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Mantle melting as a function of water content beneath back-arc basins

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[1] Subduction zone magmas are characterized by high concentrations of H_2O, presumably derived from the subducted plate and ultimately responsible for melting at this tectonic setting. Previous studies of the role of water during mantle melting beneath back-arc basins found positive correlations between the H_2O concentration of the mantle (H_2O_o) and the extent of melting (F), in contrast to the negative correlations observed at mid-ocean ridges. Here we examine data compiled from six back-arc basins and three mid-ocean ridge regions. We use TiO_2 as a proxy for F, then use F to calculate H_2O_o from measured H_2O concentrations of submarine basalts. Back-arc basins record up to 0.5 wt % H_2O or more in their mantle sources and define positive, approximately linear correlations between H_2O_o and F that vary regionally in slope and intercept. Ridge-like mantle potential temperatures at back-arc basins, constrained from Na-Fe systematics (1350°–1500°C), correlate with variations in axial depth and wet melt productivity (~30–80% F/wt % H_2O_o). Water concentrations in back-arc mantle sources increase toward the trench, and back-arc spreading segments with the highest mean H_2O_o are at anomalously shallow water depths, consistent with increases in crustal thickness and total melt production resulting from high H_2O. These results contrast with those from ridges, which record low H_2O_o (<0.05 wt %) and broadly negative correlations between H_2O_o and F that result from purely passive melting and efficient melt focusing, where water and melt distribution are governed by the solid flow field. Back-arc basin spreading combines ridge-like adiabatic melting with nonadiabatic mantle melting paths that may be independent of the solid flow field and derive from the H_2O supply from the subducting plate. These factors combine significant quantitative and qualitative differences in the integrated influence of water on melting phenomena in back-arc basin and mid-ocean ridge settings.

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1. Introduction

[2] Water has long been understood to play a major role in the generation of subduction zone magmas. In essence, water lowers the mantle solidus [e.g., *Kushiro et al.*, 1968], which ultimately drives melting of the mantle wedge beneath arcs and back-arc basins due to a flux of water originating from the dehydrating, subducting slab. This has been the working paradigm for subduction zone magma genesis for more than 20 years, supported by the widespread observations that subduction zone lavas are vesicular and explosive, contain hydrous phenocrysts [e.g., *Gill*, 1981], follow hydrous liquid lines of descent [*Sisson and Grove*, 1993a], are enriched in fluid-mobile trace elements [e.g., *Morris et al.*, 1990], and carry trace element and isotopic signatures characteristic of subducting sediment and oceanic crust [e.g., *Tera et al.*, 1986; *Plank and Langmuir*, 1993; *Miller et al.*, 1994]. High H_2O concentrations in arc magmas have been inferred or estimated using innovative analytical and experimental techniques, [e.g., *Anderson*, 1979, 1982; *Sisson and Grove*, 1993a, 1993b; *Gaetani et al.*, 1994], and direct measurements of H_2O in submarine glasses and melt inclusions have supported these earlier estimates of dissolved H_2O in arc and back-arc lavas [e.g., *Muenow et al.*, 1980; *Aggrey et al.*, 1988; *Muenow et al.*, 1991; *Danyushevsky et al.*, 1993; *Sisson and Layne*, 1993; *Stolper and Newman*, 1994] (hereinafter referred to as S&N94). Notably, a significant increase in the number of direct measurements of volatile contents in arc and back-arc magmas in recent years [e.g., *Kamenetsky et al.*, 1997; *Roggensack et al.*, 1997; *Sisson and Bronto*, 1998; *Newman et al.*, 2000; *Fretzdorff et al.*,

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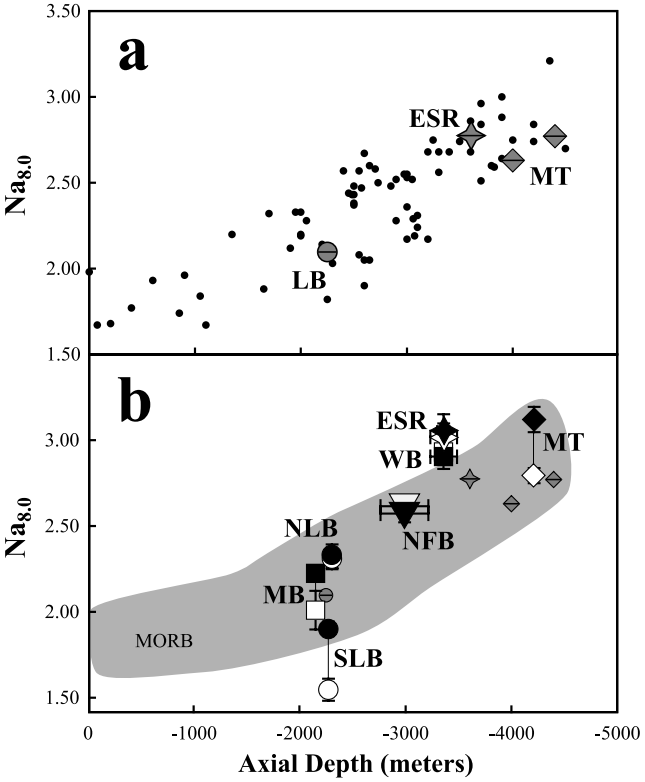

Figure 1. $Na_{8.0}$ versus axial depth in global MORB and back-arc basins. (a) MORB and back-arc basin regional averages from *Klein and Langmuir* [1987]. The black circles are MORB, and the barred shaded symbols are back arcs. Abbreviations are as follows: MT, Mariana trough; ESR, East Scotia ridge; LB, Lau basin. *Klein and Langmuir* [1987] reported two values for the Mariana trough at 17.39°N (4000 m depth) and at 18.01°N (4400 m depth). (b) Back-arc basin basalts from the compilation in this study superimposed on the data and trends of *Klein and Langmuir* [1987]. The white symbols are average back arcs (unfiltered for H_2O content (this study, Table 3)), the black symbols are average, dry back-arc basin basalts (H_2O < 0.5 wt % (this study, Table 3)), the barred shaded symbols are back arcs from *Klein and Langmuir* [1987] as in Figure 1a, and the shaded field is the MORB array from Figure 1a. Abbreviations are as follows: WB, Woodlark basin; NFB, North Fiji basin; NLB, northern Lau basin; SLB, southern Lau basin; MB, Manus basin; other abbreviations as in Figure 1a.

2002; *Kent et al.*, 2002; *Cervantes and Wallace*, 2003; *Sinton et al.*, 2003; *Walker et al.*, 2003] has set the stage for a quantitative investigation of water in natural subduction zone settings.

[3] Back-arc basins are natural places to begin such a study because they can be treated, in many ways, like mid-ocean ridges. The driest back-arc basin melts are compositionally equivalent to mid-ocean ridge melts and thus, like mid-ocean ridge magmas, can be interpreted as melts generated by varying extents of adiabatic decompression melting of ascending mantle. For example, the driest back-arc basin basalts overlap with MORBs on $Na_{8.0}$-axial depth plots (Figure 1 [see also *Klein and Langmuir*, 1987]) and plots of Na_2O versus FeO (Figure 2 [see also *Langmuir et al.*, 1992; *Taylor and Martinez*, 2003]). Both inter- and intrabasin systematics of back arcs were recently explored by *Taylor and Martinez* [2003], who compared major and trace element variations (e.g., $Na_{8.0}$, $Fe_{8.0}$, $Ti_{8.0}$, $H_2O_{(8.0)}$, Ba/La) of four back-arc basins to global MORB; their work emphasizes the hybrid nature of the back-arc basin melting process, identifying MORB-like geochemical systematics in relatively dry back-arc melts and showing how these systematics are perturbed in wetter samples by the addition of H_2O-rich material from the subducted slab. The melting process in back arcs is also not complicated, as it may be in arcs, by the effects of overlying crust, since crustal assimilation is a small concern and crustal thickness will not impact the melting column because it is a passive consequence of melting just as at ridges. Additionally, back arcs are less subject to the thermal influence of the cold subducting slab, and may reflect ambient upper mantle temperature variations as ridges do. Back-arc basins are therefore potentially ideal sites for studying the effects of variations in mantle temperature and water contents on mantle melting with fewer complicating factors than arcs.

[4] The quantitative relationship between water and mantle melting over the range of tectonic settings has been investigated in only a few localities, with contrasting results. S&N94 evaluated the importance of water in the formation of magmas at the Mariana trough back-arc spreading center

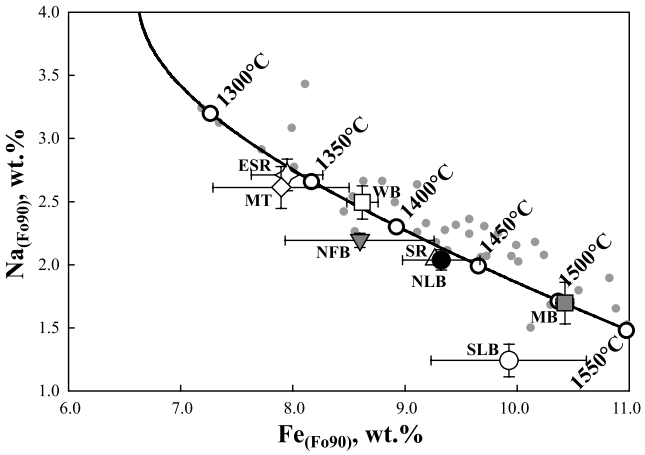

Figure 2. Regionally averaged $Na_{(Fo90)}$ versus $Fe_{(Fo90)}$ in MORB and back-arc basin basalts (Na_2O and FeO concentrations corrected to equilibrium with Fo_{90}; see section 2). The shaded points are MORB from *Langmuir et al.* [1992]. The black line connected by open circles shows the MORB model curve with potential temperatures from *Langmuir et al.* [1992]. Symbols are averages of the driest back-arc basin basalts from each region (<0.5 wt % H_2O; same as solid symbols in Figure 1b and Table 3); the open diamond is the Mariana trough (MT); the open star is the East Scotia ridge (ESR); the open square is the Woodlark basin (WB); the shaded inverted triangle is the North Fiji basin (NFB); the open triangle is the Sumisu rift (SR); the solid circle is the northern Lau basin (NLB); the open circle is the southern Lau basin (SLB); and the shaded square is the Manus basin (MB). The Sumisu rift and southern Lau basin contain no dry basalts, but average data for these regions are shown for reference.

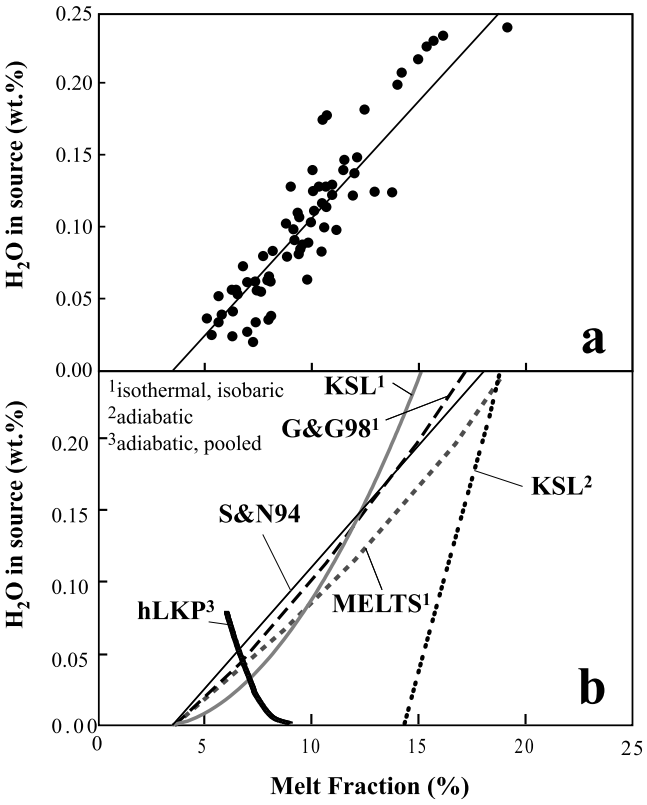

Figure 3. H_2O_o versus F from the literature. (a) Original figure from S&N94 showing their Mariana trough data and the linear best fit. (b) Models compiled from the literature for wet melting of peridotite. The thin, solid black line (S&N94) is the S&N94 Mariana trough best fit line from Figure 3a; the thick, short-dashed, shaded line (MELTS) is the isobaric, isothermal MELTS calculation from *Hirschmann et al.* [1999] at 1350°C and 1 GPa; the thin, long-dashed, black line (G&G98) is the fit to the isobaric, isothermal experimental results from G&G98 at 1350°C and 1.5 GPa; the triple-dotted black line (KSL^2) is an adiabatic model result of *Katz et al.* [2003] for T_p = 1250°C; the solid shaded line (KSL^1) is an isobaric, isothermal model result from *Katz et al.* [2003] at 1250°C and 1 GPa; the bold black line (hLKP) is an adiabatic model result from *Asimow and Langmuir* [2003] for T_p = 1350°C, pooled from a 2-D melting triangle.

using concentrations of volatile, major, and trace elements from submarine glasses. This work was the first to recognize correlations between trace elements and water contents of subduction-related lavas and suggested a positive, roughly linear relationship between the extent of mantle melting (i.e., melt fraction; F) and the H_2O concentration of the mantle (Figure 3a). The basis for pure "flux melting" models originates from this study. Correlations between H_2O and trace elements in volcanic arc melts [e.g., *Grove et al.*, 2002; *Cervantes and Wallace*, 2003; *Walker et al.*, 2003] provide supporting evidence for the importance of the flux melting mechanism beneath arcs, although nominally dry arc melts from Japan [*Tatsumi et al.*, 1983], the Cascades [*Bartels et al.*, 1991], and Java [*Sisson and Bronto*, 1998] implicate mantle decompression as another means of melt generation beneath arcs. In contrast to the positive correlation between H_2O and F described or inferred for back-arc basin and arc lavas, recent models of hot spot-influenced, "damp" mid-ocean ridges [*Detrick et al.*, 2002; *Asimow et al.*, 2004; *Cushman et al.*, 2004] suggest a negative correlation between H_2O and F (e.g., Figure 3b, hLKP curve). The differences in the relationships between water content and extent of melting in two such similar tectonic settings as back-arc basins and mid-ocean ridges present a dilemma that we address in this study.

[5] Theoretical, empirical, and experimental investigations of the role of water in mantle melting underscore the differences observed in natural settings. For example, experimental work [*Gaetani and Grove*, 1998] (hereinafter referred to as G&G98), thermodynamic modeling [*Hirschmann et al.*, 1999], and empirical parameterizations of experimental and thermodynamic data [*Katz et al.*, 2003] have shown that the positive melting trend observed in the Mariana trough is readily reproduced through isothermal, isobaric equilibration of fertile mantle with increasing H_2O content (Figure 3b, MELTS, G&G98, and KSL^1 curves), although such conditions do not translate easily into a physical melting process in the earth [S&N94; *Hirschmann et al.*, 1999; *Asimow and Langmuir*, 2003]. Parameters such as temperature and pressure certainly vary in natural systems, and the apparent similarities between these studies and the Mariana trough trend may indeed be only coincidental. On the other hand, *Asimow and Langmuir* [2003] used simple thermodynamic models to propose that adding water to decompression melting regimes at mid-ocean ridges will, seemingly paradoxically, increase total melt production but simultaneously decrease the mean extent of melting. This situation arises when pooling melts from a two-dimensional (2-D) triangular melting regime because low-F, wet melts originate from the large source volume at the triangle's base. Such a geometry can thus lead to a decrease in mean F with increasing H_2O in the mantle (Figure 3b, hLKP curve). Although this effect has been documented at hot spot-influenced mid-ocean ridges [*Detrick et al.*, 2002; *Asimow et al.*, 2004; *Cushman et al.*, 2004], it has not been seen in the high-H_2O lavas from back-arc basins. *Katz et al.* [2003] also developed an empirical parameterization of hydrous, adiabatic melting. The relationship between H_2O content and extent of melting in their hydrous, adiabatic melting model, however, contrasts with both the observed trend in the Mariana trough and with other efforts to model hydrous, adiabatic melting (Figure 3b, KSL^2 curve). Their model produces positive correlations between H_2O in the source and F, but the steepness of the trend suggests that H_2O addition does not lead to a significant increase in extent of melting, and thus that H_2O-fluxed melting contributes volumetrically very little to the final melt produced by adiabatic decompression in this model. For back-arc basins, therefore, we must presently choose between, on the one hand, an unlikely melting process (i.e., flux-induced, isothermal, isobaric melting) that produces trends similar to those observed in nature or, on the other hand, an apparently more realistic melting process (i.e., polybaric, hydrous, adiabatic melting) for which current models do not reproduce observed trends.

[6] In the work reported in this paper, we studied aspects of the differences between ridge and back-arc melting

systematics, focusing in particular on discriminating between the competing effects of H_2O, decompression, and mantle temperature on melt production. Using available measurements of the water contents of glasses from several back-arc basins, we first show that the positive H_2O-F trend of the Mariana trough found by S&N94 is a general phenomenon in back-arc basin settings. In detail, however, these trends differ in important ways from one back-arc region to another. We then use these differences and their correlations with other geochemical and geophysical parameters and the contrasts with comparable data from mid-ocean ridge basalts to demonstrate the unique way in which water drives melting beneath back-arc spreading centers.

2. Methods and Data Treatment

[7] In order to test the competing models of wet melting, we must be able to determine and compare the extent of mantle melting (F) and the concentration of H_2O in the mantle (H_2O_o) over a broad range of mid-ocean ridge and back-arc basin basalt compositions. Here, we present the methods we used to filter and correct the data, to calculate F and H_2O_o, and to estimate other relevant factors such as mantle potential temperature (T_p), axial depth, and crustal thickness in a self-consistent manner. Readers more interested in an outline of the methodology than in the details may skip to the methods summary presented in section 2.8.

2.1. Back-Arc Basin Basalt Data

[8] We compiled a global data set of major element and volatile measurements for 408 glasses and olivine-hosted melt inclusions from six back-arc basin localities (Mariana trough [*Volpe et al.*, 1987; *Hawkins et al.*, 1990; S&N94; *Gribble et al.*, 1996, 1998; *Newman et al.*, 2000], Sumisu rift [*Hochstaedter et al.*, 1990a], Manus basin [*Danyushevsky et al.*, 1993; *Kamenetsky et al.*, 2001; *Sinton et al.*, 2003; *Shaw et al.*, 2004], North Fiji basin [*Aggrey et al.*, 1988; *Danyushevsky et al.*, 1993], Lau basin [*Aggrey et al.*, 1988; *Danyushevsky et al.*, 1993; *Sinton et al.*, 1993; *Pearce et al.*, 1995; *Kamenetsky et al.*, 1997; *Peate et al.*, 2001; *Kent et al.*, 2002; A. J. Kent, unpublished data, 2002; S. Newman, unpublished data, 1994] and East Scotia ridge [*Muenow et al.*, 1980; *Leat et al.*, 2000; *Fretzdorff et al.*, 2002; *Newman and Stolper*, 1995; S. Newman, unpublished data, 2002]) and three mid-ocean ridge and ridge-like regions (Galápagos spreading center [*Cushman et al.*, 2004], Mid-Atlantic Ridge and Azores platform [*Dixon et al.*, 2002], and Woodlark basin [*Muenow et al.*, 1991; *Danyushevsky et al.*, 1993]).

