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A physical model for the volume and composition of melt produced by hydrous fluxing above subduction zones

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Thermal models of subduction zones, restrict the melt source region to a domain at sufficiently high temperature with water present (either as a free phase or in hydrous minerals). Water, released into the mantle by slab dehydration, traverses the wedge horizontally by a combination of (i) vertical movement as a fluid phase and (ii) fixed in amphiboles carried by the induced mantle flow; only in mantle hotter than amphibole stability can melts escape upwards. We develop a one-dimensional model for the source region fluxed with water. The induced mantle flow advects heat laterally to balance the latent heat of melting, in a column where the liquidus of the melt is depressed by its water content. Melt flux, fraction, temperature and water content are calculated assuming steady state. Melt compositions are predicted from the melt fraction distribution as a function of depth, constrained by the experimental data of Green. On investigating a range of plausible models, we find that the average degrees of melting predicted vary from *ca.* 2 to 8%. The predicted primary magmas are mafic high magnesium basalts with water contents ranging from 1.6 to 6 wt%, and temperatures from 1160 to 1290 °C. Models with shallower depths of segregation have higher degrees of melting and lower water contents. The volumes predicted by the physical model are a strong function of the water flux assumed to enter the source region. Previous estimates of arc growth would suggest either low water fluxes or that not all the melt reaches the arc crust.

1. Introduction

The petrogenesis of magmas generated above subduction zones is as controversial today as it was when the petrological implications of subduction were first realized (Green & Ringwood 1968). The problem in part stems from the multiplicity of potential magma sources and in part from the additional petrological complexities of melting in a hydrous system. There are a number of lines of evidence for the involvement of water in the petrogenesis of arc magmas. Volcanicity is often explosive, lavas are hydrous and may contain OH-bearing phenocryst phases, high-pressure modules may contain amphibole, while eruption temperatures are often between 1100 and 1200 °C, lower than their dry compositional equivalents (Gill 1981, p. 65). A number of isotopic tracers in arc magmas, notably ^{10}Be (Tera *et al.* 1986), ^{207}Pb and ^{87}Sr and possibly some of the more incompatible trace elements (Pearce 1982) are derived from the subducted hydrated oceanic crust and sediments.

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Thermal models of subduction zones show that temperatures in the downgoing lithosphere are too low to allow extensive melting of the subducted crust (Davies & Stevenson 1991; Honda 1985). However, experiments on amphibole stability indicate that downgoing oceanic lithosphere will dehydrate by a near-isobaric amphibole to eclogite transition at *ca.* 2.5 GPa (Tatsumi 1986). Figure 1 illustrates how water is transported away from the subduction zone. Water released from the subducted oceanic crust would amphibolitize the overlying mantle and then be carried down by the mantle flow in the wedge to a similar amphibole breakdown reaction in peridotite at *ca.* 3.0 GPa (Tatsumi 1986). From there the free water rises vertically through the already amphibolitized mantle until it reaches fresh mantle. There it reacts again forming amphibole. It is then carried down again by the induced mantle flow. The process repeats at *ca.* 3.0 GPa where the amphibole breaks down again. The net effect is the lateral transport of water across the mantle wedge, towards the wet solidus (Davies & Stevenson 1991). Addition of water to mantle above its wet solidus would cause melting. In this paper we model melt production by such hydrous fluxing, the aim being to quantify the rate of melt production, the degree of melting, melt water contents, temperatures and their major element compositions.

2. Thermal model

Davies & Stevenson (1991) have reviewed the numerous thermal models of subduction zones published over the past two decades. The differences between models mainly reflect either the incorporation of mechanisms which have since been shown to be of much less significance, especially strain heating at the slab surface; or the neglect of others that are accepted to be significant, such as induced wedge flow. The resulting thermal models (Andrews & Sleep 1974; Honda 1985; Davies & Stevenson 1991) do not differ greatly, and can be used as a basis for discussing the source region of subduction zone magmatism. The prediction of volumes and compositions of magmas from these thermal models will provide further tests and constraints on these models.

We shall consider that the water reaches the source region by the mechanism outlined in the introduction (Davies & Stevenson 1991), illustrated in figure 1. Note that we will get some melting at the water saturated solidus, but the volumes of melt produced are so low (since the amount of free water present is so low) that they will percolate upwards but are highly unlikely to segregate into cracks. Hence these melts remain in local chemical equilibrium and in effect they are similar to the hydrous silica-rich phase in that they transport water vertically, continuing the net horizontal movement of the water. The start of the source region is defined as the point where in the vertical path of the wet melt, it reaches mantle that is hotter than the thermal stability of amphibole. The width of the source region in this model is controlled by the position of the amphibole breakdown curve. It is assumed that over this width and height one gets sufficient melting to allow segregation into cracks. The segregation height could be at the maximum temperature of the source region, but given our poor understanding of magma segregation we have investigated the effect of varying this height.

