

EXPANDED DISCUSSION OF METHODS AND RESULTS

An expanded version of Table 1, showing additional calculations and results, is shown in Table 1 Supp. Table 2 Supp. shows isotopic data compiled from GEOROC and other sources. Table 3 Supp. shows parental liquid compositions, derived from methods discussed in Putirka et al. (2007), and used here to estimate T^{ol-liq} and T_p .

Calculation Strategy and Selection of Input Parameters

Estimates of mantle potential temperature, T_p can be derived from olivine-liquid equilibration temperatures (T^{ol-liq}) provided we know: 1) the FeO content of mantle-derived liquids that are “parental” (FeO^p) to erupted lavas, 2) the maximum forsterite (Fo_{max}) content of mantle olivines with which such parental liquids equilibrated, 3) the oxygen fugacity ($f\text{O}_2$) conditions during melting, and 4) the Fe-Mg exchange coefficient between olivine and liquid, $K_D(\text{Fe-Mg})^{ol-liq} = (X_{\text{Mg}}^{\text{liq}} X_{\text{Fe}}^{\text{ol}}) / (X_{\text{Mg}}^{\text{ol}} X_{\text{Fe}}^{\text{liq}})$ (where X_i^j is the cation fraction of element i in phase j).

Estimating FeO^p

The quantity FeO^p can be derived from plots of MgO v. FeO, when whole rocks or glasses lie on an “olivine control” line. The ocean islands evaluated by this study are limited to those where available data indicate that olivine control was achieved. Olivine control means that olivine addition or subtraction controls whole rock and liquid compositions (Powers, 1955). This occurs for primitive, parental liquids because parental liquids have high MgO, and precipitate only olivine when delivered to shallow depths. It has long been recognized (e.g., Langmuir and Hanson, 1980) that because Fe is partitioned nearly equally between olivine and co-existing liquid, trends of olivine fractionation/addition tend to be flat (note that the liquid lines of descent in Fig. 1 for

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Siqueiros and Hawaii do not provide a mechanisms to generate high FeO, high MgO liquids, from low FeO, high MgO parental liquids). FeO along such trends thus reflects the FeO content of parental, mantle-derived liquids (Langmuir et al., 1992; Putirka, 2005). The 28 ocean islands of this study were selected due to availability of high MgO compositions; nearly all ocean islands examined reach olivine control at >10% MgO (Fig. 1). Jan Mayen and Galapagos are exceptions. Most lavas in these islands trend to very low FeO at high MgO, but both regions are also near spreading centers, and so may be affected by mixing with ambient mantle. For example, Fernandina is the only Galapagos island to yield very high $^3\text{He}/^4\text{He}$, but published data from this island yield a mix of low and high FeO lavas, with no clear FeO-MgO trend, and so T_p at Fernandina cannot be reliably estimated.

Other input parameters

As for the remaining quantities needed for temperature estimation, they are assumed equal for ocean islands and MORs: $f\text{O}_2$ is QMF-1 (Bezos and Humler, 2005), and $K_D(\text{Fe-Mg})^{\text{ol-liq}}=0.31$ (Herzberg and O'Hara, 2002; Putirka, 2005; Toplis, 2005), and $\text{Fo}_{\text{max}} = 91.5$ (Albarede, 1992; Kelemen et al., 1997; Asimow and Longhi, 2004; Putirka et al., 2007). Reported Fe analyses are re-calculated to Fe^{2+}O -total (FeO) (Bezos and Humler, 2005; Kress and Carmichael, 1988). A database of 3,070 analyses of ocean island olivine compositions confirms that globally, Fo_{max} for ocean islands lies in the range 91.5 ± 0.5 . It cannot be precluded that $f\text{O}_2$ conditions are different for ocean islands and MOR, or vary from one ocean island to another. However, $f\text{O}_2$ studies at Hawaii (Rhodes and Vollinger, 2005) and Iceland (Heister et al., 2004) indicate broadly similar conditions as at MORs, that if adopted for all ocean islands, would not bias results by

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more than 10°C. Water also affects T^{ol-liq} , and it has been suggested that hot spots are not hot, but wet (Bonatti, 1990; Green et al., 2001). Since Dixon et al. (2002) have established that ocean island source regions indeed contain greater water contents (0.075 wt. % H₂O, compared to 0.01% H₂O in the depleted MORB source), their source differences and bulk distribution coefficient for H₂O, are adopted for calculating T^{ol-liq} (Table 1). (To calculate H₂O, melt fraction, F , is calculated from equations A1 and A2 of Putirka et al. (2007); see Calculation details, below).