[9] Before using these data to model F and H_2O_o, we first considered how the composition of each glass may have changed through degassing and crystal fractionation. To assess how degassing may have affected H_2O, we examined another volatile species, CO_2, which has much lower solubility than H_2O in silicate melt and degasses nearly to completion before H_2O degassing initiates during open system degassing [*Dixon et al.*, 1995]. The presence of measurable CO_2 is thus an indicator that H_2O has likely not been significantly affected by degassing, but CO_2 data are not published for many of the samples used in this study. The combined concentrations of H_2O and CO_2 in a melt, however, are pressure sensitive and, in submarine lavas, usually reflect vapor saturation or supersaturation at or near the hydrostatic pressure of eruption [*Dixon and Stolper*, 1995]. We thus calculated the saturation pressure of each sample with respect to pure H_2O (i.e., CO_2 = 0 ppm). Glasses with pure H_2O saturation pressures equal to or greater than their collection depths probably reached saturation with H_2O vapor and experienced some H_2O degassing. To account for uncertainties in eruption/collection depth and the chance that lava flowed downhill from the eruption site, we included only those samples with pure H_2O saturation pressures at least 30 bars less than the pressure at the collection depth. Exceptions to these criteria were made in the case of basaltic melt inclusions, which are often trapped at high pressure within the magma chamber and, if H_2O concentrations are high (e.g., at the Valu Fa ridge), may record H_2O saturation pressures much higher than those at the sample collection depth. In these cases, only melts with measured CO_2 concentrations >30 ppm were considered undegassed.

[10] To compensate for the effects of crystal fractionation, all samples that were determined to have escaped significant H_2O degassing were back corrected to primary mantle melts using several steps. The regional data sets were first filtered to exclude all glasses with MgO < 7 wt %, in order to eliminate highly fractionated compositions. One exception to this criterion was made in the case of the Manus basin, where basalt compositions with MgO $\geq$ 6.25 wt % were included in order to allow a greater number of samples to represent this region. Of the remaining samples, those with MgO < 8 wt % were corrected to 8 wt % MgO using the $Fe_{8.0}$ and $Na_{8.0}$ expressions from *Klein and Langmuir* [1987] and the $TiO_{2(8.0)}$ and $H_2O_{(8.0)}$ expressions from *Taylor and Martinez* [2003]:

$$Fe_{8.0} = FeO - 1.66 \times (8.0 - MgO), \quad (1)$$

$$Na_{8.0} = Na_2O - 0.373 \times (8.0 - MgO), \quad (2)$$

$$TiO_{2(8.0)} = TiO_2 \times MgO^{1.7}/8.0^{1.7}, \quad (3)$$

$$H_2O_{(8.0)} = H_2O \times MgO^{1.7}/8.0^{1.7}. \quad (4)$$

We then employed an additional fractionation correction to 8.5 wt % MgO because most melts are still on olivine + plagioclase cotectics at 8 wt % MgO and adding olivine only (see below) at this point leads to false high Fe. The fractionation slopes for this step are shallower than those above, but consistent with olivine + plag cotectic slopes predicted by the liquid line of descent algorithm of *Weaver and Langmuir* [1990]:

$$Fe_{8.5} = Fe_{8.0} - 1.2 \times (8.5 - 8.0), \quad (5)$$

$$Na_{8.5} = Na_{8.0} - 0.135 \times (8.5 - 8.0), \quad (6)$$

$$TiO_{2(8.5)} = TiO_{2(8.0)} - 0.3 \times (8.5 - 8.0), \quad (7)$$

$$H_2O_{(8.5)} = H_2O_{(8.0)} - 0.05 \times (8.5 - 8.0). \quad (8)$$

Glasses with initial MgO between 8.0 and 8.5 wt % were corrected to 8.5 wt % MgO only along these linear ol + plag cotectics by substituting the measured glass concentrations of FeO, Na_2O, TiO_2 and H_2O for $Fe_{8.0}$, $Na_{8.0}$, $TiO_{2(8.0)}$ and $H_2O_{(8.0)}$ and substituting the measured concentration of MgO for the 8.0 term inside the parentheses in equations (5), (6), (7), and (8). We then back corrected these 8.5% MgO-equivalent compositions, as well as all glasses with initial MgO $\geq$ 8.5 wt % (to which no correction had yet been applied), to primary mantle melts by adding equilibrium olivine (using $K_D^{ol\text{-}liq}$[Mg – Fe] = 0.3) to each glass composition in 1% increments until in equilibrium with Fo_{90} (as in the work of S&N94). These corrected liquid compositions provide $Fe_{(Fo90)}$ and $Na_{(Fo90)}$, which are discussed throughout and are used to calculate mantle potential temperature (see section 2.7), and C^l_{Ti} and $C^l_{H_2O}$ (i.e., $TiO_{2(Fo90)}$ and $H_2O_{(Fo90)}$), which are input to equations (9) and (10) below. Most samples required $\leq$15% olivine addition to reach Fo_{90} equilibrium, but to avoid artifacts from overcorrection, the few samples requiring >30% olivine addition were excluded. These three criteria (degassing, MgO content, and fractionation) are the only constraints used to filter the data; ~50% of the original data compilation (n = 224) passed the filter and were used for the modeling.

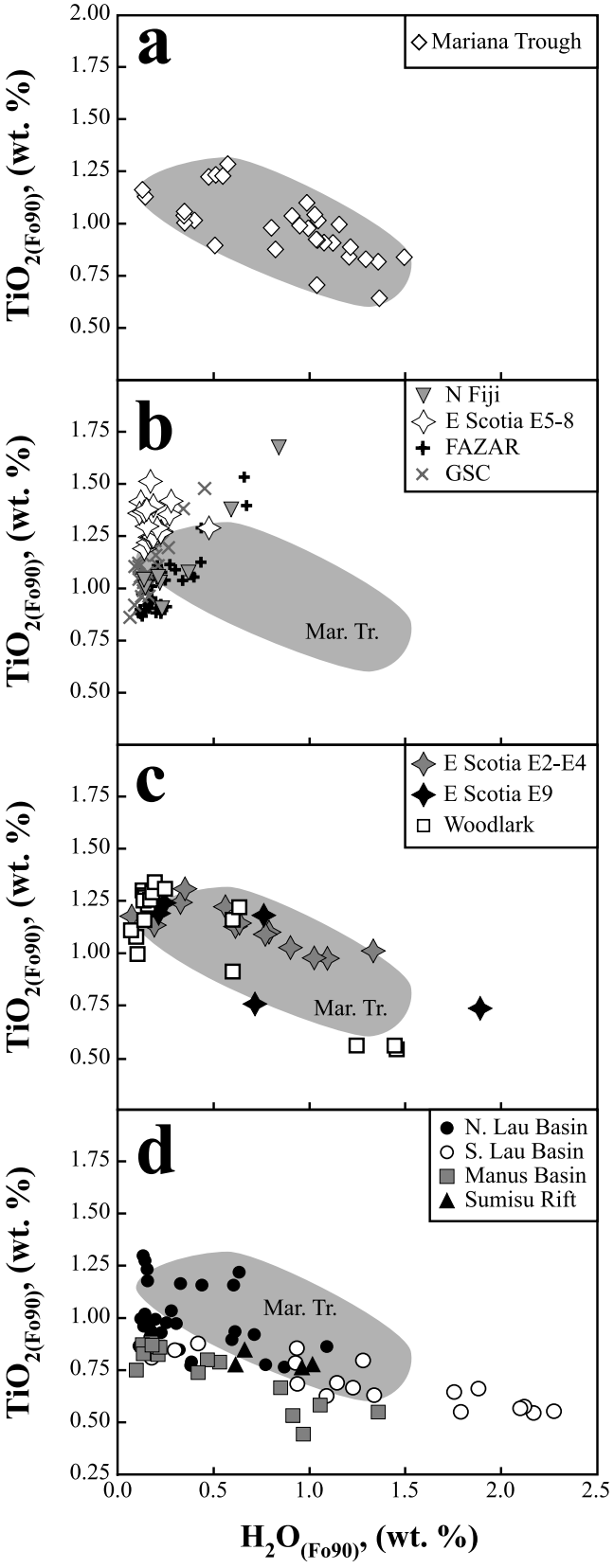

Figure 4. TiO_2 versus H_2O in back-arc basin basalts, calculated to equilibrium with Fo_{90} (see text): (a) Mariana trough (after S&N94); (b) mid-ocean ridges (Mid-Atlantic Ridge (FAZAR), Galápagos spreading center (GSC)) and back-arc basin segments (North Fiji basin, East Scotia ridge E5–E8) showing positive correlations; (c) mid-ocean ridge (Woodlark basin) and back-arc basin segments (East Scotia ridge E2–E4 and E9) showing both negative and positive correlations; and (d) back-arc basin segments showing negative correlations (northern and southern Lau basin, Manus basin, Sumisu rift).

2.2. Modeling Mantle Melt Fractions and Water Contents of Mantle Sources

[11] S&N94 performed a multielement inversion on a suite of Mariana trough glasses to constrain melt fraction and source composition. This inversion was based on the simplifying assumption that variations in Mariana trough basalt chemistry reflect variable extents of melting correlated with mixing of two source components: one similar to the mantle source of normal mid-ocean ridge basalt (NMORB) and the other a water-rich component derived from the subducted slab. They chose the most MORB-like basalt composition and assumed it reflected 5% batch melting of unmodified NMORB source to obtain the composition of this source component. They then used an iterative procedure based on the H_2O, TiO_2, K_2O, Na_2O, and P_2O_5 contents of each glass to solve simultaneously for the composition of the H_2O-rich component, the fraction of this component in the source of each sample, and the extent of melting of the source by which each sample was produced. Although this method succeeds for the Mariana trough, the assumptions (1) of a two-component system (requiring that enough data exist to define clear mixing relationships and that additional components are not involved), (2) of a H_2O-poor NMORB mantle source and a H_2O-rich component that are constant in composition, and (3) of 5% melting to produce the NMORB end-member

Table 1. Mantle Source Constraints for Back-Arc Basins and Ridges

Basin	Region/Segment	Source Model[a]	*f* Removed, %	C^{o}_{Ti}	+[b]	−[b]
Mariana trough		DMM	0	0.133	0.000	0.012
Sumisu rift		DMM	0.10	0.130	0.003	0.003
Lau basin	CLSC	DMM	1.00	0.107	0.014	0.002
	ELSC-VFR	DMM	2.50	0.083	0.024	0.006
	ILSC	DMM	0	0.133	0.000	0.003
	MTJ	DMM	0	0.133	0.000	0.003
Manus basin	MSC	DMM	0.80	0.112	0.015	0.005
	ETZ	DMM	0.80	0.112	0.015	0.005
	ER	DMM	0.15	0.128	0.005	0.004
	SR	DMM	0.80	0.112	0.015	0.005
East Scotia ridge	E2–E4	DMM	0	0.133	0.000	0.006
	E5–E6	DMM	0.15	0.128	0.005	0.004
	E7–E8	DMM	0.40	0.121	0.006	0.005
	E9	Ti/Y	...	0.192	0.007	0.007
North Fiji basin	TJ	Ti/Y	...	0.151	0.012	0.012
	N160°	Ti/Y	...	0.198	0.024	0.024
Woodlark basin	west/D'Entrecasteaux	Ti/Y	...	0.172	0.013	0.013
	center	DMM	0.50	0.119	0.014	0.003
	east	DMM	0.40	0.121	0.012	0.005
	northeast	DMM	0.10	0.130	0.003	0.003
GSC	E	*Cushman et al.* [2004]	...	0.143	...	...
	T	*Cushman et al.* [2004]	...	0.108	...	...
	N	*Cushman et al.* [2004]	...	0.105	...	...
Azores	KP	Ti/Y	...	0.189	0.014	0.014
	PO	Ti/Y	...	0.143	0.002	0.002
	OH	Ti/Y	...	0.139	0.003	0.003
	HA	Ti/Y	...	0.137	0.003	0.003

[a]DMM source model denotes the use of prior melt extraction to constrain source variation. Ti/Y source model denotes cases where the mantle source is enriched, and the TiO_2/Y model is used to constrain source characteristics.
[b]Columns denote errors on the value of C^{o}_{Ti}.

may not apply to other regions. An alternative approach is therefore necessary because, in many regions, data sets are limited, the compositions of the mantle source and/or slab-derived component may vary regionally, and the extent of melting of the MORB-like end-member is unlikely to be constant for all back arcs.

[12] Rather than performing a best fit based on several relatively incompatible elements (including some likely present in significant concentrations in the slab-derived components) to constrain F for each sample, we used TiO_2 in glasses as a single-element proxy for melt fraction (F). Like any other element that is incompatible during mantle melting, TiO_2 decreases monotonically in the melt with increasing F of a single source. While it is difficult to identify a truly "conservative" element in subduction zones (i.e., an element not added from the subducted slab [*Pearce and Parkinson*, 1993]), TiO_2 holds promise in this regard because of its low overall abundance in arc lavas and the possibility that it, like other high field strength elements, may be retained in residual rutile in the slab during dehydration [e.g., *Ryerson and Watson*, 1987]. We thus adopted the approximation that, except for dilution, the TiO_2 content of sources in a given mantle wedge are independent of the amount of slab-derived component they contain. Moreover, systematics of Mariana trough basalts are consistent with low TiO_2 in the H_2O-rich component (S&N94), as the data show that increasing H_2O concentrations correlate with decreasing TiO_2 and, by proxy, increasing F (Figure 4a (see also S&N94). This correlation is the starting point of the linear wet melting function and was indeed the observation that led S&N94 to first suspect a positive correlation between the water content and melt fraction of the source of Mariana trough lavas.

[13] On the basis of this observation for the Mariana trough (Figure 4a), we used TiO_2 to calculate F for all the filtered and corrected back-arc basin basalt and MORB data. To model F, the fraction of melting, we solved the batch melting equation to yield

$$F = \frac{\left(C^{o}_{Ti}/C^{l}_{Ti}\right) - D_{Ti}}{(1 - D_{Ti})}, \tag{9}$$

where C^{o}_{Ti} is the concentration of TiO_2 in the mantle source (Table 1; see sections 2.4–2.5), C^{l}_{Ti} is the concentration of TiO_2 in the melt in equilibrium with Fo_{90} (from basalt data, see section 2.1), and D_{Ti} is the bulk distribution coefficient for Ti during mantle melting (Table 2; see section 2.3). To obtain the concentration of H_2O in the source, we re-solved this equation in terms of H_2O to yield

$$C^{o}_{H_2O} = C^{l}_{H_2O}[F(1 - D_{H_2O}) + D_{H_2O}], \tag{10}$$

where $C^{o}_{H_2O}$ is the concentration of H_2O in the mantle source (hereinafter referred to as H_2O_o), $C^{l}_{H_2O}$ is the concentration of H_2O in the melt in equilibrium with Fo_{90} (from basalt data, see section 2.1), F is the output of equation (9), and D_{H_2O} is the bulk distribution coefficient

Table 2. Mantle/Melt Partition Coefficients and Starting Composition

Element	D (Mantle/Melt)	DMM[a]
Th, ppm	0.0015	0.0137
Nb, ppm	0.003	0.21
Ta, ppm	0.005	0.0138
La, ppm	0.0075	0.234
Ce, ppm	0.011	0.772
H_2O, ppm	0.012	116
Pr, ppm	0.015	0.131
Nd, ppm	0.02	0.713
Sm, ppm	0.03	0.27
Zr, ppm	0.03	7.94
Eu, ppm	0.035	0.107
Gd, ppm	0.04	0.395
TiO_2, wt %	0.04	0.133
Tb, ppm	0.048	0.075
Dy, ppm	0.06	0.531
Ho, ppm	0.07	0.122
Er, ppm	0.08	0.371
Yb, ppm	0.095	0.401
Y, ppm	0.095	4.07
Lu, ppm	0.105	0.063

[a]Depleted MORB mantle composition from *Salters and Stracke* [2004].

for H_2O during mantle melting (from S&N94; see Table 2 and section 2.3).

[14] While end-member batch melting is probably not a realistic melting process, it is a useful approximation for determining the bulk F of basaltic partial melts of the mantle. For example, equation (9) provides a fair approximation of the bulk melt fraction (as reflected in incompatible element concentrations) at mid-ocean ridges despite the likely complexity of the melting regime and melting process in this environment [*Plank and Langmuir*, 1992; *Plank et al.*, 1995]. As a relevant example, the $Na_{8.0}$ and $Ti_{8.0}$ variations in mid-ocean ridge basalts are well described by pooling of melts produced by polybaric, adiabatic, fractional melting with Ds that vary as a function of pressure, temperature, composition, and mode (Figure 5 [see also *Langmuir et al.*, 1992]). The $Na_{8.0}$-$Ti_{8.0}$ trend, however, is also adequately reproduced with the simplest model (batch melting with constant C^o and Ds (Figure 5)). Since back arcs tap mantle similar to MORB, and because we know less about the melting process in these settings, we adopted this simple model as our starting point.

2.3. Partition Coefficients

[15] The partition coefficients used in the batch melting model impact the results of these calculations. The D_{Ti} used here derives from the simplified MORB model above (Figure 5) and describes partitioning between melt and lherzolitic mantle. Published mantle/melt Ds for Ti range up to 0.11 [*Kelemen et al.*, 1993], but a value this high is not permitted in the batch melting model, as it produces negative values for F in high-Ti samples. The value used here (0.04) reflects an average D_{Ti} within the range predicted for appropriate variations in pressure, temperature, and melt composition along a polybaric path (e.g., 0.01–0.06 [*Langmuir et al.*, 1992]; <0.01–0.08 [*Baker et al.*, 1995]), and we assess the effect of this uncertainty on the calculations below. In rare cases, extents of melting predicted by the batch melting model exceed 20%. At such high melt fractions, one might expect clinopyroxene (cpx) to become exhausted from the mantle, and the liquid to instead reflect equilibrium partitioning between melt and a two-phase harzburgitic mantle. For the few back-arc samples where cpx may have been exhausted from the mantle (e.g., the Manus basin, based on CaO/Al_2O_3), we selected a harzburgite/melt D_{Ti} of 0.05 (average of Ds reported by *Grove et al.* [2002], *Pearce and Parkinson* [1993], and *Kelemen et al.* [1993]). Because this D is similar to the lherzolite D, the change has a small effect on the calculated F and H_2O_o. The D_{H_2O} used here (0.012) is adopted from S&N94 in order to be consistent with that study. The value is similar to that for the LREE, consistent with variations observed in MORB [*Michael*, 1988; *Dixon et al.*, 2002]. We evaluate the effect of lowering D_{H_2O} in section 2.6 below.

2.4. Modeling Depleted Mantle Source Compositions

[16] The composition of the depleted mantle end-member in the sources of back-arc basin magmas is crucial to the model results, and so we carefully consider here the likely variation of C^o_{Ti} in global back-arc basin and ridge regions. Two variables primarily control the TiO_2 concentration of a partial melt of mantle peridotite: the initial concentration of TiO_2 in the peridotite and the melt fraction. So while *Taylor and Martinez* [2003] attribute TiO_2 variations in back-arc basins largely to variations in source composition, variations in F would also lead to TiO_2 variations for a constant source composition, and the challenge is to distinguish whether a melt with low TiO_2 reflects high F, or a highly depleted mantle, or both. The key to doing so is to use not just TiO_2, but several elements with different Ds during mantle melting; i.e., although the concentration of a single element in a melt can be modeled by an infinite number of combinations of source concentration and F, the relative abundances of elements with differing mantle/melt Ds can

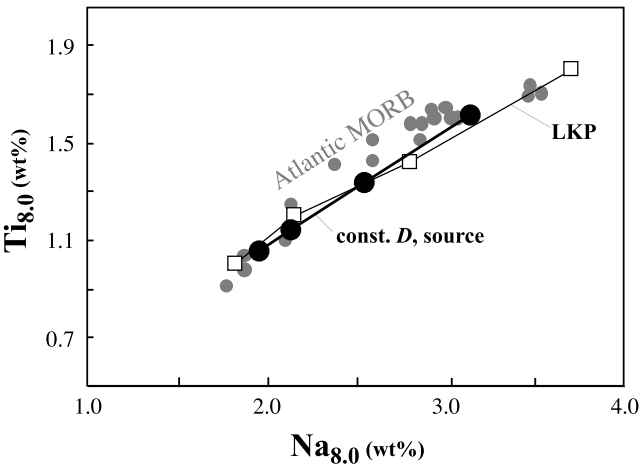

Figure 5. $Ti_{8.0}$ versus $Na_{8.0}$ in North Atlantic MORB (shaded circles [*Langmuir et al.*, 1992]) with model fits to the data trend. The thin black line with open squares (LKP) is a polybaric, adiabatic, fractional, pooled melting model, with Ds varying as a function of pressure, temperature, composition, and mode (taken from *Langmuir et al.* [1992, Figure 55]). The bold line with solid circles (const. D, this study) is batch melting with constant Ds (D_{Na} = 0.02, D_{Ti} = 0.04) and constant mantle source (C^o_{Na} = 0.29 wt %, C^o_{Ti} = 0.133 wt %).

