiites, although including some lavas that are depleted in Th as well as Nb, Ta, K, Rb and Ba with respect to LREE. Leat et al. (1988) speculate that their Group 3 magmas are derived from a modified subduction-related source which lost those elements via fluid fluxing prior to Group 3 magma genesis. During the Miocene, the NW Colorado Group 3 mantle source behaved first as asthenosphere, but then as lithosphere after the introduction of OIB-source asthenosphere beneath the region at about 10 Ma (Leat et al. 1988, 1989). This can be rationalised if the group 3 source is located in a sluggish thermal boundary layer between asthenosphere and lithosphere (Parsons and McKenzie 1978; McKenzie and Bickle 1988).

As an elaboration of this model, we suggest that the Group 3 source was emplaced beneath the northern RGR during the phase of active subduction prior to 30 Ma, as phlogopite-bearing arc-source-like mantle. It may have become incorporated into the thermal boundary layer, or perhaps been accreted to the base of the lithosphere. The latter is more likely if hydrous phases melt quantitatively close to the solidus under the prevailing conditions, as found by Olafsson and Eggler (1983), although McKenzie (1985, 1987) has stressed that silicate liquids are only retained in deforming mantle at extremely low melt fractions, which may allow hydrous phases to persist in a thermal boundary layer. During the early extensional phase, lithospheric thinning may have permitted some of this material to rise through the phlogopite/amphibole boundary and lose K, Ba and Rb. Subsequent melting tapped this source (beginning at 24 Ma in Colorado), which later became underlain by OIB-source asthenosphere. Unlike NW Colorado, there is – as far as we know at present – no ancient lithospheric mantle component in any volcanic rocks associated with the Española Basin. Perhaps it was removed from the immediate rift area in the early phase of wide extension; if so, thermal boundary layer mantle may have been permitted to rise into the sub-rift region.

The CdR hawaiite/NW Colorado Group 3 mantle source is not apparent in the Pliocene Servilleta lavas of the Taos Plateau, for which Dungan et al. (1986) have convincingly argued that MORB-like mantle-derived tholeiitic magmas were contaminated to varying degrees by (mainly lower) continental crust. The magmatic volume of the Taos Plateau volcanic field (>250 km^3) is much greater than either the CdR or the volume of magmas erupted in NW Colorado following the first appearance of an ocean-like asthenospheric component. If these differences are due to variable extents of mantle melting beneath different parts of the RGR, with the ocean-like asthenosphere source predominating at higher degrees of melting, then the CdR hawaiite/NW Colorado Group 3 source may be volumetrically minor, consistent with our suggestion of its production at shallow mantle levels as a consequence of regional extension.

Conclusions

(1) Pliocene lavas of the NW Cerros del Rio volcanic field comprise a range of compositions of weakly alkaline affinity, together with late-erupted tholeiites. The entire field has a volume of ~ 60 km^3.
(2) $\delta^{18}O$ values of quartz grains in the lavas of alkaline affinity demonstrate that they are crustally-derived xenocrysts.
(3) Tholeiitic CdR lavas have been contaminated by upper, and probably lower, continental crust. When the effects of contamination are stripped off, they resemble E-MORB and OIB tholeiites. They are derived from an asthenospheric mantle source similar to that of most oceanic magmas.
(4) CdR hawaiites are derived from a mantle source depleted in K, Rb, Ba, Nb and Ta (but not Th) relative to LREE. These magmas experienced contamination by low-$^{87}Sr/^{86}Sr$, low-$^{143}Nd/^{144}Nd$ continental crust, which yielded intermediate lavas with up to 63% SiO_2 at the highest degree of contamination.
(5) A few CdR lavas represent mixing between hawaiite-source and and tholeiite-source mantle components.
(6) The hawaiite source component also appears in the RGR-related lavas of NW Colorado, and therefore is a regional mantle type. It most likely represents subduction-related mantle modified by subsequent lithospheric extension. Removal of K, Rb and Ba from initial arc-source-like mantle may have been achieved through uprise of a phlogopite-bearing mantle thermal boundary layer to a depth of <70 km, with consequent conversion of phlogopite to an amphibole phase. If so, similar trace element patterns should be apparent in the magmas of other extensional regions with a previous history of subduction.

Appendix

Analytical methods

Major elements and Rb, Sr, Ba, Cu, Ni, Zn, Cr, V, Y, Zr and Nb in samples with numbers beginning 6- were determined by XRF at the New Mexico Bureau of Mines and Mineral Resources using standard techniques. Major elements and Be, Ba, Sr, Ni, Cu, Co, Zn, Cr and V were determined on remaining samples by ICP at Imperial College (University of London) using a borate flux-fusion in nitric acid solution. In addition to standards, one sample (TA33) was used as a comparability monitor; no significant differences in reported concentrations for elements determined by both techniques were found. REE, Sc, Hf, Th, U and Ta were determined by INAA at the Imperial College Reactor Centre and at Durham University using a high-purity Ge detector. Full details of INAA analytical procedures are given by Leat et al. (1990).

Oxygen and hydrogen isotopes were determined in the stable isotope laboratory at Southern Methodist University. Whole-rock and hand-picked mineral separates for oxygen isotope analysis were prepared in a fluorination extraction line similar to that described by Clayton and Mayeda (1963), and isotope ratios determined on a Finnegan MAT-251 mass spectrometer. H_2O was extracted in an apparatus similar to that of Friedman and Smith (1958), and reduced to H_2 in a uranium furnace prior to D/H ratio determination on the Finnegan MAT-251. NBS-28 quartz has a $\delta^{18}O$ (SMOW) value of +5.66‰, and NBS-70 biotite a δD(SMOW) value of − 60‰, in the SMU laboratory. Analytical precision is ± 0.15‰ for $\delta^{18}O$ and ± 2‰ for δD.

Sr and Nd isotope ratios on CdR lavas were measured on a VG354 mass spectrometer at McMaster University, and for the two basement granite samples on a Finnegan MAT 251 solid-source mass spectrometer at the University of Texas at Dallas. Strontium ratios are normalised to $^{86}Sr/^{88}Sr = 0.1194$ and neodymium ratios to $^{146}Nd/^{144}Nd = 0.7219$. 2σ precision at McMaster is typically ± 0.000015 for $^{87}Sr/^{86}Sr$, and ± 0.000013 for $^{143}Nd/^{144}Nd$, and at Dallas is better than ± 0.000018 and ± 0.000015 respectively.

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