Some anti-plume advocates (e.g., Falloon et al., 2007) suggest that only glass compositions (which may have lower MgO and FeO compared to whole rocks) should be considered for T estimation. But “liquids” provide a highly homogenized account of mantle melting processes (e.g., Langmuir et al., 1992) and do not preserve the most primitive melt compositions produced by partial melting. For example, because olivine-rich magmas are denser than their evolved counterparts, it has long been recognized that the MgO contents of glass compositions, and the sum of all erupted products (crystals and liquids), necessarily yield minimum estimates for parental MgO^{liq} (e.g., Wright, 1973; Clague et al., 1991). Olivine and whole rock compositions from the recent Hawaiian Scientific Drilling Project validate this view, that Hawaiian parental liquids have picrite compositions (Putirka et al., 2007). Many studies (e.g., Langmuir and Hanson, 1980; Clague et al., 1995; Rhodes and Vollinger, 2004; Asimow and Longhi, 2004; Herzberg et al., 2007), even those by anti-plume advocates (Falloon et al., 2007), also quantitatively demonstrate that FeO contents along olivine control lines cannot be generated by fractionation of liquids with substantially lower FeO contents—they must derive from parental liquids that have similar, or higher FeO. These conclusions are unaffected by the

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recognition that some high-SiO₂ liquids may derive by reaction with harzburgite (Wagner and Grove, 1998; Stolper et al., 2004), as such wall-rock reaction decreases FeO contents (Fig. 1).

Calculations and Results

T^{ol-liq} is estimated (Fig. 2) using thermometers from Putirka et al. (2007), which account for variations in P and H₂O, and anhydrous models from Beattie (1993) (modified to apply at very high P ; Herzberg and O'Hara, 2002). Though calibrated with different experimental data, these models exhibit remarkably similar levels of accuracy and precision (± 40 -50°C) (Putirka et al., 2007) and yield similar T^{ol-liq} estimates (Table 1). All three thermometers account for variations in SiO₂^P and (Na₂O+K₂O)^P (as in FeO^P, these are the SiO₂ and Na₂O+K₂O for primitive liquids). Estimates of SiO₂^P and (Na₂O+K₂O)^P at each ocean island are obtained by averaging lavas that have 10wt.%<MgO<15wt.%, and adding olivine until such liquids achieve Fe-Mg exchange equilibrium with Fo91.5 olivine; F is calculated using equation A1 (Table 1) from Putirka et al. (2007)) from these calculated liquid compositions; F averages 11% for ocean island primitive liquids. F can also be calculated using equation A2 in Putirka et al. (2007), which is based on experiments performed using enriched peridotite compositions from Pickering-Witter and Johnston (2000); using equation A2, the average for F among ocean islands is 20%. Table 1 Supp shows calculations for T_p and T_{ex} a moderately enriched mantle source, $T_p(En)$ and $T_{ex}(En)$, based on the mean F calculated from equations A1 and A2 (average ocean island $F = 16\%$), for mantle potential temperature,. Olivine compositions are selected by projecting FeO-MgO trend lines to their intersection with an

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olivine composition trend line; if olivine is added along liquid lines of descent (the olivine fractionation trend in Fig. 1), nearly identical parental liquid compositions are calculated. Melt fraction, F , is calculated using the composition- and pressure-dependent equation A1 of Putirka et al. (2007), and this F is used to calculate H_2O^p (using Dixon et al.'s (2002) estimates for H_2O in the ocean island and MOR source regions). T^{ol-liq} values are converted to T_p using F , a mantle adiabat, dT/dP , of 12.1 °C/GPa, and a thermal correction for the heat of fusion ($\Delta T^{corr} = F\Delta H_{fusion}/Cp$; $\Delta H_{fusion} = 128.3$ kJ/mole; $Cp = 192$ J/mole•K; Putirka, 2005); $T_p = T^{ol-liq} + \Delta T^{corr} - P(dT/dP)$. Results are in Table 1.