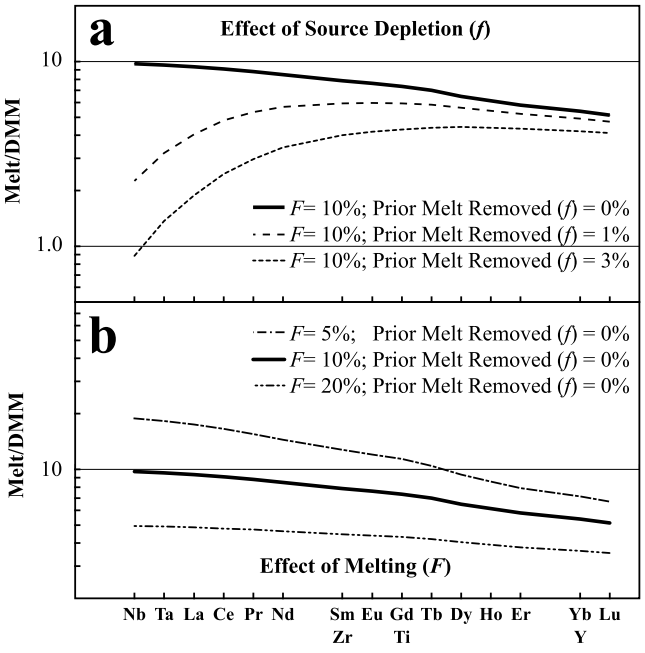

Figure 6. Model melt trace element compositions normalized to DMM of *Salters and Stracke* [2004]. (a) Illustrating the effect of mantle source variation on the shape and position of the trace element pattern, keeping F constant at 10%. The curves show the effects of 0, 1, and 3% prior batch melt removal from source on the shape of the pattern. (b) Illustrating the effect of variable melt fraction on the shape and position of the trace element pattern, keeping source concentrations constant (at DMM (Table 2)). Curves show 5, 10, and 20% batch melts of unmodified DMM, using Ds as in Table 2.

distinguish variations in source composition from variations due to differing extents of melting. For example, source depletion affects highly incompatible elements (e.g., Nb) more than moderately incompatible elements (e.g., Ti (Figure 6a)), while high degrees of melting will affect both elements almost equally provided $F \gg D$ (Figure 6b). Here, we attempt to account for TiO_2 variations in mantle sources using a model based on the systematic variations of a suite of elements in lavas from each basin.

[17] A complicating factor in separating variations in the MORB-like mantle composition from variations in extent of melting at subduction zones is that there are probably no elements in the mantle wedge that are completely unaffected by the mass flux from the subducting slab. Nevertheless, for back-arc basin lavas, we will assume most trace element enrichments (e.g., LREE) are additions from the slab-derived, H_2O-rich component, whereas incompatible element depletion is a feature of the MORB-like mantle. We therefore chose several relatively conservative elements, spanning a range of mantle/melt Ds (Y, Ti, Zr, Nb, and Ta) to set constraints on mantle source variations. These high field strength elements vary systematically during mantle melting at mid-ocean ridges and, unlike the light rare earth elements and Th, are expected to be retained in the slab by residual rutile [*Pearce and Parkinson*, 1993]. These are thus among the elements in the mantle least affected by addition of slab-derived components.

[18] Figure 7 illustrates how we determined the composition of the end-member mantle sources for the back-arc spreading centers, to ultimately derive C^o_{Ti}. We did this by fitting the conservative element pattern (rather than the full trace element pattern, most of which is sensitive to slab additions (e.g., Figure 7)) of natural lava samples with a melt removal model to determine the source composition of the ''slab-free'' mantle end-member for each basin or subregion within a basin. Mantle source variation is expressed as a function of prior melt removal from an average mantle composition. We chose as the starting point the depleted MORB mantle (DMM) composition of *Salters and Stracke* [2004], which represents an average composition of the mantle that provides the source of MORB. The melt removal model controls the trace element characteristics of the mantle source by removing a specified fraction of melt (f) from DMM using batch melting and the bulk mantle/melt Ds in Table 2. That is, f represents a melt fraction that has been removed from the mantle prior to the melting that takes place within the ridge or back-arc magma source. The removal of f creates depletion of the mantle before it enters into the ridge and back-arc sources. Subsequent melting of these depleted sources, which is the melting step that generates mid-ocean ridge and back-arc basin magmas (F, as defined in section 2.2), employs these same Ds. Source depletion (the removal of f from DMM) has a large effect on the shape of the trace element pattern of a melt (Figure 6a), whereas F controls the absolute abundances of all elements, shifting the whole pattern up or down with a minimal effect on shape (Figure 6b), because f is usually a small value (<3%) and F is a large value (typically 10–20%) with respect to the Ds modeled. Prior melt removal (f) has the largest effect on the most incom-

Figure 7. Trace element models constraining mantle source characteristics for basins and segments with depleted mantle sources (equal to or more depleted than DMM), normalized to DMM of *Salters and Stracke* [2004]. Figures for each region are split between two panels. Top panels show complete trace element patterns for two melt compositions encompassing the geochemical range of the region, with Th and the REE in filled symbols and the conservative elements (CEs; see text) in open symbols. Bottom panels isolate the CEs, with bold and dashed lines showing model curves for the CE patterns, varying F and C^o_{Ti} as illustrated in Figure 6. The shaded fields around these model curves illustrate the errors on the model (see Table 1), constrained by determining the range of model pattern shapes permissible by any pair of CEs (including 10% errors in the concentration of each element; see text). Within a given basin, segments with similar mantle characteristics are grouped together. (a, b) Mariana trough. (c, d) Manus basin, ER segment. (e, f) Manus basin, MSC, ETZ, and SR segments. (g, h) Lau basin, MTJ and ILSC segments. (i, j) Lau basin, CLSC segment. (k, l) Lau basin, ELSC and VFR segments, with additional curve (solid shaded line) illustrating the misfit of a variable C^o_{Ti}-constant F model. (m, n) Sumisu rift. (o, p) Woodlark basin, NE segment. (q, r) Woodlark basin, east segment. (s, t) East Scotia ridge, E2–E4 segments. (u, v) East Scotia ridge, E5–E6 segments. (w, x) East Scotia ridge, E7–E8 segments. See text for data sources.

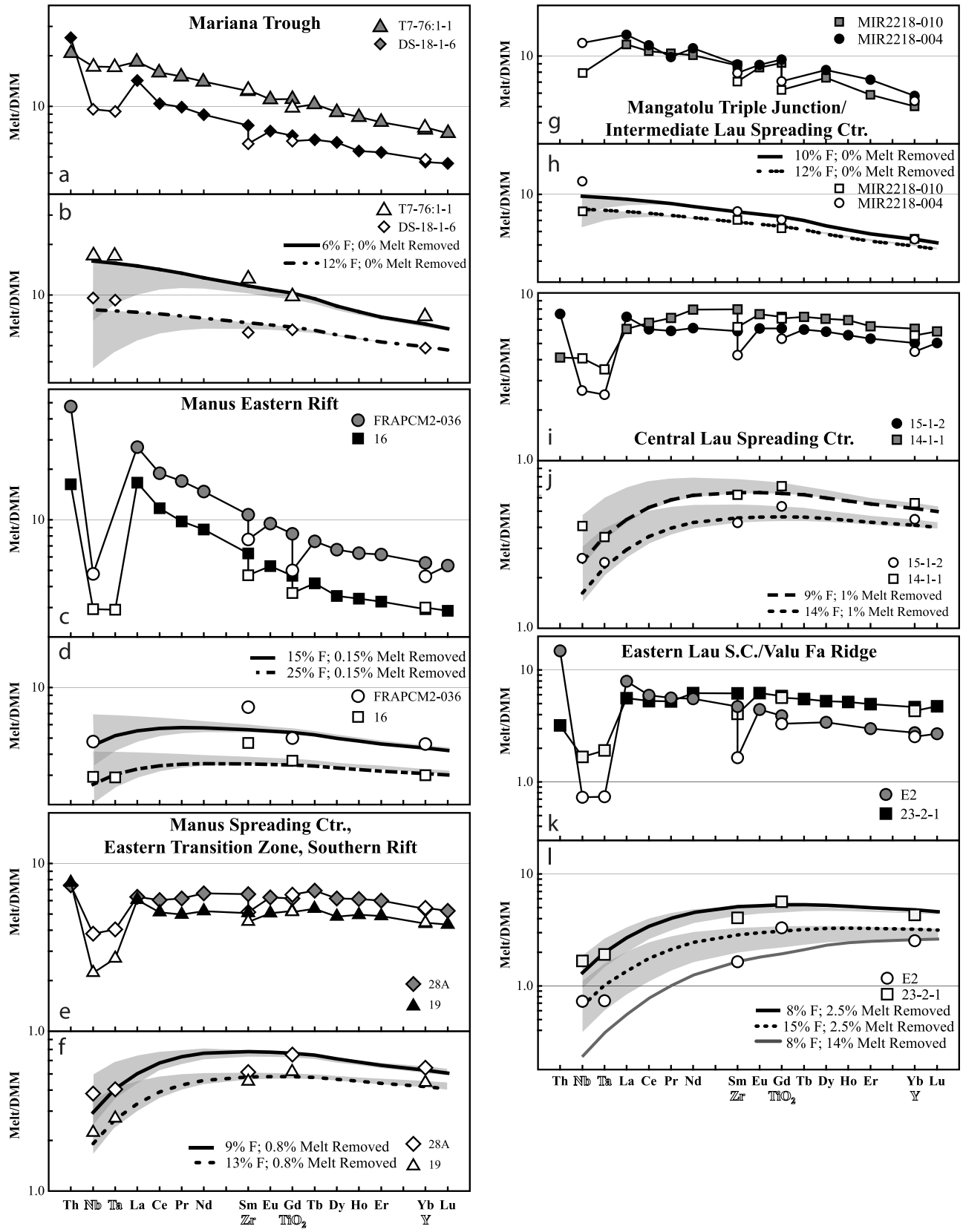
Mariana Trough
T7-76:1-1
DS-18-1-6
a
b
6% F; 0% Melt Removed
12% F; 0% Melt Removed
Manus Eastern Rift
FRAPCM2-036
16
c
d
15% F; 0.15% Melt Removed
25% F; 0.15% Melt Removed
Manus Spreading Ctr.,
Eastern Transition Zone, Southern Rift
28A
19
e
f
9% F; 0.8% Melt Removed
13% F; 0.8% Melt Removed
MIR2218-010
MIR2218-004
g
Mangatolu Triple Junction/
Intermediate Lau Spreading Ctr.
h
10% F; 0% Melt Removed
12% F; 0% Melt Removed
i
Central Lau Spreading Ctr.
15-1-2
14-1-1
j
9% F; 1% Melt Removed
14% F; 1% Melt Removed
Eastern Lau S.C./Valu Fa Ridge
E2
23-2-1
k
l
8% F; 2.5% Melt Removed
15% F; 2.5% Melt Removed
8% F; 14% Melt Removed
Melt/DMM
Th Nb Ta La Ce Pr Nd Sm Eu Gd Tb Dy Ho Er Yb Lu
Zr TiO2 Y

Figure 7

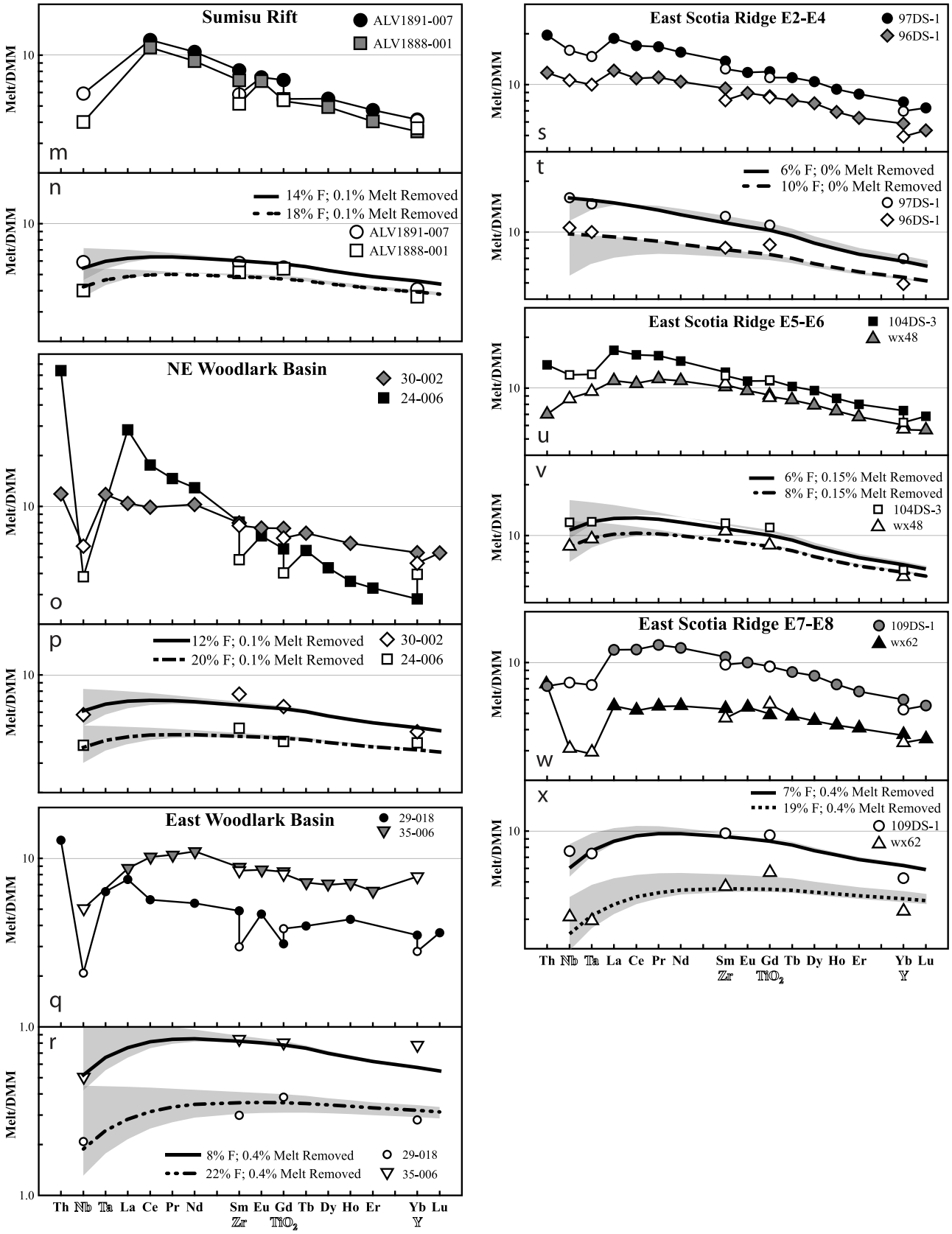

Figure 7. (continued)

patible elements (e.g., Nb and Ta), which may change in source concentration by a factor of 10 with as little as 2.5% melt removed, whereas TiO_2 would vary by only a factor of 1.5, using the batch melting model. Removing melt using a fractional melting model would lead to even smaller changes in TiO_2 relative to the more incompatible elements, and we thus used the batch removal model because it allows for greater variation of TiO_2 in the mantle source. The concentration of Nb or Ta, relative to TiO_2 (and other elements), therefore provides a constraint on the amount of depletion necessary in the mantle source of each region. This constraint is a minimum in the sense that Nb and Ta could theoretically be added from the slab, causing the mantle to appear less depleted than it actually is. These constraints are then used to estimate the mantle source composition and C^o_{Ti} for most back-arc spreading segments (Table 1, Figure 7), and we apply a constant source composition to each segment. We unfortunately cannot, in many cases, perform this analysis with the exact samples used to model F and H_2O_o because most of these glasses do not have the necessary trace element data. In such cases, trace element data from whole rock samples in the same local area were used to establish regional source characteristics beneath each spreading segment [*Hochstaedter et al.*, 1990b; *Nohara et al.*, 1994; *Pearce et al.*, 1995; *Dril et al.*, 1997; *Fretzdorff et al.*, 2002; *Sinton et al.*, 2003].

[19] Figure 7 illustrates back-arc regions where the mantle sources were successfully modeled relative to DMM. The conservative elements (CEs) Nb, Ta, Zr, Ti, and Y reflect mantle source characteristics, whereas rare earth elements and Th indicate the extent of geochemical enrichment from the slab. As a general rule, lavas with lower absolute concentrations of CEs (i.e., higher F) show greater absolute abundances in slab-derived elements than low-F lavas from the same regions, suggesting a link between F and the extent of slab-derived input at back-arc basins (such that larger extents of melting are associated with greater slab input of the non-CEs). The mantle depletion model developed here makes a fair approximation of the CE patterns and thus the mantle source characteristics of basins and intrabasin segments tapping variably depleted mantle. In fact, most regions require little to no prior melt removal ($f <$ 0.2%) to explain the CE patterns (e.g., Mariana trough, Sumisu rift, MTJ, ILSC, Manus ER, ESR E2–E6, Woodlark basin NE (Table 1)). The CE patterns are also generally parallel within a given basin spreading segment (F variation) rather than of variable shape (source variation), which supports the assertion that F is the dominant variable controlling CE variation on local scales. A few regions require more depleted mantle (~1–2.5% f; e.g., CLSC, ELSC-VFR, Manus MSC-ETZ (Table 1)), which relates to low Nb concentrations (<1 ppm).