It is unclear how big a role adiabatic decompression plays in generating subduction zone magmas relative to hydrous fluxing (Davies & Stevenson 1991). In this model we restrict ourselves totally to the hydrous fluxing end-member. Adiabatic decompression of dry mantle was addressed in McKenzie & Bickle (1988). Note

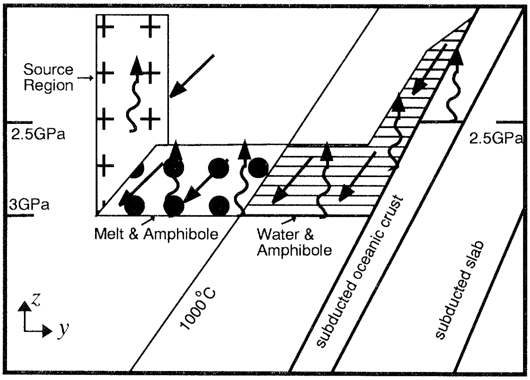

Figure 1. Schematic diagram of the proposed water transport mechanism and the orientation of the assumed source region where amphibole is no longer stable. The straight arrows represent the motion of matrix, while the wiggly arrows represent motion of either melt or water. Note there is a region where amphibole and melt is stable, and another where water and amphibole is stable. The source region and z axis are assumed to be vertical; the y axis horizontal, and the x axis is assumed to be horizontal into the paper, i.e. along strike of the subduction zone. The model considers v_s to be the horizontal component of the matrix velocity.

melting cools the mantle; this is true for melting due to hydrous fluxing as well as melting due to adiabatic decompression.

3. Wet melting

The main controls on the amount of melting are the rate of heat advected into the melt region by the induced flow in the mantle wedge, the water flux from the subducted plate and the melt thermodynamics; that is the relationship between melt temperature, water content and melt fraction. To a first approximation this can be modelled by a one-dimensional column of partial melt rising into hotter mantle. Hot unmelted dry mantle flows into the side of the column. This melts by reaction with the wet magma and any excess water flowing up the column. The latent heat of melting results in a temperature drop. Cooler partly melted (i.e. residual) mantle leaves the melt region on its downstream side.

We assume steady state, with constant water flux up the column. Melt, water dissolved in the melt, and any excess water are assumed to travel with the same velocity. The melt is in local thermal equilibrium with solid. We ignore both advection of heat by the melt and conduction of heat up the melt column. This can be shown *a posteriori* to be reasonable. Heat will be conducted horizontally into the column but this would correspondingly reduce the heat advected into the column. We ignore possible conductive gradients across the melting column.

The amount of melting is calculated from conservation of energy. We balance the latent heat of melting with the heat advected in by the solid flow. To do this we need to be able to calculate the (weight) fraction of peridotite which melts as a function of temperature and water content. Experiments on liquidus depression give an empirical linear relationship between mole fraction water and liquidus depression, given a mole equivalent mass of *ca*. 255 g mole^{-1} for melt, not dissimilar to the values derived by Burnham (1979). However, this relationship does not adequately describe melting of wet peridotite as melt composition changes across the melt region. For example an olivine tholeiite melt which has a dry liquidus consistent with the dry peridotite solidus has a water saturated liquidus at 3 GPa of *ca.* 1100 °C (Stern *et al.* 1975) which is inconsistent with the 3 GPa peridotite wet solidus at *ca.* 1000 °C (Green 1973). This compositional dependency is approximated by adopting a larger constant a_1 in the relation describing liquidus depression (see equation (1) below). The larger value for a_1 correctly describes the separation of dry and water saturated solidi. Melt temperature increases as a function of increasing melt fraction. This is accommodated by relating the melt temperature, to the dry melt temperature, at the same melt fraction and pressure, by

$$T_{\rm l} = T_{\rm d} - a_1 X^{\rm m}_{\rm w}, \tag{1}$$

where $T_{\rm l}$ is the wet melt temperature, a_1 is a constant, $X^{\rm m}_{\rm w}$ is the mole fraction content of water and $T_{\rm d}(X, P)$ is the temperature of a dry melt at temperature P and degree of melting X, calculated from the parametrization of McKenzie & Bickle (1988, eqns (18)–(20)). The liquidus depression data (Stern *et al.* 1975; Green 1972) indicate no detectable variation in a_1 with pressure between 2 and 3 GPa, the region of interest. Maximum solubility of water in melts is taken as 25 wt% at 3 GPa (Green 1973; Stern & Wyllie 1973). The pressure dependence of maximum water solubility is unimportant as the melts are only saturated over a small vertical interval.

The conservation of energy equation is

$$L\,{\rm d}M/{\rm d}t = v_{\rm s} C_p \rho_{\rm s}(T_{\rm i} - T_{\rm l})/\Delta y, \tag{2}$$

where L is the latent heat of melting, $v_{\rm s}$ is the matrix velocity, C_p is the specific heat capacity at constant pressure, $\rho_{\rm s}$ is the mantle density, Δy is the source width, while $T_{\rm i}$ is the mantle temperature just before it enters the source region. Note since ${\rm d}M/{\rm d}t$ is the mass rate of melting per cubic metre, $\Delta y\,{\rm d}M/{\rm d}t$ gives the melt production rate at height z per metre of arc length. The width of the melt region, Δy, does not affect the solution.