For ambient mantle, temperatures are calculated using mean MORB FeOt, which happens to be nearly identical to mean FeOt for primitive lavas from Siqueiros (Perfit et al., 1996) (Fig. 1). Putirka et al. (2007) have shown that, globally, MORB vary over a very narrow temperature interval, with 1σ as low as ± 35 -40°C. This result is not inconsistent with temperature estimates by Langmuir et al. (1992), who did not calculate a standard deviation or an average MORB temperature, but rather a range of MORB temperatures, which included Iceland (see Putirka et al., 2007).

Since equilibration of parental liquids with Fo91.5 olivine does not occur at Earth's surface, but most likely below the base of the crust (Putirka, 1997; Klügel and Klein, 2006), all calculations are performed at 1 GPa. However, although this assumed value for P affects estimates of T^{ol-liq} , it has only a minimal impact on T_p . For example, by calculating T_p (from T^{ol-liq}) at MORs in 1GPa intervals, from 1 atm to 5 GPa, the average T_p and standard deviation are: $T_p^{MOR} = 1414 \pm 24^\circ\text{C}$. This standard deviation is less than thermometer error, and results from the fact that higher T^{ol-liq} estimates (at higher P) are

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offset by a greater degree of cooling due to adiabatic upwelling, and a smaller positive correction to T_p related to the heat of fusion (Putirka et al., 2007).

Interestingly, the anhydrous thermometer of Beattie (1993) yields intermediate T^{ol-liq} estimates (Table 1) compared to H₂O-dependent models (Putirka et al., 2007) (Table 1 Supp), indicating that H₂O contents at ocean islands might be low enough that the effects of inter-ocean island variations in H₂O are likely subsumed by model error. If all oceanic volcanic sources are assumed to be anhydrous, mean T_{ex} is between 195-205°C, and the temperature for ambient mantle, $T_p^{MORs} = 1411^\circ\text{C}$. The excess temperatures of Table 1 and Table 1 Supp, are also similar to those derived by an entirely different method, involving partial melting (Herzberg et al., 2007).

Differences between these T estimates, and Putirka et al. (2007)

The estimates for T_p^{MORs} , T_p^{Samoa} and T_p^{Hawaii} are less than the estimates of Putirka et al. (2007). Most fall within or near the 1σ error of 80°C for error on T_p (Putirka et al., 2007). The present estimate for MORB is lower by 30-40°C because in Putirka et al. (2007), they considered all MORB with 7 wt. % < FeOt < 9.5 wt. %, to calculate mean FeOt for MORB, regardless of MgO. However, MORB with >9.5% MgO have slightly lower FeOt (8.18%) because by using high MgO samples, those MORB that have been affected by plagioclase fractionation are completely excluded (plagioclase fractionation increases FeOt). The Putirka et al. (2007) estimate of mean MORB FeOt (8.8%) is therefore an upper limit to mean FeOt for MORB, as is their estimate of T_p^{MORs} . Notably, the average obtained using high MgO MORB from PETDB is remarkably similar to that obtained from picrites from Siquieros (8.14%; Perfit et al., 1996). Temperatures are also slightly lower because here, values for T^{ol-liq} are calculated using an average of models

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(2) and (4) from Putirka et al. (2007) and Beattie (1993). Tests of these models in Putirka et al. (2007) show that each of these models yield similar mean errors for experimental data, even though they are calibrated using different variables and equations. The current judgment is that it is probably best to average results from all three expressions. Finally, for Samoa, FeO_t contents for data published in GEOROC trend to somewhat lower values than considered in Putirka et al. (2007), and use of this larger database results in lower temperatures.

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Table DR1. Expanded version of Table 1 in paper.