[20] We acknowledge that the model curves are not perfect fits to the CE patterns in Figure 7, likely due to variable data quality (particularly for Nb and Ta) and possible additions from the slab in some cases. Conservative trace element variations could, alternatively, arise dominantly or exclusively from variations in mantle source composition. We evaluated this possibility by holding F constant and attempting to fit the CE patterns of lavas only through changing mantle depletion. This model falls short in two ways. First, the extent of prior melt removal necessary to explain variations in the less incompatible elements Zr and Y removes too much Nb and Ta to match the shape of the full CE pattern in multiple samples from a given region (e.g., Figure 7l). Niobium and tantalum could, however, be added from the slab, which would elevate their concentrations above the ambient mantle and explain why the full CE pattern is poorly modeled. Second, the extents of prior melt removal (f) necessary to broadly reproduce the

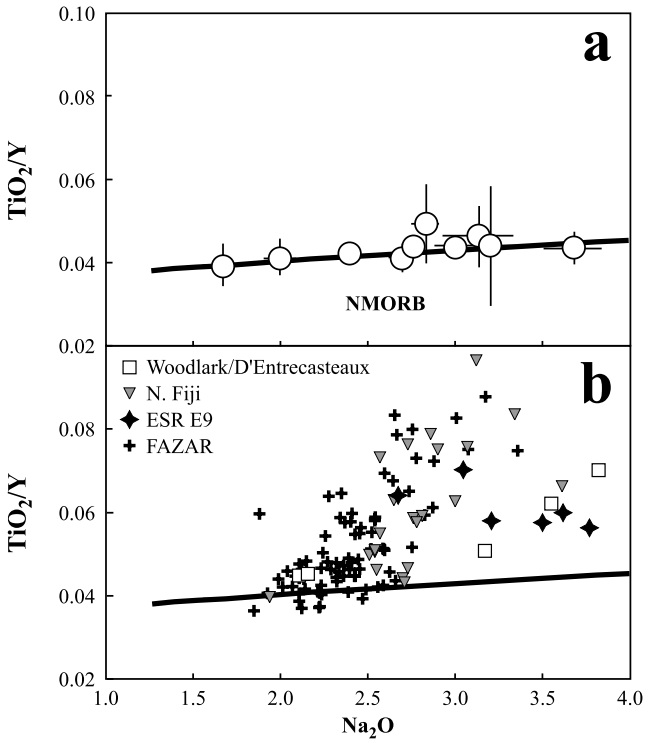

Figure 8. TiO_2/Y versus Na_2O, demonstrating the model used to constrain mantle source characteristics for segments with enriched mantle sources (more trace element enriched than the DMM of *Salters and Stracke* [2004]). TiO_2 and Y have similar mantle/melt Ds for N-MORB, whereas absolute concentrations of incompatible elements like Na_2O, which are relatively invariant in the mantle source, vary systematically as a function of F. (a) TiO_2/Y versus Na_2O in MORB, with a batch melting model curve (bold line) approximating the MORB trend, showing the relative invariance of the TiO_2/Y ratio with extent of melting at mid-ocean ridges. Regional MORB averages (samples with 7.5–8.5 wt % MgO) are from the following regions: the Kolbeinsey ridge [*Devey et al.*, 1994], the Australian-Antarctic discordance [*Klein et al.*, 1991], the Indian Ocean triple junction [*Price et al.*, 1986], the Tamayo region [*Bender et al.*, 1984], the Pacific seamounts [*Niu and Batiza*, 1997], the Cayman rise and the Reykjanes ridge (www.petdb.org), and the Mid-Atlantic Ridge between the Kane and Hayes fracture zones (www.petdb.org). Error bars are one standard deviation of the regional average. (b) TiO_2/Y versus Na_2O in ridge and back-arc regions with enriched mantle (Table 1) showing high TiO_2/Y, attributed here to source enrichment. The points shown are individual glass samples. The bold line is the MORB batch melting model curve from Figure 8a.

geochemical range recorded, for example, along the Lau basin (Figure 7l) would correlate with variations in H_2O concentration along the length of the basin, and indeed in global basins. The Lau basin mantle composition clearly varies along strike regardless of how we model it, but there is no strong argument for why large variations of H_2O concentrations in the source regions of arcs and back arcs should correlate globally with the extent of prior mantle depletion [*Eiler et al.*, 2000]. Water concentrations more sensibly correlate with additions of slab-derived components and with F. We therefore elect to constrain variations in the water-poor, MORB-like component of the mantle beneath back-arc basins using the approach described above whereby this component is assumed to have been derived from DMM by variable amounts of prior batch melt extraction, and where the amount of this prior melt extraction is constrained using a fit to the HFSE that assumes they are truly conservative (i.e., uninfluenced by the addition of water-rich, slab-derived components). To the degree that these conservative elements might have been influenced by addition from the slab, the absolute magnitude of their variations in the MORB-like component in the sources of back-arc magmas may require modification, and the source constraints we present (Table 1) would represent the minimum amount of depletion (and therefore the maximum C^o_{Ti}) necessary to explain the geochemistry of these back-arc basins.

2.5. Modeling Enriched Mantle Source Compositions

[21] The melt removal–based source model also only succeeds in regions where the mantle is similar to, or more depleted than, the initial DMM composition. In regions where the mantle is more trace element–enriched than DMM, the trace element patterns have steeper negative slopes than unmodified DMM and cannot be explained by prior melt removal from this MORB-like source. In such cases we employed another approach to approximating the H_2O-poor mantle sources. In the case of the Galápagos spreading center, we adopted the source constraints of *Cushman et al.* [2004], but in other regions where enriched mantle source characteristics have not been so rigorously evaluated, we applied a simple model based on TiO_2/Y systematics. The ratio of TiO_2 to Y varies little during mantle melting at mid-ocean ridges (TiO_2/Y = 0.04–0.05 for regionally averaged MORB (Figure 8)) because these elements have similar mantle/melt Ds. On the other hand, regions near hot spots or having previously recognized, isotopically enriched mantle have high TiO_2/Y ratios (e.g., FAZAR, N. Fiji basin (Figure 8) [*Eissen et al.*, 1994; *Asimow et al.*, 2004]). Some of the high ratios could arise from a higher D_Y in these sources (which could contain garnet), but we elect to treat TiO_2/Y ratios in excess of mean MORB entirely as TiO_2 enrichment of the H_2O-poor mantle source, in order to provide a maximum constraint on C^o_{Ti}; that is,

$$C^o_{Ti} = \left[\frac{(TiO_2/Y)_{sample}}{(TiO_2/Y)_{MORB}}\right] \times C^{DMM}_{Ti}, \quad (11)$$

where $(TiO_2/Y)_{sample}$ is the TiO_2/Y ratio of the glass, $(TiO_2/Y)_{MORB}$ is 0.04, and C^{DMM}_{Ti} is 0.133 (Table 2). This likely overestimates C^o_{Ti} and leads to maximum values for F and H_2O_o. We thus present the results below using a range of C^o_{Ti} values, but in all cases comparing results using the constant source model (DMM [*Salters and Stracke*, 2004]) to maximum C^o_{Ti} constraints from TiO_2/Y in enriched regions and CE constraints for depleted regions (Table 1).

2.6. Uncertainties and Errors

[22] The effects of uncertainties in measurements and model assumptions on the final calculations of F and H_2O_o require careful consideration. We first addressed the confidence level of each source of uncertainty. For analytical measurements (TiO_2 and H_2O), we used the standard deviation of replicate measurements on any given sample to reflect the uncertainty in the data (±5% for TiO_2, ±10% for H_2O). For C^l_{Ti}, both the correction for crystal fractionation and the possibility of slab-derived TiO_2 in the melt contribute to errors in this value, and we used an uncertainty of ±20%. Minimum errors in the mantle source composition (C^o_{Ti}) arise from analytical uncertainties in the measurement of the CEs used to constrain the source characteristics. Uncertainties on C^o_{Ti} were thus constrained by allowing 10% variation in all of the CEs in the melt removal model for each region and by constraining the full range of f values (as shown by pattern shapes in Figure 7) permissible by any pair of CEs within this 10% concentration variation (see Table 1). On average, this analysis results in $\leq$10% error in C^o_{Ti}. We explored the effect of a ±50% variation in D_{Ti} (0.02–0.06), which encompasses the range over which the instantaneous value of D_{Ti} may vary along a polybaric path (see section 2.3 [*Langmuir et al.*, 1992]). The D_{H_2O} (0.012) used here originates from S&N94 and is consistent with the D_{Ce} used here (Table 2), but recent experiments suggest lower values for both D_{H_2O} and D_{Ce} during batch melting of spinel lherzolite (0.007–0.009 [*Aubaud et al.*, 2004; *Hauri et al.*, 2006]. Studies of MORB and OIB data also suggest a lower value for D_{H_2O}, closer to D_{La} (0.01 [*Dixon et al.*, 2002]), and we thus also evaluated the effect of lowering D_{H_2O} in our error analysis but use D_{H_2O} = 0.012 for the final calculations to remain consistent with S&N94. An additional test of the modeling outcome, using Na_2O instead of TiO_2 as a proxy for F, is provided in the auxiliary material[1]. Using this alternate proxy for F yields results similar to those for TiO_2. Our results are thus not dependent on the use of TiO_2 as a proxy for F, nor on factors specific to TiO_2 (e.g., the presence of rutile, ilmenite, and Ti-clinohumite in the mantle).

[23] We used a Monte Carlo error analysis, which allows each parameter to vary simultaneously within its assigned uncertainty during the calculation of F and H_2O_o, to evaluate the effects of these uncertainties on the model trends. This technique produces error ellipses around the modeled points in H_2O_o versus F, which represent the outcome of 90% of the trial calculations (Figure 9a). Elongation of the ellipses relative to the origin is largely a product of the uncertainties in C^o_{Ti} and D_{Ti}, but also serves to emphasize that errors on this diagram are highly correlated. F is used to calculate H_2O_o, and any error contributing to the determination of F therefore propagates into error in

[1] Auxiliary materials are available in the HTML. doi:10.1029/2005JB003732.

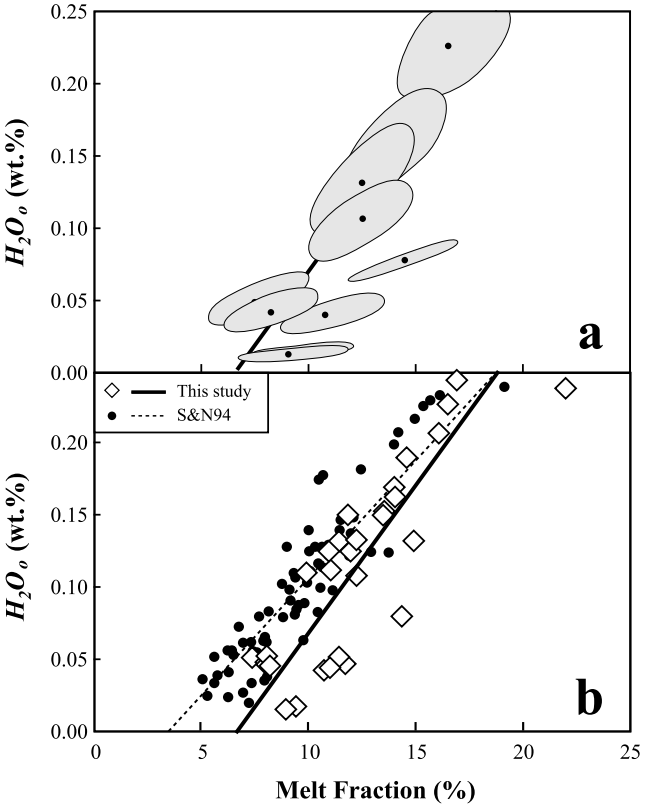

Figure 9. H_2O_o (H_2O concentration of the mantle source) versus F (mantle melt fraction) showing (a) error ellipses for 10 example points in the Mariana trough (MT). The bold line is a linear regression through the full MT data set, as shown in Figure 10a. The equation of the line is $y = 2.082x - 0.142$ ($r^2 = 0.798$), errors on the slope are ±10%, and errors on the intercept are ±30%. (b) Comparison between results of S&N94 and this study. The black circles and short-dashed line are the results and linear best fit from S&N94 ($y = 1.628x - 0.055$; $r^2 = 0.840$). The open diamonds are the results from this study, utilizing a larger initial glass data set than S&N94, and the solid line is the best fit line to these points ($y = 2.082x - 0.142$; $r^2 = 0.798$). This study filtered the data differently than did S&N94, which is why this study plots fewer points.

H_2O_o. This analysis results in up to 30% uncertainty in the F-intercept (melt fraction at zero H_2O_o) of a regressed line but relatively little variation in the slope (~10%), which indicates that the slopes of the model trends are more robust than the intercepts.

[24] Both the S&N94 method and the procedure outlined here result in similar trends despite employing different geochemical constraints and data sets for determining F and H_2O_o in the Mariana trough. Figure 9b compares the results from S&N94 with the results from this study. For the two best-fit lines shown in Figure 9b (one regressed through the S&N94 points and the other through the points from this work), the absolute difference in F-intercept is large (3.4 versus 7.0%), owing in part to the extra steps taken here to correct for crystal fractionation, which result in lower TiO_2 in the modeled melt compositions. The slopes of the two trends, however, differ by only 28% (1.63 versus 2.08). This difference in slope translates into relatively small differences in H_2O_o at a given F (e.g., 0.07 versus 0.10 wt % H_2O_o at F = 10%), and is probably within the analytical uncertainty in the H_2O measurement. We regard this comparison as a successful reproduction of the original S&N94 trend, using the TiO_2 proxy for F.

2.7. Mantle Temperature, Axial Depth, Crustal Thickness, and Melt Production

[25] Mantle potential temperature (T_p) is a primary factor in melt production at mid-ocean ridges, and it likely strongly influences melt generation beneath back-arc basins as well. At ridges, mantle T_p controls the depth/pressure at which melting initiates (P_o) and thus the overall extent of melting (F), which relates to the length of the melting column [*Klein and Langmuir*, 1987; *McKenzie and Bickle*, 1988]. The FeO concentration of a melt is sensitive to P_o, whereas the concentrations of incompatible elements such as Na_2O or TiO_2 vary inversely with F [*Klein and Langmuir*, 1987; *Kinzler and Grove*, 1992]. In Figure 2, the correlation of global MORBs in $Na_{(Fo90)}$ versus $Fe_{(Fo90)}$ (i.e., Na_2O and FeO concentrations corrected for crystal fractionation to equilibrium with Fo_{90}; see section 2.1) is consistent with variations in T_p and P_o [*Langmuir et al.*, 1992; *Kinzler*, 1997; *Asimow et al.*, 2001]. We used the pooled, accumulated fractional melting model of *Langmuir et al.* [1992] to calculate T_p from dry, MORB-like back-arc basin melts ($H_2O \leq 0.5$ wt % (Figure 2, Tables 3 and 4)) using $Na_{(Fo90)}$ and $Fe_{(Fo90)}$ parameterized as follows:

$$T_p(\mathrm{Fe}) = \left[3.4381 \times \left(\mathrm{Fe}_{(\mathrm{Fo90})}\right)^2\right] + \left[4.154 \times \left(\mathrm{Fe}_{(\mathrm{Fo90})}\right)\right] + 1088.6 \quad (12)$$

$$T_p(\mathrm{Na}) = \left[41.164 \times \left(\mathrm{Na}_{(\mathrm{Fo90})}\right)^2\right] - \left[336.78 \times \left(\mathrm{Na}_{(\mathrm{Fo90})}\right)\right] + 1956.1. \quad (13)$$

Values of T_p were calculated for each mid-ocean ridge and back-arc basin by averaging T_p(Fe) and T_p(Na) for each sample, then averaging T_p(sample) for all samples in each basin (Table 4). In the few cases where the standard deviation between the two T_p calculations from a single glass composition was >50°C, then only T_p(Na) was used in the regional average because the correction of Fe concentrations in high-Fe glasses to Fo_{90} has greater error than the Na correction.

[26] Axial depth and crustal thickness at mid-ocean ridges are also linked to total melt production, reflecting the integrated effects of mantle temperature, wet melting, and melt removal. High mantle temperature and large extents of melting combine to generate thick oceanic crust that, due to isostasy, lies at shallower water depths (i.e., lower axial depth) than regions with thinner crust. Axial depth and crustal thickness relate quantitatively to the extent of melting as reflected, for example, by the correlation of depth with melt fraction proxies (e.g., $Na_{8.0}$ [*Klein and Langmuir*, 1987]; see Figure 1). In order to explore the extent to which the chemical and physical relationships observed at mid-ocean ridges extend to back-arc basins, we determined mean axial depths at back-arc basins either from

Table 3. Geochemical Variables at Back-Arc Basins

		Dry Samples Only (H_2O < 0.5 wt %)						All Samples					
Basin	Segment	CaO/Al_2O_3	1σ	$Fe_{(Fo90)}$	1σ	$Na_{(Fo90)}$	1σ	CaO/Al_2O_3	1σ	$Fe_{(Fo90)}$	1σ	$Na_{(Fo90)}$	1σ
						Small-Scale Averages							
East Scotia ridge	E2–E4	0.69	0.02	7.48	0.46	2.66	0.13	0.68	0.01	6.62	0.25	2.62	0.07
	E5–E8	0.73	0.01	8.18	0.11	2.69	0.07	0.73	0.01	8.07	0.15	2.70	0.07
	E9	0.71	0.02	7.03	0.08	3.06	0.04	0.71	0.02	6.49	0.41	2.85	0.20
Lau basin	MTJ	0.76	0.02	9.08	0.18	2.20	0.04	0.79	0.02	8.24	0.27	2.17	0.07
	CLSC	0.82	0.02	9.51	0.25	1.92	0.03	0.82	0.02	9.51	0.25	1.92	0.03
	ILSC	0.76	0.03	8.22	0.47	1.84	0.18	0.76	0.03	8.22	0.47	1.84	0.18
	ELSC	0.79	0.01	8.76	0.40	1.55	0.04	0.80	0.01	8.73	0.23	1.57	0.07
	VFR							0.86	0.01	10.34	0.41	1.12	0.06
Manus basin	ER							0.76	0.09	8.35	0.00	1.28	0.00
	MSC-ETZ	0.85	0.00	10.46	0.08	1.68	0.04	0.83	0.02	9.54	0.41	1.63	0.04
Mariana trough	15°–17°N	0.69	0.07	8.04	0.14	2.68	0.12	0.68	0.01	7.01	0.30	2.52	0.08
	17°–19°N	0.69	0.05	7.22	1.08	2.79	0.07	0.66	0.01	6.45	0.23	2.54	0.06
	19°–21°N							0.67	0.01	6.70	0.49	2.39	0.15
North Fiji basin	N160°	0.73	0.00	8.60	0.00	2.30	0.00	0.67	0.03	6.11	0.57	2.53	0.14
	TJ	0.73	0.04	8.59	0.37	2.17	0.02	0.73	0.04	8.59	0.37	2.17	0.02
Woodlark basin	D'Entrecasteaux							0.66	0.00	6.24	0.06	2.73	0.02
	center	0.76	0.00	8.71	0.06	2.32	0.03	0.76	0.00	8.71	0.06	2.32	0.03
	east	0.71	0.02	8.46	0.14	2.77	0.04	0.67	0.03	7.74	0.73	2.58	0.18
						Basin-Scale Averages							
East Scotia ridge	whole	0.72	0.01					0.70	0.01				
Lau basin	north	0.79	0.01	8.98	0.17	2.04	0.04	0.80	0.01	8.73	0.21	2.04	0.05
	south	0.79	0.01	8.76	0.40	1.55	0.04	0.85	0.01	9.91	0.34	1.24	0.06
Manus basin	whole	0.85	0.00	10.46	0.08	1.68	0.04	0.81	0.03	9.32	0.38	1.60	0.04
Mariana trough	whole	0.69	0.04	7.66	0.36	2.72	0.09	0.67	0.01	6.69	0.18	2.50	0.05
Sumisu rift	whole							0.70	0.03	9.27	0.59	2.04	0.02
North Fiji basin	whole	0.74	0.01	8.59	0.33	2.19	0.03	0.70	0.01	7.97	0.49	2.28	0.07
Woodlark basin	whole	0.74	0.01	8.61	0.07	2.49	0.07	0.74	0.03	7.93	0.33	2.49	0.07

the axial profiles of *Taylor and Martinez* [2003] for the four basins they examined or from averaging the sample collection depths for the other basins we have examined in our study (see Table 4). We also estimated crustal thickness (see Table 4) from seismic profiles of the few basins where detailed geophysical data exist [*Bibee et al.*, 1980; *LaTraille and Hussong*, 1980; *Ambos and Hussong*, 1982; *Turner et al.*, 1999; *Crawford et al.*, 2003; *Martinez and Taylor*, 2003].

2.8. Summary of Methods

[27] We have described in detail the several steps in our treatment of glass compositions from back-arc basins, and we summarize them here. The compiled basalt data were first screened to exclude fractionated compositions (MgO < 7.0 wt %) and degassed samples (pure H_2O saturation pressure <30 bars from eruption pressure). All remaining glasses were then corrected for the effects of crystal fractionation, adjusting the melt compositions to olivine-enriched compositions that would be in equilibrium with Fo_{90}. The extent of melting of a peridotitic mantle source (F) required to produce each corrected back-arc basin glass composition was calculated based on its TiO_2 content relative to its mantle source using the batch melting equation and a constant D_{Ti}. The concentration of H_2O in the mantle source of each sample was then calculated using this value of F and a constant D_{H_2O}. Note that the calculations of F require knowledge of the TiO_2 content of the mantle source (C^o_{Ti}) of each sample prior to melting. We have shown that TiO_2 contents of most back-arc mantle sources (prior to addition of slab-derived components) can be modeled using the DMM composition of *Salters and Stracke* [2004], but in some cases the deviation from this end-member appears to be significant. For mantle lower in TiO_2 than DMM, we adjusted C^o_{Ti} for each spreading segment using conservative elements (Nb, Ta, Zr, TiO_2, Y), assuming that the concentrations of these elements deviate from DMM due to a previous episode of melt extraction (assumed to be batch melting). For sources enriched in trace elements relative to DMM (again, prior to addition of slab-derived components), we modeled C^o_{Ti} based on their TiO_2/Y ratios. The total variation in C^o_{Ti} from region to region is a little more than a factor of 2. A Monte Carlo error analysis reveals uncertainties on the linear regression through the modeled Mariana trough data of ~30% on the F-intercept and ~10% on the slope of the line, and the trend of F and H_2O_o determined using this procedure is similar to that first shown in this region by S&N94 (see Figure 9b).

3. Results

[28] Here we present a summary of the modeling outcome for back-arc basins and mid-ocean ridges. Detailed discussions of the results for each specific region are presented in the auxiliary material. The correlation of $TiO_{2(Fo90)}$ and $H_2O_{(Fo90)}$ in the fractionation-corrected glass data illustrate the first-order observations suggesting a relationship between H_2O and mantle melting in back-arc basin settings. Most of the back-arc basin data show dominant trends of decreasing $TiO_{2(Fo90)}$ with increasing $H_2O_{(Fo90)}$ (Figure 4). Because TiO_2 is incompatible during mantle melting, and we assume TiO_2 originates only from the mantle, we expect that its concentration in the melt will be progressively diluted as the melt fraction (F) increases. The relationships shown in Figures 4a, 4c, and 4d thus suggest

Table 4. Physical Variables at Back-Arc Basins

Basin	Segment	Arc Distance, km	1σ	Trench Distance, km	1σ	Axial Depth, m	1σ	C_T, km	Mean F	1σ	Mean H_2O_o, wt %	1σ	T_p, °C	±	dF/dH_2O_o	±	F-Intercept, %	r^2
Small-Scale Averages																		
East Scotia ridge	E2–E4	200	5	361	4	−3275	73		0.10	0.00	0.08	0.01	1343	12				
	E5–E8	217	7	370	6	−3748	53		0.07	0.00	0.02	0.00	1355	7				
	E9	117	3	291	10	−2690	124		0.21	0.11	0.21	0.04	1299	0				
Lau basin	MTJ	158	24	293	18	−2152	58		0.12	0.01	0.07	0.01	1410	8				
	CLSC	268	1	400	0	−2296	11	7–8	0.11	0.00	0.04	0.00	1461	6				
	ILSC	209	8	346	4	−2950	100		0.15	0.03	0.06	0.02						
	ELSC	162	8	313	2	−2541	15	5.5–6.5	0.09	0.00	0.04	0.01						
	VFR	127	1	260	1	−1957	12	8	0.14	0.01	0.22	0.04						
Manus basin	ER	174	3	280	18	−1843	75		0.38		0.38							
	MSC-ETZ	274	8	402	5	−2329	34		0.18	0.02	0.10	0.03						
Mariana trough	15°–17°N	98	2	323	2	−3997	24		0.12	0.01	0.11	0.02	1355	10				
	17°–19°N	107	2	317	1	−4287	28		0.13	0.01	0.13	0.02	1338	8				
	19°–21°N	111	6	331	7	−4347	25		0.12	0.01	0.12	0.03	1415	16				
Woodlark basin	D'Entrecasteaux					−2669	0		0.15	0.02	0.10	0.01						
	center					−3819	85		0.06	0.00	0.01	0.00	1391	3				
	east					−3238	174		0.12	0.04	0.08	0.07	1355	3				
North Fiji basin	N160°	626	30	926	73	−3554	52		0.12	0.03	0.07	0.00	1389	14				
	TJ	621	15	751	20	−2658	260		0.14	0.01	0.03	0.00	1409	11				
Basin-Scale Averages																		
East Scotia ridge	whole	197	6	356	5	−3363	59		0.09	0.01	0.05	0.01	1348	7	0.32	0.04	5.1	0.848
Lau basin	north	207	15	342	13	−2309	9		0.12	0.01	0.05	0.01	1440	7	0.66	0.07	7.9	0.710
	south	136	4	273	6	−2273	19		0.14	0.01	0.22	0.04			0.29	0.02	7.7	0.910
Manus basin	whole	249	17	372	21	−2153	12		0.12	0.04	0.19	0.03	1505	9	0.77	0.05	10.1	0.951
Mariana trough	whole	105	2	321	2	−4222	16	4–7	0.15	0.03	0.19	0.05	1354	10	0.48	0.04	6.9	0.798
Woodlark basin	whole					−3366	127		0.10	0.02	0.05	0.03	1377	5	0.67	0.05	6.7	0.923
Sumisu rift	whole	16	2	229	3	−1911	123		0.18	0.01	0.16	0.02						
North Fiji basin	whole	623	13	817	42	−2994	227		0.13	0.01	0.05	0.01	1406	9				

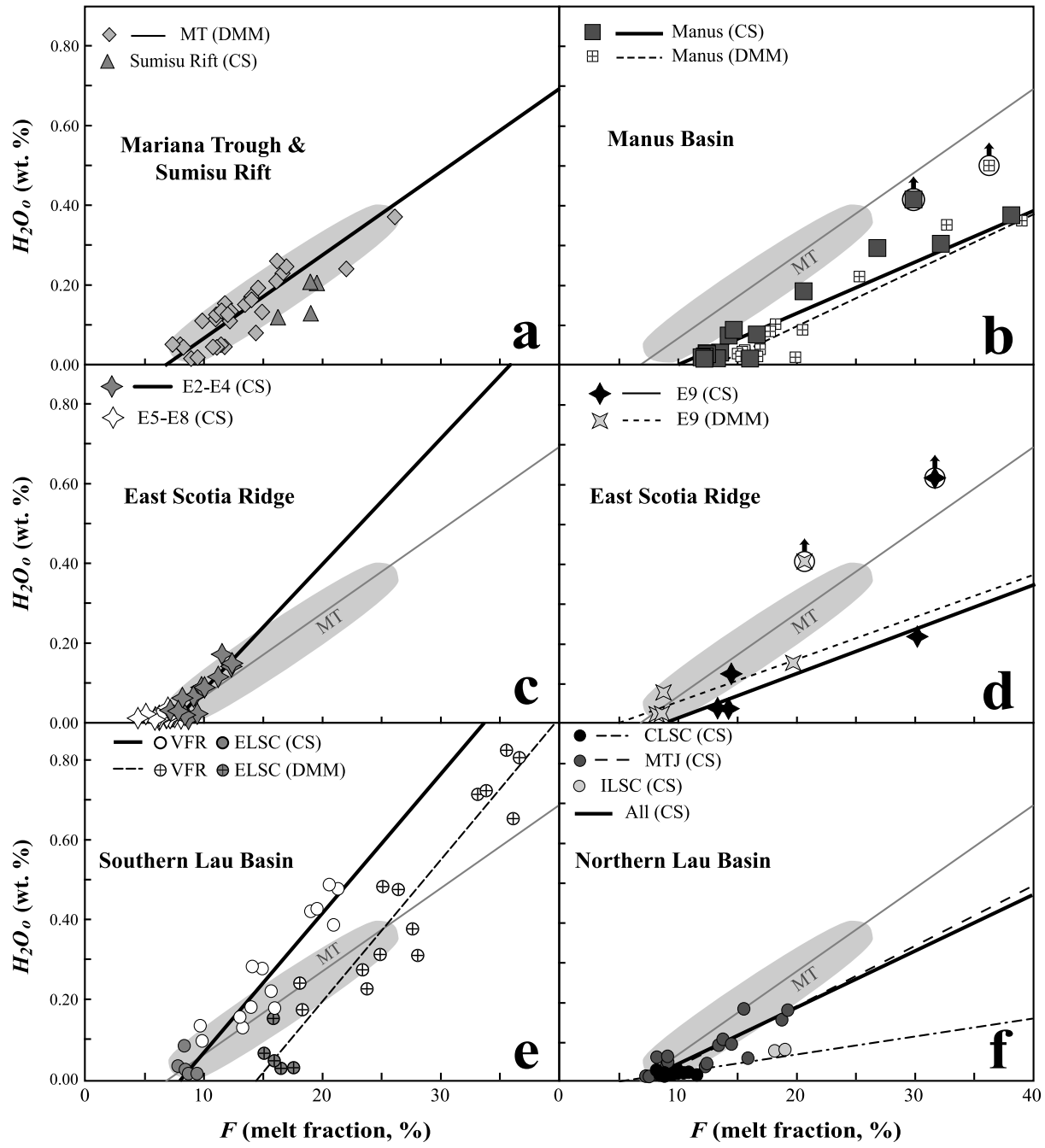

Figure 10. H_2O_o versus F at back-arc basins closely associated with subduction. The light shaded field and the dark shaded line are the range of Mariana trough model points and the linear regression from Figure 9b. The results using a constant MORB mantle source ($C^o_{Ti} = 0.133$) are labeled DMM, and results using variable mantle source constraints modeled in this work are labeled CS (corrected source; C^o_{Ti} given in Table 1). In many cases, the DMM and CS results were too similar to distinguish, and in these cases, only CS results are shown. (a) Mariana trough (CS, $y = 2.082x - 0.142$; $r^2 = 0.798$) and the Sumisu rift (CS, $y = 3.152x - 0.414$; $r^2 = 0.448$). (b) Manus basin (CS, $y = 1.299x - 0.131$; $r^2 = 0.951$). (c) East Scotia ridge segments E2–E4 (CS, $y = 3.171x - 0.238$; $r^2 = 0.796$) and E5–E8. (d) East Scotia ridge segment E9 (CS, $y = 1.106x - 0.098$; $r^2 = 0.831$). (e) Southern Lau basin Valu Fa ridge (VFR) and the eastern Lau spreading center (ELSC) segments (CS, $y = 3.450x - 0.265$; $r^2 = 0.910$); note the difference between DMM and CS models in this region. (f) Northern Lau basin central Lau spreading center (CLSC), intermediate Lau spreading center (ILSC), and Mangatolu triple junction (MTJ) segments (CS, $y = 1.524x - 0.119$; $r^2 = 0.710$). Note one ESR-E9 point and one Manus point highlighted with dark circles and arrows, which were excluded from modeling because they likely degassed. Arrows indicate the direction these points would shift had they not degassed. Data sources are given in the text, and details for each region are given in the auxiliary material.

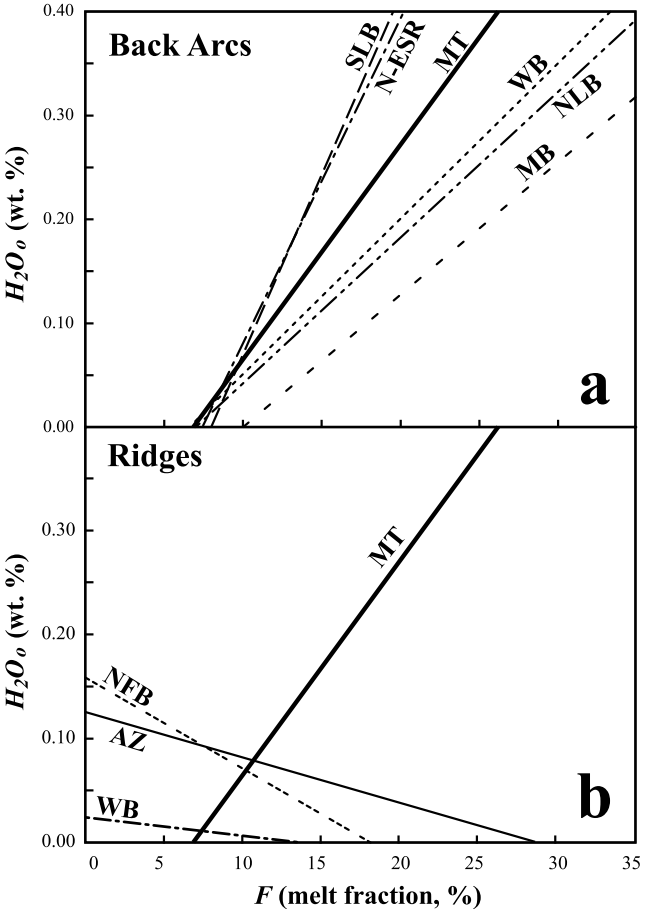

Figure 11. Compilation of H_2O_o versus F best fits for mid-ocean ridges and back-arc basins. (a) Results for back-arc basins, showing trends with positive, variable slopes at East Scotia ridge segments E2–E4 (N-ESR), the southern Lau basin (SLB), the Woodlark basin (WB), the Sumisu rift (SR), the Manus basin (MB), and the northern Lau basin (NLB). (b) Results for mid-ocean ridges, showing trends with negative slopes at the North Fiji basin (NFB), wet samples from the Azores platform region of the Mid-Atlantic Ridge (AZ), and dry samples from the Woodlark basin (WB). The Mariana trough (MT) from Figure 11a is shown for reference. In both panels, all lines shown are for regressions through CS results (as distinguished in Figures 10 and 12).

that increasing F correlates with increasing H_2O. Using the fractionation corrections and melting model developed here, we find that the maximum H_2O concentrations in both the lavas and the mantle at back-arc basins far exceed those at mid-ocean ridges, and the principal relationship between H_2O_o and F is positive (Figure 10). Mantle H_2O concentrations at back arcs range from MORB-like values (0.01–0.05 wt %), which are present at all basins, to maxima of 0.4–0.5 wt % H_2O at the Manus basin and the Valu Fa ridge. Positive trends in H_2O_o versus F vary regionally in slope and intercept (Table 4), with the East Scotia ridge segments E2–E4 and the southern Lau basin defining the steepest trends and the Manus and northern Lau basins the shallowest (Figure 11a). The calculated F is <20% in most samples from all back-arc basins, with only a few extreme samples indicating $F > 20\%$. Correcting for mantle source depletion has the largest effect for the Valu Fa ridge, where the mantle is the most depleted relative to DMM (Figure 10e). All arrays defined by back-arc basins indicate finite melt fraction at zero (or low) H_2O_o, which strongly implicates decompression melting if we accept that melting in the absence of a flux agent like H_2O must be generated by adiabatic decompression. However, the important point in our view is the robust and recurring positive relationship between H_2O and extent of melting at back-arc basins.

[29] The negatively sloping relationship between H_2O_o and F proposed for mid-ocean ridges by *Asimow and Langmuir* [2003] and *Cushman et al.* [2004] is apparent at four regions (Figure 12): two mid-ocean ridges (GSC, Azores platform), and two spreading centers with distant or past associations with subduction (N. Fiji basin, Woodlark basin). In these regions, melt fraction broadly decreases as a function of increasing H_2O_o, although these relationships are much weaker than for back arcs. Maximum H_2O_o concentrations reached at these places are low (max. 0.08 wt % H_2O_o (Figure 12)) compared to active back-arc basins, and are well within the storage capacity of H_2O in nominally anhydrous upper mantle phases such as olivine, orthopyroxene, and clinopyroxene (e.g., 0.12–0.38 wt % H_2O at 3 GPa [*Hirschmann et al.*, 2005]). The negatively sloped trends in H_2O_o versus F observed at classic mid-ocean ridges (Azores, GSC) are in regions of ridge–hot spot interaction, where constraints on mantle source variation are essential for successful modeling (and, indeed, TiO_2 and H_2O generally vary with isotope ratios). We are not fully satisfied with any of the approaches for quantifying chemical source variations of the mantle in these regions (including the approach based on TiO_2/Y ratios presented here), and consequently the significance of these negative slopes is uncertain. We conclude, however, that such negative slopes are indeed real phenomena, given their persistence at two of three ridges regardless of which approach is taken to constrain variations in source composition and the appearance of such a trend at part of the Woodlark basin. This basic difference in the character of the trend of H_2O_o versus F at mid-ocean ridges versus back-arc basins suggests to us a fundamental difference in the role of H_2O in the melting processes beneath back arcs and mid-ocean ridges.

4. Discussion

4.1. Effects of Variable Mantle Temperature

[30] Constraining mantle temperature variations beneath back-arc basins is important for understanding subduction zone petrogenesis, structure, and evolution. The 150°C temperature range calculated here (i.e., 1350°–1500°C (Figure 2, Table 4)), using only the driest back-arc basalts, is within the range of global MORBs, which sample upper mantle that spans a ~250°C range in T_p [*Klein and Langmuir*, 1987; *McKenzie and Bickle*, 1988; *Langmuir et al.*, 1992]. Back arcs thus appear to sample mantle of similar temperature to mid-ocean ridges, reflecting normal thermal variations in the upper mantle. The high-T end of the MORB range, however, is due largely to hot spot proximity (e.g., the Mid-Atlantic Ridge in the vicinity of

Iceland). Such a hot spot influence at back-arc basins seems unlikely since the slab would generally shield the overlying mantle wedge from the thermal effects of deep upwelling plumes. At the northern segments of the Lau basin, however, infiltration of the Samoan mantle plume through a tear in the subducting plate has been proposed to account for along-strike geochemical variations [*Poreda and Craig*, 1992] and this could have thermal consequences as well. Among the back-arc basins considered here, the Manus basin taps mantle of the highest T_p; geochemical evidence such as high $^3He/^4He$ ratios [*Shaw et al.*, 2004] is also consistent with a hot spot contribution to this elevated T_p. The mantle temperatures calculated by us for back-arc basins also correlate with variations in the *S* wave velocity structure of mantle beneath four back-arc basins at 40–100 km depth [*Wiens et al.*, 2006]. Specifically, the *S* wave velocities decrease with increasing temperature, with the Mariana trough as the coldest/fastest and the northern Lau basin as the hottest/slowest, strongly supporting our determination of significant thermal variations in these settings.