	H ₂ O ^{pl}	F	(1) T_{ol}^{ol-} liq(°C)	(2) T_{ol}^{ol-} liq(°C)	(3) T_{ol}^{ol-} liq(°C)	Corr T to Tp	Mean T_{ol}^{ol-} liq(°C)	Mean T_p (°C)	T_{ex} (°C)	H ₂ O ^p to get $T_{ex}=0$	T_p (En)	T_{ex} (En)
Ambient Mantle												
MOR	0.11	8.0	1353	1368	1344	41.2	1355	1396	-	-	1410	0
Ocean Islands												
Azores	1.07	9.3	1466	1511	1437	50.0	1471	1521	124	6.00	1554	144
Canary	1.37	9.9	1498	1553	1470	53.7	1507	1561	164	7.51	1597	188
Cape Verde	1.78	4.7	1497	1526	1452	19.2	1492	1511	114	6.09	1558	148
Caroline Is.	1.44	7.5	1504	1553	1470	37.8	1509	1547	150	7.11	1587	178
Crozet	0.63	10.9	1472	1520	1450	60.8	1481	1542	145	6.77	1575	165
Fernando de	2.17	7.4	1530	1586	1492	37.0	1536	1573	176	8.06	1617	207
Galapagos	1.06	6.1	1385	1386	1354	28.9	1375	1404	7	1.34	1429	19
Galapagos (VE)	1.06	6.1	1458	1497	1431	65.0	1462	1527	130	8.34	1556	147
Gough	1.52	12.5	1470	1529	1442	71.1	1481	1552	155	7.03	1581	171
Grande Comore	1.49	9.8	1496	1547	1466	53.4	1503	1557	160	7.32	1593	183
Hawaii	0.65	25.6	1512	1576	1497	158.7	1528	1687	290	12.31	1712	302
Heard	1.68	12.9	1487	1558	1460	73.7	1502	1576	179	7.99	1607	197
Iceland	0.34	21.0	1447	1483	1433	128.3	1454	1583	186	7.96	1599	189
Jan Mayen	1.20	9.0	1418	1456	1395	48.2	1423	1471	74	4.02	1497	87
Juan	1.91	11.5	1507	1569	1476	64.6	1517	1582	185	8.20	1617	207
Kerguelen	0.39	18.4	1487	1531	1468	110.6	1496	1606	209	9.15	1636	226
Louisville	1.20	14.4	1503	1557	1480	83.8	1513	1597	200	8.81	1631	221
Madeira	1.24	10.2	1496	1545	1467	55.9	1503	1558	161	7.40	1595	185
Marquesas	1.18	11.6	1489	1545	1463	65.2	1499	1564	167	7.59	1598	188
Ninetyeast	0.47	15.1	1469	1508	1448	88.4	1475	1563	166	7.48	1593	183
Pitcairn	1.17	13.0	1498	1567	1468	74.5	1511	1586	189	8.33	1618	208
Reunion	0.52	13.5	1485	1537	1463	77.9	1495	1573	176	7.91	1606	196
Ross/Erebus	0.99	6.6	1490	1539	1448	31.9	1493	1525	128	6.27	1563	153
Samoa	1.40	16.5	1507	1575	1486	97.9	1522	1620	223	9.67	1652	242
Selvagen Is.	2.00	7.3	1482	1522	1444	36.3	1483	1519	122	5.99	1555	145
St. Helena	1.04	11.7	1486	1538	1462	65.8	1495	1561	164	7.48	1595	185
Tahiti/Society	1.35	12.4	1499	1562	1474	70.5	1511	1582	185	8.27	1617	207
Trindade	2.33	9.3	1572	1654	1537	50.0	1588	1639	242	10.16	1684	274
Tristan de	1.75	11.2	1500	1559	1473	62.7	1511	1573	176	7.95	1609	199

F is percent of partial melting determined from calculated primitive liquid compositions and equation A1 from Putirka et al. (2007);
H₂O^p is wt. % water in the primitive liquid, which is based on *F*, and the bulk distribution and source concentration values in Dixon et