[31] Although back-arc basin basalts thus appear similar to MORBs with respect to mantle T_p, our results demonstrate they are quite different in that they have much higher H_2O_o concentrations (approaching 0.5 wt % in some cases) and define positive, linear trends in H_2O_o versus *F*. Moreover, the variations in slope correlate with the T_p variations described above based on independent measures (Na_2O and FeO). The relationship between slope (dF/dH_2O_o, an expression of wet melt productivity) and temperature (Figure 13) is not an unexpected result, since isothermal-isobaric models predict this effect [S&N94; *Hirose and Kawamoto*, 1995; G&G98; *Hirschmann et al.*, 1999; *Reiners et al.*, 2000; *Katz et al.*, 2003]. On the basis of our results, melt productivities in the sources of back-arc basin magmas increase from ~30 to 80%/wt % (i.e., % *F* per wt % H_2O_o) as T_p increases by 150°C; this result is similar to those of MELTS calculations, which predict wet

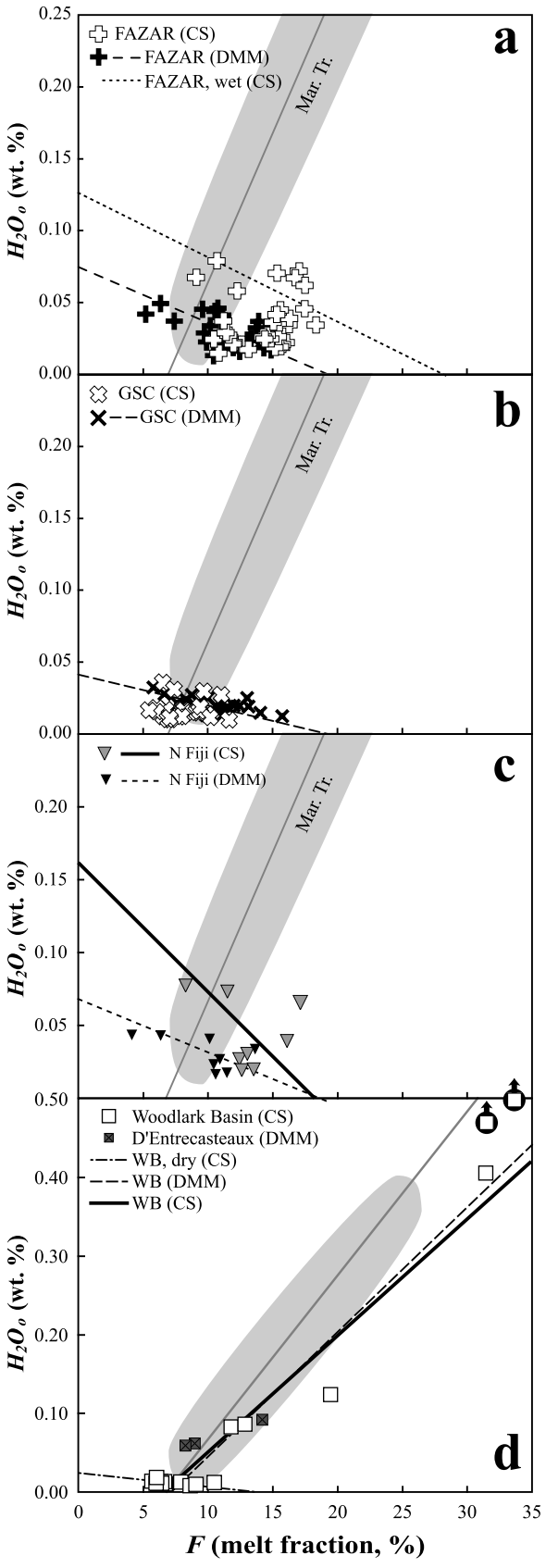

Figure 12. H_2O_o versus *F* at mid-ocean ridges and spreading centers distant from subduction zones. The shaded region shows part or all of the range of Mariana trough data from Figure 9b, and the line through the shaded region shows the best fit line. As in Figure 10, model results using a constant MORB mantle source are labeled DMM (C^o_{Ti} = 0.133); results using variable mantle source constraints modeled in this work are labeled CS (corrected source; C^o_{Ti} given in Table 1). (a) Mid-Atlantic Ridge and Azores platform (FAZAR) (DMM, $y = -0.390x + 0.075$; $r^2 = 0.229$); the CS results here indicate a negatively sloped trend for the 10 wet samples from the Azores platform (wet CS, $y = -0.447x + 0.128$; $r^2 = 0.0823$) that may be distinct from the rest of the ridge, which is comparatively dry and alone does not produce a resolvable trend. (b) Galápagos spreading center (GSC) (DMM, $y = -0.214x + 0.040$; $r^2 = 0.284$); note that the CS results do not produce a resolvable trend. (c) North Fiji basin (CS, $y = -0.886x + 0.163$; $r^2 = 0.069$). (d) Woodlark basin (WB), which shows two distinct, oppositely sloping trends in the wet (all CS, $y = 1.482x - 0.097$; $r^2 = 0.923$) and dry samples (dry CS, $y = -0.173x + 0.025$; $r^2 = 0.106$). Note two WB model points highlighted with dark circles and arrows, which were excluded from modeling because they likely degassed some H_2O. Arrows indicate the direction these points would shift had they not degassed. Data sources are given in the text, and details for each region are given in the auxiliary material.

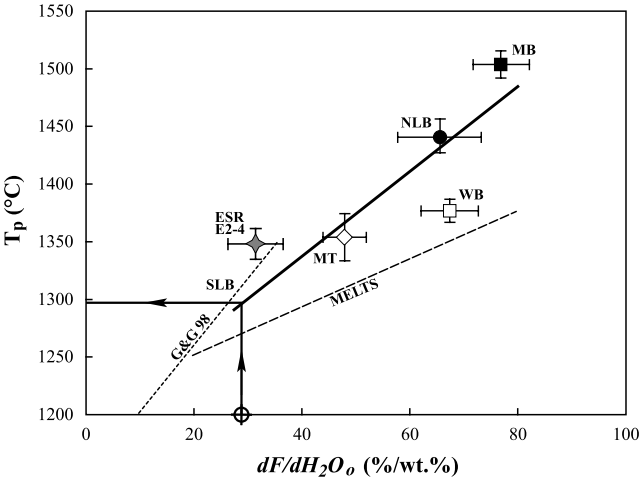

Figure 13. Mantle potential temperature (T_p) from Figure 2 versus wet melt productivity (dF/dH_2O_o) from Figure 11 at back-arc basins (values given in Table 4). T_p is constrained from dry glasses from each basin (<0.5 wt % H_2O), whereas dF/dH_2O_o is constrained using wet and dry samples as shown in Figures 10–12. Regions shown are the large-scale basin averages for the Manus basin (MB; solid square), the northern Lau basin (NLB; solid circle), the Woodlark basin (WB; open square), the Mariana trough (MT; open diamond), and the East Scotia ridge (ESR; shaded star) segments E2–E4. Error bars for dF/dH_2O_o are the standard error on the linear regression, and error bars for T_p are 1σ deviations from the regional, basin-scale averages (Table 4). The solid black line is a linear regression through these points ($y = 368.6x + 1191$; $r^2 = 0.66$). The southern Lau basin (SLB) is shown as a crossed circle on the x axis because this region illustrates how measured wet melt productivity may be used to constrain T_p in places where T_p cannot be independently determined through Na-Fe systematics. The long-dashed line is the range of wet melt productivities from *Hirschmann et al.* [1999] (MELTS), and the short-dashed line is the trend of wet melt productivities from G&G98.

melt productivities of 20–80%/wt % over a 125°C temperature range, although offset to lower absolute temperatures and pressure (1250°–1375°C at 1 GPa [*Hirschmann et al.*, 1999]; see Figure 13). The experimental results of G&G98 indicate melt productivities generally lower than those we have inferred (10–36%/wt % at 1200°–1350°C and 1.5 GPa; see Figure 13) but they overlap with our empirical trend at ~1300°C.

[32] The correlation between dF/dH_2O_o and T_p has the potential to provide an alternative approach to estimating mantle T_p; this would be particularly useful for regions with H_2O concentrations too high to permit successful modeling of Na and Fe variations, in regions at which variations in source composition lead to deviations from the global MORB trend in Fe-Na, and for volcanic arc fronts [*Kelley et al.*, 2003]. At the Valu Fa ridge (VFR), for example, water contents of erupted glasses and melt inclusions do not overlap with MORB, and the relationship between Na and Fe in these samples is disrupted by uniformly high H_2O. Samples from the VFR do, however, form a steep linear trend in H_2O_o versus F (Figure 12e), indicating less efficient melt productivity, which is consistent with cooler mantle T_p beneath the VFR than the northern Lau basin (Figure 13) and suggests an along-strike thermal gradient in the Lau basin. In principle, the F-intercept of the trend could also provide constraints on mantle T_p, but high errors on the intercept relative to the slope make dF/dH_2O_o a more robust indicator.

[33] Additional factors related to mantle temperature beneath back-arc basins include cooling from the subducting plate, convergence rate, and back-arc spreading rate [e.g., *Peacock*, 1996; *Kincaid and Sacks*, 1997; *Kincaid and Hall*, 2003; *England and Wilkins*, 2004], and the calculated mantle temperatures for various back-arc basins based on our results can be used to evaluate the effects of these various factors on T_p. For example, in the Lau basin, distance from the spreading axis to the trench (and slab) decreases toward the south (Figure 14), and this may be partially responsible for cooler T_p at the VFR relative to the northern basin. Likewise, convergence rate and back-arc spreading rate are thought to be linked, since fast subduc-

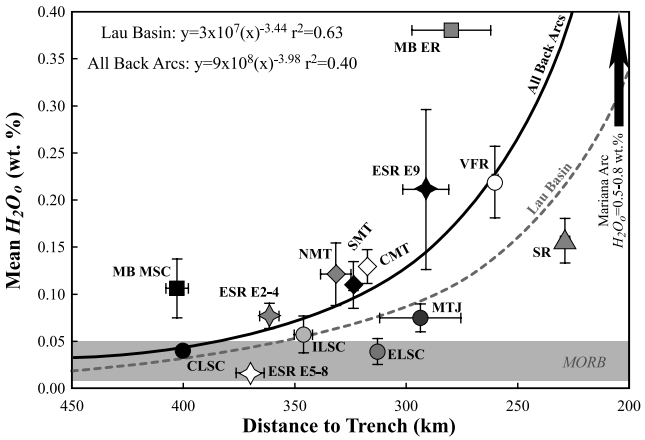

Figure 14. Mean H_2O_o versus distance to the trench at back-arc basins (values given in Table 4). The back-arc basin data are regional averages of the Manus basin Eastern Rifts (MB ER; shaded square) and the Manus spreading center/eastern transform zone (MB MSC; solid square), the Lau basin central Lau spreading center (CLSC; solid circle), the intermediate Lau spreading center (ILSC; light shaded circle), the Mangatolu triple junction (MTJ; dark shaded circle), the eastern Lau spreading center (ELSC; medium shaded circle) and the Valu Fa ridge (VFR; open circle), the East Scotia ridge (ESR) segments E2–E4 (shaded star), E5–E8 (open star), and E9 (solid star), and the Mariana trough northern third (NMT, 19°N–21°N; shaded diamond), central third (CMT, 17°N–19°N, open diamond) and southern third (SMT 15°N–17°N, solid diamond). Error bars are 1σ deviations from the small-scale averages (Table 4). The shaded field is the range of H_2O_o in MORB from this study (Figure 12). The black arrow in the upper right indicates the direction that volcanic arcs are predicted to plot (off-scale). The range for the Mariana arc is from *Kelley et al.* [2003]. The dashed shaded curve is a power law regression through the five Lau basin segment averages. The solid black curve is a power law regression through all of the back-arc segments shown.

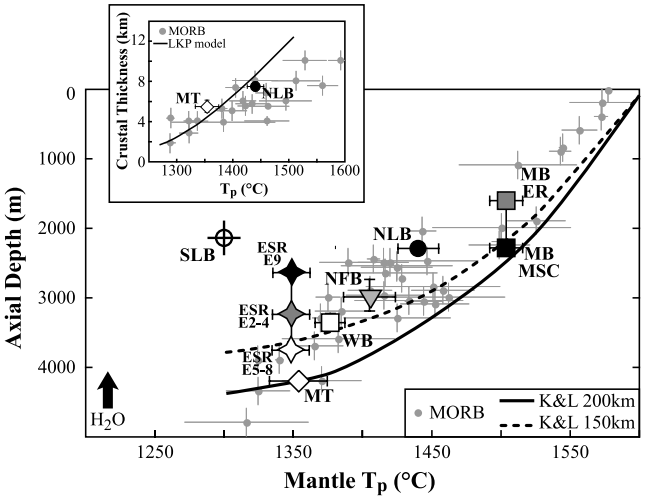

Figure 15. Mantle potential temperature (T_p) versus axial depth at back-arc basins. Shaded circles are MORB from *Langmuir et al.* [1992], with T_p as calculated here (equations (12) and (13) in text; see also Figure 2). The back-arc basin data are regional averages of the Manus basin Eastern rift (MB ER; shaded square) and Manus spreading center (MB MSC; black square) segments, the northern Lau basin (NLB, black circle), the North Fiji basin (NFB, inverted shaded triangle), the Woodlark basin (WB, open square), the East Scotia ridge segments E2–E4 (ESR E2–4, shaded star), E5–E8 (ESR E5–8, open star) and E9 (ESR E9, black star), and the Mariana trough (open diamond). Error bars are 1σ deviations from the basin-scale or small-scale averages (Table 4). The southern Lau basin (SLB) is shown as a crossed circle at the T_p indicated by Figure 13. The solid curve is the model from *Klein and Langmuir* [1987] for 200 km compensation depth; the dashed curve is for 150 km compensation depth. Inset shows T_p versus crustal thickness; the solid line is the model curve from *Langmuir et al.* [1992]. The black arrow in the lower left indicates the direction that H_2O addition to the mantle should drive axial depth and crustal thickness. Extra melting from H_2O should create more crust and shallower axial depth than a dry ridge at the same T_p.

tion is expected to drive more rapid wedge circulation and back-arc mantle upwelling, and the fast back-arc spreading will in turn lead to increased relative convergence between the overriding and subducting plates. Our results are consistent with these expectations: convergence rates are fast at the Tonga/Lau and New Britain/Manus (>140 mm/yr) subduction zones, where back-arc temperatures are warmer, and slow at the Mariana (40–70 mm/yr) and South Sandwich/East Scotia (67–79 mm/yr) subduction zones [*Taylor and Martinez*, 2003, and references therein], where back-arc T_p is cooler.

[34] An important test of the wet melt productivities calculated here is the effect on the volume of crust created, as recorded in crustal thickness or axial depth (e.g., Figure 15). Extents of melting at back-arc spreading centers are driven higher than MORB by the addition of H_2O, and high extents of melting should result in thick crust and shallow axial depth relative to mid-ocean ridges. Axial depth at back arcs varies from ~1900 to 4300 meters below sea level, within the range of mid-ocean ridges. Geochemical indicators of extent of mantle melting, such as $Na_{8.0}$, broadly correlate with axial depth at back arcs, as at ridges (Figure 1 [see also *Klein and Langmuir*, 1987]), although regions with shallow axial depth and high H_2O_o (e.g., the southern Lau basin and the Manus basin (Figure 1b, open symbols)) tend to have lower mean $Na_{8.0}$ than ridges of the same depth. In comparison to the driest melts from each region (H_2O < 0.5 wt % (Figure 1b, solid symbols)), however, average back-arc melts (Figure 1b, open symbols) are skewed to low $Na_{8.0}$ (esp. basins where mean H_2O_o is high; Mariana, Manus, southern Lau basins), consistent with high H_2O driving higher F and lower $Na_{8.0}$ (Table 3, Figure 1b).

[35] We may also compare back-arc mantle T_p and axial depth to that of ridges, which show a positive relationship owing to the greater thermal buoyancy, compositional buoyancy, and crustal thickness of hotter regions (Figure 15 [see also *Klein and Langmuir*, 1987]). We find that most back-arc spreading axes are at water depths consistent with their mantle temperatures, suggesting that the excess melt produced by high H_2O leads to small shifts in axial depth, generally within the range of ridges. For the Mariana trough and the northern Lau basin, the two basins with constraints on both crustal thickness and mantle T_p, we also find that back-arc basins are within the range of normal mid-ocean ridges, albeit plotting toward the high crustal thickness side of the ridge array (Figure 15, inset). Mantle temperature thus appears to be the dominant physical variable contributing to melt generation beneath many back-arc basins. On the other hand, clear excess melt production is present for the most water-rich back-arc segments, where anomalously shallow axial depth results from the excess melt produced by H_2O addition (see SLB, NLB, ESR E2-4/E9 and MB ER points in Figure 15). Thus water clearly contributes excess melt production in the wettest regions, whereas lavas from other regions may only reflect high extents of melting of small mantle source volumes and preferentially erupt wet melts.

4.2. Effects of Variable Mantle H_2O Concentration

[36] The H_2O_o concentrations at mid-ocean ridges, with few exceptions, do not exceed 0.01–0.05 wt % (max. H_2O_o = 0.08 wt % at the Azores platform), which is well below the minimum storage capacity of H_2O in nominally anhydrous mantle phases at pressures near the onset of melting (i.e., 0.12 wt % H_2O at 3 GPa [*Hirschmann et al.*, 2005]). On the basis of our results, however, the sources of back-arc basin basalts have H_2O_o concentrations of 0.01–0.50 wt % (and possibly higher when considering degassed samples (Figure 10)). In many regions, these concentrations clearly locally exceed the minimum storage capacity of nominally anhydrous mantle minerals (i.e., 0.18 wt % H_2O at 6 GPa [*Hirschmann et al.*, 2005]). We do note that *Hirschmann et al.* [2005] report a large range of permissible mantle H_2O storage capacities (0.18–0.63 wt % at 6 GPa) and their preferred model indicates a storage capacity of 0.44 wt% H_2O at 6 GPa, which would only be exceeded by the wettest local regions indicated by our study. We emphasize, however, that many regions locally exceed the minimum storage capacity, and back-arc basin mantle

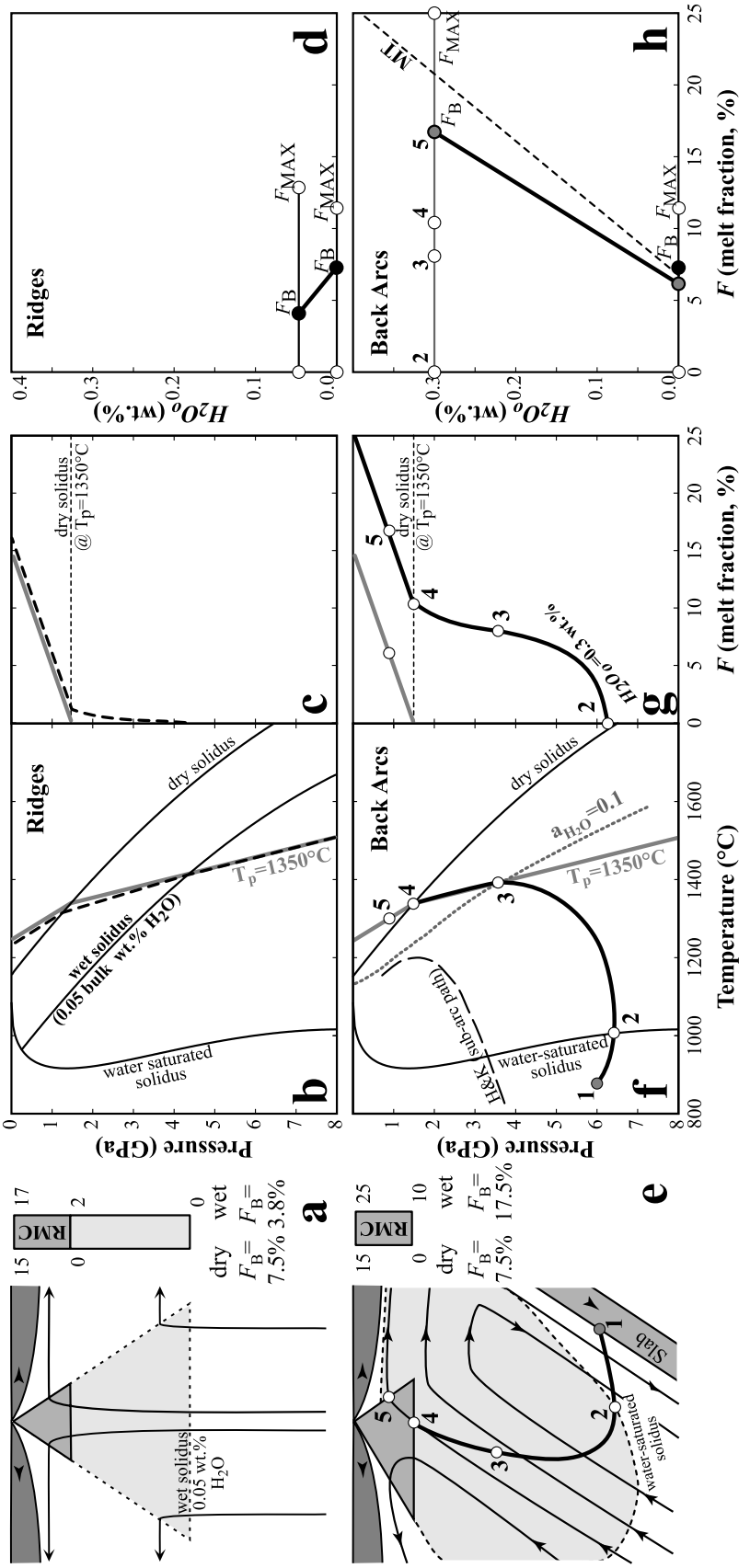
a
RMC
wet solidus
0.05 wt.%
H2O
dry wet
FB= FB=
7.5% 3.8%
b
Ridges
dry solidus
wet solidus
(0.05 bulk wt.% H2O)
Tp=1350°C
water saturated
solidus
Pressure (GPa)
c
dry solidus
@ Tp=1350°C
d
Ridges
H2Oo (wt.%)
FB
FMAX
e
RMC
dry wet
FB= FB=
7.5% 17.5%
Slab
water-saturated
solidus
f
Back Arcs
dry solidus
aH2O=0.1
Tp=1350°C
H&K
sub-arc path
water-saturated
solidus
Temperature (°C)
g
dry solidus
@ Tp=1350°C
H2Oo=0.3 wt.%
F (melt fraction, %)
h
Back Arcs
MIT
FB
FMAX
H2Oo (wt.%)
F (melt fraction, %)

Figure 16

sources may thus contain some H_2O in excess of the nominally anhydrous minerals. As shown in Figure 14, water concentrations in back-arc mantle tend to increase as spreading axes approach the trench. The wettest spreading segments (e.g., Valu Fa, Manus ER, East Scotia E9, Sumisu) are located ~300 km or less from the trench. Figure 14 shows some back-arc mantle H_2O in excess of average MORB up to 400 km from the trench, but the average H_2O_o concentrations increase rapidly (fit by a power law) as trench distance decreases below ~300 km. There is also evidence of even higher H_2O_o values beneath the Mariana arc (Figure 14 [*Kelley et al.*, 2003]), suggesting that the trenchward increase in H_2O_o defined by back-arc basins may extend to the mantle beneath arcs. At the other extreme, the North Fiji basin has an average H_2O_o similar to MORB (0.05 wt % (Table 4)) and is ~800 km east of the New Hebrides/Vanuatu trench. The distance to the trench in this case is consistent with the low mantle H_2O concentrations, and is clearly too far for this ridge to incorporate direct H_2O additions from the slab. These relationships strongly suggest that the subducting slab is the ultimate source of elevated H_2O at back arcs, as well as at arcs, and provide observational constraints on how H_2O may be lost from the slab and transported through the mantle wedge.

4.3. Different Roles of H_2O at Ridges and Back-Arc Basins

[37] On the basis of our results, there appears to be a distinction between basalts from mid-ocean ridges (and spreading centers distant from subduction zones), where increasing H_2O is accompanied by decreasing extents of melting (i.e., $dF/dH_2O_o < 0$), and basalts from back-arc basins, where high H_2O is associated with high extents of melting (i.e., $dF/dH_2O_o > 0$). How can we explain the different behavior of ridges and back-arc basins during wet melting? Although there are many possibilities, we focus here on the hypothesis that the contrasting trends relate to the relative concentrations of H_2O in the two environments. At ridges, water is present in low concentrations, likely dissolved in solid, nominally anhydrous phases such as olivine and pyroxene prior to melting. As such, the shape of the melting regime, and the production of melts and their mixing proportions, are governed to first order by decompression paths along the solid flow field [*Langmuir et al.*, 1992]. At back arcs, on the other hand, H_2O added from the subducting plate will, according to our results, at least locally exceed the nominal storage capacity of the mantle. In such cases, excess H_2O is supplied to the mantle melting region via fluids or melts, driving high degrees of melting in a process that may be independent of the solid flow field of the mantle. In this final section, we develop these ideas further, and explore some of the consequences of this hypothesis. Figures 16 and 17 portray the different processes that contribute to melting beneath mid-ocean ridges and back-arc basins. Note that these scaled cartoons are designed to be more illustrative than exclusive.

4.3.1. A Back-Arc Melting Model

[38] In the mantle wedge above subduction zones, *P-T* conditions experienced by ascending H_2O, melt and mantle are unlikely to be adiabatic in the classic sense, in contrast to the conditions beneath a ridge (Figure 16). In particular, the thermal gradient above the slab is influenced by heat flow from the mantle wedge into the cold slab, so as water released from the slab (and any subsequent melt) ascends to lower pressure, its temperature increases as it migrates away from the cold slab and into the warm interior of the mantle

Figure 16. Contrasting *P*, *T*, *F*, and H_2O_o relationships for mid-ocean ridges and back-arc basins. (a) Scaled cartoon and residual mantle column (RMC). (b) Pressure versus temperature phase diagram and (c) pressure versus *F*, illustrating the melting mechanism responsible for (d) decreasing bulk *F* with increasing H_2O_o at mid-ocean ridges (after *Asimow and Langmuir* [2003]). In Figure 16b the dry solidus is from *Langmuir et al.* [1992]; the wet solidi are from *Katz et al.* [2003]; the heavy, solid shaded line is an adiabatic path for anhydrous mantle of T_p = 1350°C; the long-dashed black line is an adiabatic path for hydrous mantle (H_2O_o = 0.05 wt %). In Figure 16c the solid shaded line is *F* along the anhydrous, adiabatic path, and the long-dashed black line is *F* along the hydrous, adiabatic path, showing how mantle intersects the solidus deeper in the wet case than the dry case, but the wet mantle experiences only low extents of melting before reaching the dry solidus. In Figure 16d the integrated mean *F* over the full pressure range of melting (F_B) in the wet case is less than in the dry case despite reaching a higher maximum *F*. (e) Scaled cartoon and residual mantle column (RMC), (f) pressure versus temperature phase diagram, and (g) pressure versus *F*, illustrating the melting mechanism responsible for (h) increasing bulk *F* with increasing H_2O_o beneath back-arc basins. In Figure 16e the bold line illustrates one possible path that H_2O may follow from the slab through the mantle, with numbered points corresponding to the points in Figures 16f–16h. Water leaves the slab and enters the mantle at point (1), wet melting initiates at point (2), the ascending melt ± mantle achieve an adiabatic path at point (3), dry melting initiates at point (4), and melting ceases at point (5). The mantle flow lines and the asymmetric melting triangle in Figure 16e are based on the back-arc mantle flow geometry of *Conder et al.* [2002]. In Figure 16f the phase boundaries and the dry adiabat are identical to Figure 16b, and the bold line is the path of H_2O through the mantle wedge as in Figure 16e, increasing in temperature with decreasing pressure until intersecting and following the adiabatic path. The dotted shaded line is the schematically located H_2O activity (a_{H2O}) of 0.1 (after *Hirth and Kohlstedt* [1996]), and the long-dashed black curve (Figure 16f, labeled H&K) is the path of intermediate-rate, networked flow of hydrated mantle ascending beneath an arc, from *Hall and Kincaid* [2001]. In Figure 16g the solid shaded line is *F* along the anhydrous, adiabatic path as in Figure 16c, and the bold numbered curve shows possible melt production from H_2O addition along the path in Figures 16e–16f for a H_2O_o concentration of 0.3 wt %. High wet melt productivities are achieved along the path of increasing *T* (see text). In Figure 16h, melt fractions achieved by each numbered point are shown along a line of constant H_2O_o (0.3 wt %), illustrating how H_2O addition drives bulk *F* (F_B) and maximum *F* (F_{MAX}) higher than in the dry case. The series of F_B points corresponding to H_2O_o from 0 to 0.3 wt % (not shown) would produce a positive H_2O_o-*F* trend similar to the Mariana trough (MT, dashed line).

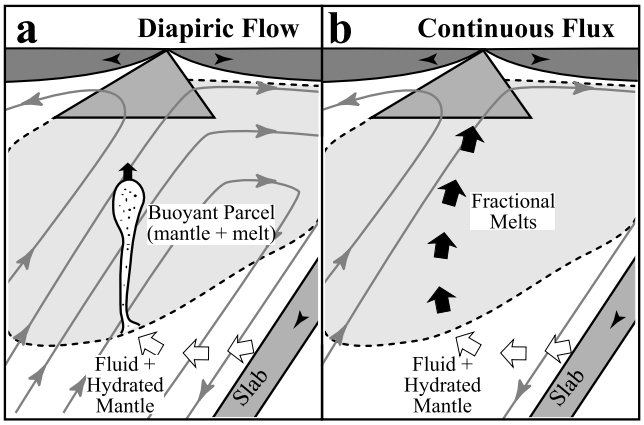

Figure 17. Cartoon illustrating possible melting mechanisms beneath back-arc basins. (a) Diapiric flow of buoyant mantle through the wedge [*Davies and Stevenson*, 1992; *Hall and Kincaid*, 2001; *Gerya and Yuen*, 2003]. Mantle and melt ascend together as a buoyant parcel above the wet solidus. (b) Continuous fluxing of H_2O from the slab through the mantle wedge. Fractional melts form at the wet solidus, rapidly separate from the mantle source, and ascend through the dry melting regime. If solid mantle flow is slow compared to the rate of H_2O flux, continuous addition of H_2O from the slab will progressively melt a single mantle volume.

wedge (Figure 16e, point 1 [*Baker et al.*, 1994; *Hall and Kincaid*, 2001; *Grove et al.*, 2002]). Melting under such conditions may also be described as adiabatic if heat is conserved within the system and the ascending material cools as it ascends. However, if the volumes of fluid and melt are small relative to the heat reservoir of the mantle wedge, the impact of H_2O transport and melting on the heat budget of the system may be negligible and the early stages of melting may thus be nonadiabatic. Melting initiates once the mobilized H_2O crosses the H_2O-saturated solidus in the mantle wedge (Figure 16e, point 2). Under the conditions proposed here near the initiation of melting (high pressure and a free H_2O-rich fluid phase), the fluid in the H_2O-silicate system could instead be supercritical, in which case the concept of H_2O-saturated melting would not apply at the earliest stages of melting [*Shen and Keppler*, 1997; *Mibe et al.*, 2004]. Constraints on the supercritical behavior of basaltic melts and aqueous fluids, however, are still in their infancy and we therefore elect to describe the initiation of melting relative to the H_2O-saturated solidus even at these high pressures. The nature of the melt migration process through the mantle wedge is not well known. Reactive porous flow of the H_2O-rich melt through the mantle is the most likely mechanism [*Davies and Stevenson*, 1992; G&G98], but melt and mantle could ascend together as a buoyantly unstable parcel that upwells beneath the arc and back arc (Figure 17a [*Davies and Stevenson*, 1992; *Hall and Kincaid*, 2001; *Gerya and Yuen*, 2003]). The buoyant parcel may follow a path similar to that for diapiric networked flow described by *Hall and Kincaid* [2001], which exchanges heat as it traverses the hot interior of the mantle wedge. Eventually, the parcel becomes entrained in the adiabatically upwelling paths beneath the back arc (Figure 16e, point 3) that cross the dry solidus (Figure 16e, point 4) and melt further due to adiabatic decompression.

[39] Other reasonable physical models involve separation of melt and solid flow. If melting is a fractional process, the melt and the mantle will not ascend together as a single parcel. Melt will instead become separated from its mantle source, drawing H_2O with it and preventing further H_2O-fluxed melting of the mantle source from taking place. In this case, however, if the mantle is resupplied and remelted by a continuous flux of H_2O from the slab, then the final pooled melt will reflect the integrated, fractional melting of a single mantle source (Figure 17b). This fractional process would produce melt compositions similar to the final melt of the batch process described above [*Plank and Langmuir*, 1992]. In this case, the continuous flux of H_2O and melt through a single mantle volume could imitate melting under isobaric, isothermal conditions, provided the process is nonadiabatic as described above, since each increment of the H_2O/melt flux experiences a similar set of conditions along a path through the mantle wedge. Reactive transport of melt through the mantle wedge [e.g., *Kelemen et al.*, 1997] might also create the appropriate H_2O-TiO_2 relationships, but quantitative models require better constraints on melt versus solid flow vectors in the mantle wedge.

[40] The resulting melt fraction, as recorded by the TiO_2 concentration of the ascending melt, along the path through the mantle wedge (as illustrated in Figure 16) may be estimated with some simple approximations, borrowing from the melting models developed by *Hirth and Kohlstedt* [1996] and *Katz et al.* [2003]. Consider mantle with 0.3 wt % H_2O_o (near the high end of the back arc spectrum). This water is supplied from the subducting plate, and drives melting in the mantle at the water-saturated solidus (Figures 16e–16f, point 2). The initial H_2O concentration of this melt is high (e.g., ~30 wt % H_2O [*Grove et al.*, 2002]), but the melt H_2O concentration is continually diluted as melting proceeds. When the parcel reaches pt. 3 in Figure 16e, the activity of H_2O in the melt (a_{H2O}) is 0.1, or 10% of saturation (note that the activity curve is schematically located in Figure 16e [*Hirth and Kohlstedt*, 1996]), and so the H_2O concentration of the melt will now be ~3.5 wt % (H_2O sat. = 35% at 4.3 GPa [*Katz et al.*, 2003]). Since melting began in a mantle with H_2O_o = 0.3 wt %, we use equation (10) to determine that the melt at pt. 3 reflects ~8% melting of this mantle (in contrast to a ridge case where, for H_2O_o = 0.05 wt %, the amount of melt at this same point reflects only F = 0.04%). As the melting path becomes vertical and mantle parcels ascend adiabatically, melting continues as convective adiabatic decompression melting, with the bulk composition of the source fixed such that H_2O concentration (and activity) in the partial melt decrease progressively. Melt productivity in the ascending mantle will continually decrease because the H_2O content of the melt is diluted as melting proceeds, moving the effective solidus to progressively higher temperatures. In any case, we assume that melt generation above point 4 occurs at a rate similar to dry systems (~1% melt per kilobar of ascent), attaining maximum, final melt fractions in excess of 20% [see also *Gaetani and Grove*, 2003]. The melting path thus reflects a combination of wet, nonadiabatic and wet, adiabatic melting (Figure 16e).

4.3.2. Breakdown of the Melting Regime "Shape" Control

[41] The driest back-arc basin magmas are more comparable to the mid-ocean ridge environment, where relatively dry mantle undergoes melting largely above the dry solidus, to an extent controlled by mantle T_p. Such compositions define the F-intercept of the wet melting trends characteristic of both back-arc basins and ridges (Figures 16d and 16h). When the water content of the mantle is low enough that all the water can be accommodated in nominally anhydrous mantle phases, melting is "organized" by the 2-D triangular shape of the melting regime, and provided all the melt from the triangle can be pooled, leads to a F versus H_2O_o trend with a negative slope [*Asimow and Langmuir*, 2003; *Cushman et al.*, 2004]. Beneath back-arc basins, the water content is large enough that significant melting takes place at depths well below those at which the shape of the melting regime has any meaning, leading to a quantitatively different set of controls on the extent of melting. Back-arc basins achieve a positive vector of increasing F with increasing H_2O_o because high-F, wet melting paths contribute large F before entering the triangular melting regime (Figures 16e–16g). In the ridge case, bulk F is dominated by the large source volume of low F, wet melts. In the back-arc case, the bulk F is dominated by high F, wet melts.

[42] The unique geometry of plate and mantle juxtaposition at subduction zones also leads to a number of dynamical consequences for both the shape of the melting regime and the interpretation of erupted melt compositions. We examine, for example, the relationship between H_2O and melt regime volume. At mid-ocean ridge settings, where H_2O is present as a trace element dissolved in solid mantle minerals, the pooling of melt within the triangle leads to a decrease in bulk melt fraction with increasing H_2O_o (e.g., Figure 16d [*Asimow and Langmuir*, 2003]). Water in such a system increases the depth at which adiabatically upwelling mantle intersects the solidus (P_o), allowing the production of low-degree, wet melts within the ascending mantle before intersecting the dry solidus (Figures 16a–16d). Increasing H_2O_o in this system increases both P_o and the volume of the low-F, wet melting region. For the maximum H_2O_o in MORB ($\sim$500 ppm), wet melting begins at $\sim$4.5 GPa ($\sim$150 km (Figure 16a)). This large, low-F, wet-melt volume below the base of the dry-melt triangle skews the bulk melt fraction over the full melting region to much lower values.

[43] Beneath back-arc basins, however, this relationship may not be so simple. In these settings, the initial depth of melting may be controlled by the water-saturated solidus, which is independent of the normal controls on melting regime shape imposed by mantle T_p, plate spreading, and passive corner flow as at ridges. The addition of water to back-arc mantle does allow a larger mantle volume to experience partial melting than at a dry ridge (Figure 16g), but increasing the flux of H_2O into the mantle does not necessarily increase the volume of the wet melting regime. Instead, an increased flux of water effectively breaks the control of the melting regime shape on the depth and extent of melting, allowing for higher extents of melting of the same mantle volume. In this view, F increases with increasing H_2O_o. Because H_2O_o increases toward the trench (i.e., H_2O_o is higher at shallower depths), melting will also take place at progressively shallower depths, although the amount of H_2O that can be dissolved in melt at these lower pressures is much less. The depth of melting thus may, in fact, decrease as H_2O_o increases in the mantle wedge.

[44] Another dynamical consequence of subduction zone geometry relates to melt focusing and pooling. Mid-ocean ridge magmas reflect the bulk melt fraction (F_B [*Plank et al.*, 1995]) removed from mantle passing through the wide melting region beneath the ridge, but back-arc melts may be biased toward the maximum melt fraction (F_{MAX}). Melt focusing and pooling across the melting regime is an efficient process beneath ridges. Mantle flow beneath back arcs, however, is perturbed by wedge circulation, which deflects flow toward the wedge corner and the slab [*Spiegelman and McKenzie*, 1987; *Conder et al.*, 2002; *Kincaid and Hall*, 2003]. This geometry could lead to preferential focusing of low-F melts from the slab side of the melting triangle into the magmatic plumbing system beneath the arc. The effect of such a process would lead to a preferential increase in the bulk F recorded by melts erupted at back arcs, and the magnitude would scale with proximity to the slab, which controls wedge circulation and H_2O_o concentration. The active flux of water to the back-arc melting region may also lead to lower viscosity melts that pool more vertically, or even organize flow into vertical instabilities (Figure 17). Thus such melts may not blend as efficiently across the melting region as at ridges. In such a case, the range of lava compositions erupted at back arcs might be greater than at ridges, and we would expect to find dry and wet melts locally erupted, as is the case in the Mariana trough. Alternatively, the positive H_2O_o-F trend could instead reflect binary mixing of separate melts from a dry, MORB melting regime and a high-F, wet melting regime, although in this case the initial mantle source composition could be different for each melting end-member. In order to further evaluate the distribution and relative importance of wet and dry melts and their mantle sources on local and regional scales, however, more data are required for more densely sampled back-arc regions.

[45] The positive relationships between H_2O_o and F at back-arc basins derive from several consequences of an active supply of water from the subducting plate. (1) Water concentrations in excess of the minimum solid mantle storage capacity (>0.18 wt %) suggest the presence of free fluid or melt along a significant portion of the back-arc mantle path, rather than incorporation of H_2O in solid flow as is thought to be the case at ridges. (2) Because of the inverted thermal gradient imposed by the cold subducting plate, and the small volume of initial melts relative to the mantle, the deepest parts of the melting paths beneath back arcs may not be adiabatic. (3) Melt production beneath back arcs is not governed solely by the shape of the melting regime generated by the solid flow field, but also by the magnitude of the H_2O flux from the slab. (4) The extent of lateral pooling beneath back arcs may be less than at ridges, leading to a dominance of vertical pooling and strong local vectors of variability.