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al. (2002). All temperatures are in °C. (1) uses $T = f(D_{Mg}, P, X_i)$ calibrated from low P , anhydrous experiments (Beattie, 1993), using empirical corrections for high P (Herzberg and O'Hara, 2002); (2) uses $D_{Mg} = f(T, P, X_i)$, calibrated from hydrous and anhydrous experiments at low and high P (Putirka et al., 2007, Eq. 2); (3) uses $T = f(D_{Mg}, P, X_i)$, calibrated as in (2) (Putirka et al., 2007; Eq. 4). All calculations use parental magma compositions calculated as noted in text, and $Fo_{max}=91.5$, except for Hawaii and Iceland, where Fo_{max} values are 91.3 and 92.0 respectively. All excess temperatures (T_{ex}) for ocean islands are relative to T_p^{MOR} (Siqueiros), which represents ambient mantle (Fig. 1). T^{corr} (°C) is the thermal correction related to the heat of fusion and adiabatic upwelling ($T_p = T^{ol-liq} + T^{corr}$); $T_{ex}(^{\circ}C) = T_p^{Oceans.} - T_p^{MOR}$; H_2O^p to get ($T_{ex}=0$) is wt. % water needed in the primitive liquid to explain differences in T^{ol-liq} without an excess temperature. $T_p(En)$ and $T_{ex}(En)$ are the mantle potential and excess temperatures when F is calculated using the mean of estimates for F from equations A1 and A2 in Putirka et al. (2007) (a moderately enriched mantle). Because F is higher for enriched mantle compositions, $T_p(En)$ and $T_{ex}(En)$ are higher, for two reasons: high F means a greater upward correction for the heat of fusion, and for water dependent models, high means that H_2O is lower.

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Table DR2. Ocean Island isotope data

Ocean Islands ¹	³ He/ ⁴ He	²⁰⁸ Pb/ ²⁰⁴ Pb	²⁰⁷ Pb/ ²⁰⁴ Pb	²⁰⁶ Pb/ ²⁰⁴ Pb	¹⁴³ Nd/ ¹⁴⁴ Nd	⁸⁷ Sr/ ⁸⁶ Sr
	max					
Azores	11.5	39.24	15.62	19.47	0.512865	0.7038
Canary		39.24	15.58	19.45	0.512903	0.7032
Cape Verde	13.8	39.04	15.58	19.42	0.512881	0.7032
Caroline Is.					0.512979	0.7032
Crozet		39.31	15.61	18.90	0.512854	0.7040
Fernando de Noronha		39.31	15.63	19.40	0.512826	0.7055
Galapagos (VE)	9.9	38.95	15.58	19.20	0.512958	0.7042
Gough	7.02	38.84	15.60	18.42	0.512574	0.7047
Grande Comore	7.08	39.41	15.57	19.08	0.512751	0.7037
Hawaii	24.8	38.01	15.47	18.36	0.512990	0.7035
Heard	18.3	38.54	15.56	18.07	0.512592	0.7055
Iceland	34.3	37.87	15.44	18.23	0.513086	0.7031
Jan Mayen						
Juan Fernandez	16.6				0.512870	0.7035
Kerguelen	17.5	37.91	15.47	17.61	0.512687	0.7054
Louisville						0.7034
Madeira	6.7	38.75	15.53	19.09	0.513082	0.7028
Marquesas		39.04	15.58	19.25	0.512837	0.7041
Ninetyeast Ridge		38.94	15.58	18.66	0.512870	0.7048
Pitcairn	10.6	39.08	15.57	19.49	0.512932	0.7033
Reunion	13.14	38.97	15.58	18.84	0.512838	0.7042
Ross/Erebus		39.52	15.64	19.89		0.7031
Samoa	17.97	39.52	15.60	19.30	0.512768	0.7051
Selvagen Is.		39.23	15.57	19.43	0.512909	0.7031
St. Helena	5.9	37.65	15.75	20.63	0.512913	0.7029
Tahiti/Society		38.85	15.59	19.10	0.512813	0.7045
Trindade		38.78	15.55	18.99	0.512791	0.7037
Tristan de Cunha	6.3	39.05	15.54	18.63	0.512585	0.7047

¹Data are from GEOROC and Class and Golstein, 1997; Class et al., 2005; Class et al., 1998; Claude-Ivanaj et al., 1998; Doucet et al., 2006; Geist et al., 2003; Honda and Woodhead (2005) Hopp and Tieloff (2005); Kurz et al., 2004; MacPherson et al., 2005; Workman et al., 2004.

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Table DR3. Parental liquid compositions (oxides in weight %), calculated using methods in Putirka et al. (2007)

Location	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅
MORB-Siqueiros	48.5	0.9	15.9	8.0	0.1	13.2	11.1	2.3	0.0	0
Azores	45.1	2.1	10.9	11.1	0.2	18.2	9.2	2.1	0.9	0.3
Canary	43.7	2.5	9.5	11.9	0.1	19.7	9.2	2.1	0.8	0.5
Cape Verde	41.5	2.7	9.5	11.8	0.2	19.5	11.2	2.1	0.8	0.6
Caroline Is.	42.5	2.6	9.4	11.9	0.1	19.8	9.9	2.3	0.9	0.6
Crozet	45.1	2.1	9.5	11.5	0.1	18.9	10.2	1.7	0.7	0.3
Fernando de Nohrona	41.4	2.6	8.7	12.3	0.1	20.7	9.9	2.5	1.0	0.7
Galapagos	46.9	0.9	14.8	9.3	0.1	14.7	11.0	2.1	0.0	0.1
Galapagos (VE)	46.4	2.0	11.9	11.0	0.1	17.9	8.2	2.2	0.4	0.2
Gough	46.7	2.3	10.3	11.1	0.1	18.1	7.5	2.2	1.3	0.3
Grande Comore	44.0	1.6	10.5	11.8	0.2	19.4	9.1	2.3	0.8	0.3
Hawaii	47.2	1.6	9.0	11.6	0.2	21.4	7.3	1.4	0.2	-
Heard	45.9	2.8	8.9	11.5	0.1	19.0	7.6	2.0	1.6	0.5
Iceland	47.2	0.7	12.3	9.3	0.2	18.2	10.5	1.4	0.1	-
Jan Mayen	46.9	2.0	10.0	10.0	0.1	16.1	11.6	1.8	1.3	0.3
Juan Fernandez	44.5	2.1	10.2	11.9	0.1	19.7	7.5	2.6	0.9	0.4
Kerguelen	47.5	0.7	11.3	12.0	0.2	19.9	6.9	1.4	0.1	0.1
Louisville	45.0	2.0	9.9	12.2	0.1	20.4	7.8	1.8	0.4	0.3
Madeira	43.9	2.1	10.4	11.9	0.2	19.7	8.8	2.0	0.6	0.4
Marquesas	44.8	2.6	9.6	11.8	0.1	19.4	8.5	1.9	0.9	0.4
Ninetyeast Ridge	47.1	0.9	11.9	11.5	0.2	18.9	7.7	1.5	0.3	0.1
Pitcairn	45.6	2.4	10.5	11.7	0.1	19.4	6.4	1.9	1.7	0.4
Reunion	45.7	1.9	10.1	11.8	0.1	19.5	8.2	1.8	0.5	0.2
Ross/Erebus	43.4	2.7	10.7	11.3	0.2	18.5	8.5	3.0	1.1	0.6
Samoa	45.3	2.5	9.2	12.0	0.2	20.5	8.0	1.8	0.7	-
Selvagen Is.	44.0	2.0	10.6	11.2	0.2	18.3	9.4	2.7	0.8	0.8
St. Helena	44.9	2.2	9.8	11.8	0.1	19.4	9.0	1.8	0.7	0.3
Tahiti/Society	44.7	2.5	9.2	12.0	0.1	19.9	8.4	1.8	1.0	0.4
Trindade	40.3	2.8	7.7	13.0	0.1	22.4	8.7	2.8	1.3	0.7
Tristan	44.1	2.4	9.8	12.0	0.1	19.9	8.5	1.9	1.0	0.4