5. Conclusions

[46] The compiled data and calculations presented here confirm the unique role of H_2O in the mantle melting process beneath back-arc basins. The first-order observation

of negatively sloping correlations between H_2O and TiO_2 in erupted back-arc basin basalts suggests a relationship between H_2O and extent of melting, since TiO_2 is incompatible during melting and is conservative within the subduction zone (i.e., all TiO_2 in the magmas is assumed to come from the mantle wedge, not from the slab). Using TiO_2 to calculate the mantle melt fraction (F) reveals positive, linear correlations between H_2O addition from the slab and F beneath back-arc basins the world over, but with slopes and intercepts that vary regionally. The positive wet melting functions at back arcs contrast with what is observed at mid-ocean ridges, where increasing mantle H_2O contents correlate with decreasing F. We conclude that back-arc mantle temperature varies regionally by $\sim$150°C and plays an important role in melt production in a manner similar to mid-ocean ridges, largely controlling physical variables such as axial depth and crustal thickness, as well as wet melt productivity (dF/dH_2O_o). Superimposed on temperature is the effect of H_2O on melt production, as high H_2O_o correlates positively with high F. The concentration of H_2O_o varies in the back-arc mantle as a function of distance from the trench and relates to the orientation (+ or −) of wet melting trends modeled at back-arc basins and mid-ocean ridges. We infer that the presence of the subducting slab in close proximity to back-arc spreading centers initiates a critical change in the role of water during melting, from a relatively passive participant in adiabatic melt generation beneath mid-ocean ridges, to an active fluxing agent that drives melt generation in the back-arc environment.

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Appendix

A. Additional tests of modeling outcome

A.1. Use of Na2O as single-element proxy for F

Modeling F using Na2O instead of TiO2 also results in a positive trend in H2Oo vs. F. Model outcome is thus not dependent on the use of a particular element to calculate F, and is consistent with the multi-element approach of [Stolper and Newman, 1994]. Results using Na2O, however, yield a steeper slope at the same intercept (Fig. A1; using same DNa parameters as in Fig. 5). The slope is steeper because Na2O is much less conservative than TiO2, and is likely added to the mantle along with water from the slab (as predicted by Stolper and Newman [1994]). High water brings Na2O with it, causing wetter magmas to have higher Na2O than if pure H2O had been added. We interpret higher Na2O as lower F, which results in a steeper slope. If Na2O added with H2O were subtracted from the melt compositions, the points would shift to higher F, and may overlap the TiO2-derived array. Despite the utility and widespread use of Na2O in MORB systems [e.g., Klein and Langmuir, 1987], we choose not to use it because TiO2 is more conservative, but this analysis demonstrates that our results are not strictly dependent on the use of TiO2 as a proxy for F.

B. Results for each back-arc basin and mid-ocean ridge region

B.1. Back-Arc Basins

In global back-arc basins, the negatively-sloping correlation between H2O and TiO2 first observed in the Mariana trough [Stolper and Newman, 1994] generally holds true. To first order, back-arc melts have higher H2O concentrations than mid-ocean ridge basalts (MORBs), coinciding with decreasing TiO2 concentrations (Fig. 4a, c-d). In detail, however, each basin forms a distinct trend in TiO2 vs. H2O, and at low H2O concentrations, select regions even show positive H2O-TiO2 trends similar to MORB (Fig. 4b-c).

B.1.1. Mariana Trough

The Mariana trough is located west of the Mariana arc in the western Pacific ocean. Water concentrations in Mariana trough lavas range from ~0.2 to 2.0 wt. % [Gribble, et al., 1996; Gribble, et al., 1998; Newman, et al., 2000; Stolper and Newman, 1994], which exceeds the upper limit of H2O concentrations in mid-ocean ridge basalts [e.g., Michael, 1988] and translates to ~0.01 to 0.37 wt.% H2O in the mantle source (Fig. 10a). We have shown that our modeling of this region agrees satisfactorily with the original study of Stolper and Newman [1994] (Fig. 9b), despite differing data sets and methodology. The results of this modeling reveal a positive, linear correlation between H2Oo and F at the Mariana trough (Fig. 10a) that contrasts with the trends from mid-ocean ridges (Fig. 11). In this region, we use a constant DMM source model because Stolper and Newman [1994] argued convincingly in favor of a constant, MORB-like mantle in the Mariana trough, as does our evaluation of the mantle source using trace element constraints (Fig. 7).

B.1.2. East Scotia Ridge

The East Scotia ridge is located in the southeast Pacific Ocean, west of the South Sandwich arc. Water data are available for eight of the nine well-defined spreading segments of this basin [Fretzdorff, et al., 2002; Muenow, et al., 1980; Newman, Unpublished data]. Water concentrations from the northern E2-E4 segments, which contain the largest trace element signature from the subducted slab [Fretzdorff, et al., 2002], overlap with the Mariana trough (0.2-2.0 wt. %) and correlate negatively with TiO2 concentration (Fig. 4c). Segments E5-E8, however, have H2O concentrations similar to normal MORB (~0.1-0.6 wt. %), and high TiO2 samples have higher H2O (Fig. 4b). The southernmost segment, E9, is more like the northern segments, with high H2O (0.3-1.0 wt. %) correlating negatively with TiO2 (Fig. 4c). One basaltic glass from segment E9 contains 2.0 wt.% H2O [Muenow, et al., 1980], but this sample was excluded because it has degassed. Our modeling reveals two positively-sloped H2Oo-F trends distinguishable at the East Scotia ridge, defined by the H2O-rich segments E2-E4 in the north and E9 in the south, respectively. The E2-E4 data define a trend steeper than the Mariana trough (Fig. 10c), whereas the sparse E9 data fall along a shallower trend (Fig. 10d). Segments E5-E8 scatter along the F-axis near zero H2Oo, similar to mid-ocean ridges. The mantle beneath segments E2-E8 is variably more depleted than DMM, and the difference between constant and variable source models is small (Table 1; Fig. 7). Segment E9, on the other hand, taps an enriched mantle, and applying source constraints from the TiO2/Y model shifts the modeled points to higher F, but this has a minor effect on the slope of the E9 trend.

B.1.3. Manus Basin

The Manus basin is located north of Papua New Guinea, behind the New Britain island arc, and comprises two main spreading segments: the Manus spreading center (MSC) and the Eastern Rifts (ER). The transform fault defining the western boundary of the MSC forms an extensional transform zone (ETZ)

that is also volcanically active. Concentrations of H2O in the modeled basaltic glasses range from 0.2-1.4 wt.% [Danyushevsky, et al., 1993; Kamenetsky, et al., 2001; Shaw, et al., 2004; Sinton, et al., 2003] and correlate negatively with TiO2 (Fig. 4d). The model results show that the Manus basin defines a positively-sloped correlation between H2Oo and F, with H2Oo concentrations approaching 0.4 wt.% (Fig. 10b), but a slope significantly shallower than the Mariana trough. Our modeling reveals that the mantle beneath the ER, MSC, and ETZ is somewhat depleted relative to DMM (Fig. 7). Recall, however, that the source depletion model used here yields a maximum TiO2 content of the mantle. The high melt fractions predicted for certain samples (29-38%; Fig. 10b) could result from over-estimation of , if TiO2, Nb, and Ta were added to the mantle from the subducted slab. Relative to a constant DMM model, however, the mantle source corrections result in a small difference in the slope of the H2Oo-F trend (Fig. 10b).

B.1.4. Lau Basin

The Lau basin is in the southwest Pacific ocean, behind the Tonga-Kermadec arc. This basin contains five active spreading segments for which glasses and melt inclusions have been analyzed for H2O: the Mangatolu triple junction (MTJ), the Central Lau spreading center (CLSC), the Intermediate Lau spreading center (ILSC), the Eastern Lau spreading center (ELSC), and the Valu Fa ridge (VFR) [Aggrey, et al., 1988; Danyushevsky, et al., 1993; Kamenetsky, et al., 1997; Kent, Unpublished data; Kent, et al., 2002; Newman, Unpublished data; Pearce, et al., 1995; Peate, et al., 2001; Sinton, et al., 1993]. The Lau basin is v-shaped and actively propagating towards the south, where the VFR approaches within 120 km of the arc. The northern MTJ, CLSC, and ILSC segments are offset from the southern ELSC-VFR segments, and although both regions show decreasing TiO2 with increasing H2O concentrations (Fig. 4d), the orientations of these trends are geographically distinct and suggest natural groupings. We have thus grouped the MTJ, CLSC, and ILSC samples together under the northern Lau basin designation (NLB), and the ELSC and VFR samples together under the southern Lau basin designation (SLB) when discussing regional characteristics of the Lau basin. The H2O concentrations of basaltic melt inclusions from the VFR are the highest of any back-arc basin (~1.1-2.8 wt.%; [Kent, Unpublished data]), and translate into some of the wettest mantle revealed by this study (H2Oo≤0.50 wt.%). Model results for the ELSC and VFR are the most affected by prior depletion of the mantle source (Fig. 10e). The low Nb and Ta concentrations require at least 2.5% prior melt removal from DMM to model CE patterns in this region (Fig. 7f), which results in low (Table 1) and shifts results to smaller melt fractions (Fig. 10e). Using the depleted source instead of DMM decreases the calculated F at the VFR by as much as half, although the slope difference between the two models is only 4%.

B.1.5. Sumisu Rift

The Sumisu rift is located immediately behind the Izu arc (~20 km), to the north of and along strike with the Mariana trough. The H2O concentrations of glasses from this region span a narrow range (1.0-1.2 wt.%; [Hochstaedter, et al., 1990]). The mantle source in this region is slightly more depleted than is inferred for the Mariana trough source. Although only a small number of Sumisu rift samples are suitable for this modeling (n=4), these show a negative correlation between H2O and TiO2 (Fig. 4d) and a positive correlation between H2Oo and F (Fig. 10a), although with so few samples the details of the model outcome in the region are less robust.

B.2. Mid-Ocean Ridges

We begin here with a review of recent results from two hotspot-influenced mid-ocean ridges: the Galápagos spreading center (GSC; [Cushman, et al., 2004; Detrick, et al., 2002]) and the Mid-Atlantic ridge approaching the Azores platform (FAZAR; [Dixon, et al., 2002; Asimow, et al., 2004]). Basalts from these ridges are "damp" (i.e., 0.1-1.2 wt. % H2O) relative to the global range of H2O concentrations in basalts from all tectonic settings (0-6 wt. % H2O; [Newman, et al., 2000]). Lavas from these regions show both H2O and TiO2 concentrations increasing with proximity to the hotspot and, consequently, positively-sloped correlations between TiO2 and H2O opposite those observed in most back arcs (Fig. 4b). For comparison, we also consider two small, anomalous spreading centers in this section: the Woodlark basin and the North Fiji basin. The Woodlark basin, although actively rifting mantle that was once behind the Solomon Island arc, is not presently a back-arc basin. The N160° and Triple Junction segments of the northern North Fiji basin are located >700 km behind the Vanuatu arc, and are the only segments from this basin with glasses in which H2O has been measured. Although technically in a back-arc setting, the lavas erupted at these segments show little to no affinity with those from other back-arc basins that are more proximal to the trench (~200-450 km), so this basin is also included with the discussion of mid-ocean ridges.

B.2.1: Galápagos Spreading Center

The Galápagos spreading center (GSC) in the Pacific ocean is a hotspot-influenced spreading ridge. From ~98°-91°W, the GSC changes from an average, intermediate-spreading ridge in the west to an inflated axial high in the east, coincident with the inferred intersection with the Galápagos plume and characterized by thicker crust, enriched lava compositions, and high H2O concentrations [Cushman, et al., 2004; Detrick, et al., 2002]. The coincidence of both incompatible element and H2O enrichment of these lavas results in a positive correlation between TiO2 and H2O along the GSC (Fig. 4b). The concentrations of H2O in the GSC glasses used in our modeling are 0.1-0.6 wt.%, with the highest H2O contents associated with the eastern axial high [Cushman, et al., 2004]. Detrick, et al. [2002] and Cushman, et al. [2004] argued for decreased mean extents of melting and increased melt production beneath the axial high, citing the ability of H2O to expand the volume of the melting region by adding a large, low-F volume at the base of the melting column (i.e., the same argument used by Asimow and Langmuir [2003]). Figure 11b shows that the melting model presented here also produces a negatively-sloped correlation between H2Oo and F at the GSC in calculations using a constant DMM source, but using the source constraints of Cushman, et al. [2004], which call for along-strike variations of the mantle source at this ridge, the negative slope disappears.

B.2.2: Mid-Atlantic Ridge near the Azores

The northern Mid-Atlantic ridge approaches the Azores hotspot between 33° and 40°N latitude. A geochemical gradient along the ridge axis (e.g., La/Sm, 143Nd/144Nd, 206Pb/204Pb; [Dosso, et al., 1999]) marks the increasing influence of enriched mantle towards the Azores platform. The range of H2O contents in basalts from this region is ~0.1-1.2 wt.% [Dixon, et al., 2002], which is within the range reported for global MORB [Michael, 1988]. Asimow, et al. [2004] use these data to model the mantle source, melting, and magmatic fractionation trends in this region, and Asimow and Langmuir [2003] draw on these Azores data to support a melting model that produces negatively-sloped correlations between H2Oo and F (Fig. 3).

These results are, however, sensitive to the assumed composition of the mantle source. For example, if we take either a constant DMM source or the modeled constant source compositions of Asimow, et al. [2004] and then use our batch melting model to determine F and H2Oo, the FAZAR sample suite defines a smooth, negative H2Oo-F trend (Fig. 11a) in agreement with the results of Asimow, et al. [2004]. If, however, we adopt a mantle source model that varies along strike (using TiO2/Y; see section 2.5), as suggested by along-strike isotopic variations [Dosso, et al., 1999], a single, negatively-sloped H2Oo-F correlation in this data set becomes less apparent (Fig. 11a). Specifically, several samples from the Azores platform with high TiO2 also have high TiO2/Y, which we interpret as more enriched mantle. Making this correction to adjusts the calculated F and H2Oo to higher values, since larger extents of melting are required to generate the high TiO2 concentrations of the lavas from a more Ti-enriched source. Correlations between radiogenic isotope ratios (e.g., 87Sr/86Sr, 143Nd/144Nd, 206Pb/204Pb) and TiO2/Y in the FAZAR sample suite suggest that the mantle source composition does indeed vary in this region, and Asimow, et al. [2004] also required TiO2 variations in the mantle source to be able to model accurately liquid lines of descent. As explained above, our TiO2/Y model provides a maximum constraint on the TiO2 concentration of the mantle and thus may overestimate F since Y concentration may also vary in the mantle or may increase if melting occurs within the garnet stability field. These calculations underscore the critical nature of accurately constraining the mantle source. The Azores platform region certainly taps a more enriched mantle relative to the rest of the FAZAR region along the MAR, and the entire FAZAR region likely taps variably enriched mantle. Uncertainties in the concentrations of incompatible trace and minor elements of the source that accompany this enrichment (especially TiO2 and Y, for our treatment of the data), however, make a definitive statement difficult. Consequently, we can at this point only conclude that basalts from Azores platform may define a negative H2Oo-F trend distinct from the rest of the FAZAR region, both of which are distinct from the Mariana trough (Fig. 11a).

B.2.3 North Fiji Basin

As described above, the N160° and Triple Junction segments of the North Fiji basin are located >500 km behind the Vanuatu arc in the southwest Pacific ocean. Although back arc in a geographic context, lavas erupted at these segments show little evidence for slab-derived components in the magma source, perhaps because the ridge is so distant from the arc. Instead, lavas from these northern segments suggest the influence of enriched, EMORB- or OIB-type mantle, likely related to either the Samoan hotspot or Fiji platform mantle sources [Eissen, et al., 1994; Nohara, et al., 1994]. The H2O concentrations of glasses from these North Fiji basin segments are ~0.2-1.0 wt.% [Aggrey, et al., 1988; Danyushevsky, et al., 1993], and H2O also correlates positively with

TiO2 (Fig. 4b) in lavas from this ridge-like basin. As at ridges, this translates into a negative relationship between H2Oo and F (Fig. 11c). Many of these samples, however, have anomalously high TiO2/Y (Fig. 8b), which leads to a large correction to higher and consequently higher F relative to the DMM model as seen on Figure 11c. Although the variable source model generates more scatter, it does not change the negative orientation of the trend.

B.2.4. Woodlark Basin

The Woodlark basin, in the western Pacific ocean, is bounded on the west by Papua New Guinea, where the spreading ridge intersects continental crust near the D'Entrecasteaux islands, and on the east by the San Cristobal trench, where the ridge is actively subducting beneath the Solomon islands. Although not presently a back-arc basin, the Woodlark spreading center is opening in a region that once was behind the Solomon Island arc (i.e., before subduction reversed direction at 5 Ma; [Muenow, et al., 1991]) or affected by subduction at the Trobriand trough [Taylor, et al., 1999]. The concentrations of H2O in glasses from the Woodlark basin used in this modeling are 0.2-1.6 wt.% [Danyushevsky, et al., 1993; Muenow, et al., 1991], with the most H2O-rich samples from the western segment nearest the D'Entrecasteaux islands and from one eastern segment close to the San Cristobal trench. Water concentrations as high as 1.9 wt.% are reported for basalts from the eastern Woodlark basin [Danyushevsky, et al., 1993], which translate to H2Oo concentrations of up to 0.50 wt.%. These melts, however, have unknown CO2 concentrations and were saturated with pure H2O at their depth of collection (and were therefore excluded from modeling); they may, however, indicate that the H2O concentration of the mantle in this region extended to >0.5 wt.%, placing this among the wettest portions of the oceanic mantle. Basalts from the central region of the basin have low H2O concentrations similar to MORB (0.2-0.3 wt.%) that correlate positively with TiO2 (Fig. 4c), but basalts from the H2O-rich segments in the east and west have negative correlations between TiO2 and H2O (Fig. 4c). As a result of these trends, the relatively dry samples from the central part of the Woodlark basin define a tight negatively-sloped correlation between H2Oo and F (Fig. 10d) like mid-ocean ridges, but the wet samples delineate a positive H2Oo-F trend similar to the Mariana trough. Note that the mantle source composition is also inferred to vary along-strike in this region, with the central and eastern segments melting mantle more depleted than DMM (Table 1; Fig. 7), and the western segment near the D'Entrecasteaux Islands tapping an enriched source (TiO2/Y=0.054). Applying these corrections results in relatively minor changes in the calculated F and H2Oo values of samples from the central and eastern segments but results in a more noticeable shift in points from the western segment (Figure 10d). In any case, both constant and variable source models show the same result in this region: a negative correlation between H2Oo and F in the dry samples, and a positive correlation among the wet ones. Although the positive trend at this basin is strengthened by combining data from the geographically separate eastern and western segments, which may not be a valid connection, the data from each wet segment taken independently also indicate positive relationships between H2Oo and F.

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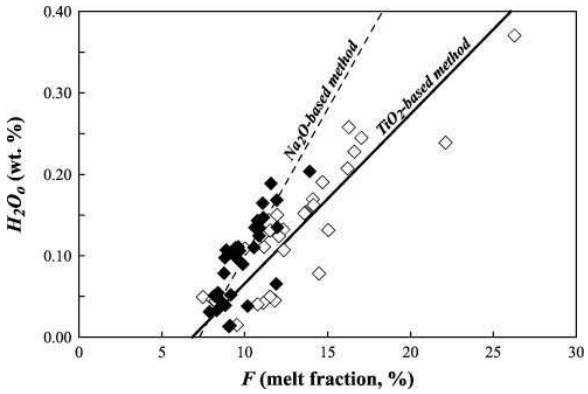

Fig. A1. H2Oo vs. F for the Mariana trough, comparing results of calculations using TiO2 and Na2O as proxies for F. The open symbols and solid line are the results and linear best fit using TiO2 as the proxy for F (y=2.082x-0.142; r2=0.798). Filled symbols and dashed line are the results and linear best fit using Na2O as the proxy for F (y=3.646x-0.266; r2=0.604).