The equation for the mole fraction of water in the melt is

$$X^{\rm m}_{\rm w} = \frac{M_{\rm e}}{18.02(1 - W^{\rm m}_{\rm w})/W^{\rm m}_{\rm w} + M_{\rm e}} = \frac{M_{\rm e}}{18.02W/F + M_{\rm e}} \tag{3}$$

and

$$X^{\rm m}_{\rm w} = \frac{M_{\rm e}}{18.02(1 - W^{\rm m}_{\rm w}({\rm sat}))/W^{\rm m}_{\rm w}({\rm sat}) + M_{\rm e}}, \tag{4}$$

where $X^{\rm m}_{\rm w}$ is the mole fraction of water in the melt, $W^{\rm m}_{\rm w}$ is the weight fraction of water in the melt, $M_{\rm e}$ is the mole equivalent mass for the melt, W is the melt flux (kg m^{-2} s^{-1}), while F is the water flux (kg m^{-2} s^{-1}), if the melt is water-saturated, which it must be at infinitesimal degrees of melting in a hydrous system. Equations (1) and (3) together self-consistently describe the behaviour of the wet solidus and liquidus. The above parametrization is also continuous with the dry parametrization.

From the conservation of mass we can relate the melting rate to the melt flux gradient by

$${\rm d}M/{\rm d}t = {\rm d}W/{\rm d}z. \tag{5}$$

Where z is the height above the base of the melting column. The bottom of the melting column, z_0 is at $z = 0$ km (at a depth of 100 km); while z_1, the top of the melting column, is at a height, $z = 40$ km (i.e. a depth of 60 km) for the basic model. Hence from substituting equations (5) and (1) into equation (2) we get

$${\rm d}W/{\rm d}z = (v_{\rm s} C_p \rho_{\rm s}/L\,\Delta y)\,(T_{\rm i} - T_{\rm d} + a_1 X^{\rm w}_{\rm m}). \tag{6}$$

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Table 1

symbol	parameters		
	name	value	units
C_p	specific heat capacity	10^3	J kg^{-1} °C^{-1}
L	latent heat	4.5×10^5	J kg^{-1}
Δy	effective source region width	2×10^4	m
M_e	molar equivalent mass of melt	255.0	g mole^{-1}
a_1	constant relating X_w^m to (T_i-T_d)	642.0	°C
F	water flux	0.6×10^{-8}	kg m^{-2} s^{-1}
ρ_s	density of solid	3.34×10^3	kg m^{-3}
v_s	velocity perpendicular to isotherms	10^{-9}	m s^{-1}
z_0	bottom (at 100 km depth)	0.0	km
z_1	top (at 60 km depth)	40.0	km
T_i^{min}	T_i at z_0	1000	°C
T_i^{max}	T_i at z_1	1250	°C
W	melt flux	—	kg m^{-2} s^{-1}
dM/dt	rate of melting	—	kg m^{-3} s^{-1}

The degree of melting, X, is the ratio of the rate of melting to the solid flux,

$$X = (\mathrm{d}W/\mathrm{d}z)\,\Delta y/\rho_s v_s. \tag{7}$$

We vary T_i from T_i^{min} (usually 1000 °C) linearly to T_i^{max} at the top of the source region. The Davies & Stevenson (1991) thermal model suggests T_i^{max} of *ca.* 1250 °C. Since T_i varies with height in the source region this problem has to be solved numerically. Equation (6) must be solved by iteration, since T_d is a nonlinear function of X; this can be seen since to evaluate X we require T_d via dW/dz.

We assume that the melts rise slowly through the source region, remaining in hydrous as well as thermal equilibrium. By hydrous equilibrium we mean that the water content of the melt is the water content of the whole local system. We take no account of the change in the bulk composition with height as a result of advection, except to assume that water contents are appropriately diluted. Since the system is in steady state the relative contribution of melt produced at height z to the final melt segregating from the top of the source region is in proportion to its rate of production.

The parameters used for the basic model are in table 1. Many of the parameters are uncertain while others naturally vary. The temperatures T_i, melting column height, z_1-z_0 and velocity structure v_s approximate models of Davies & Stevenson (1991). The water flux from subducted crust of 0.6×10^{-8} kg m^{-2} s^{-1} is equivalent to dehydration of 2 km of amphibolite containing 1 % H_2O subducting at 7 cm a^{-1}. This is less than some previous estimates (Peacock 1990); but much of the additional water held in minerals other than amphibole would be released at lower temperatures and pressures, and would escape to the surface. We have investigated the effect of variability of these parameters by evaluating the results for our basic model, changing only one parameter at a time.

4. Results

The effect of varying one parameter at a time, away from the basic model listed in table 1, is illustrated in figure 2. In varying the parameter a_1, we needed to increase

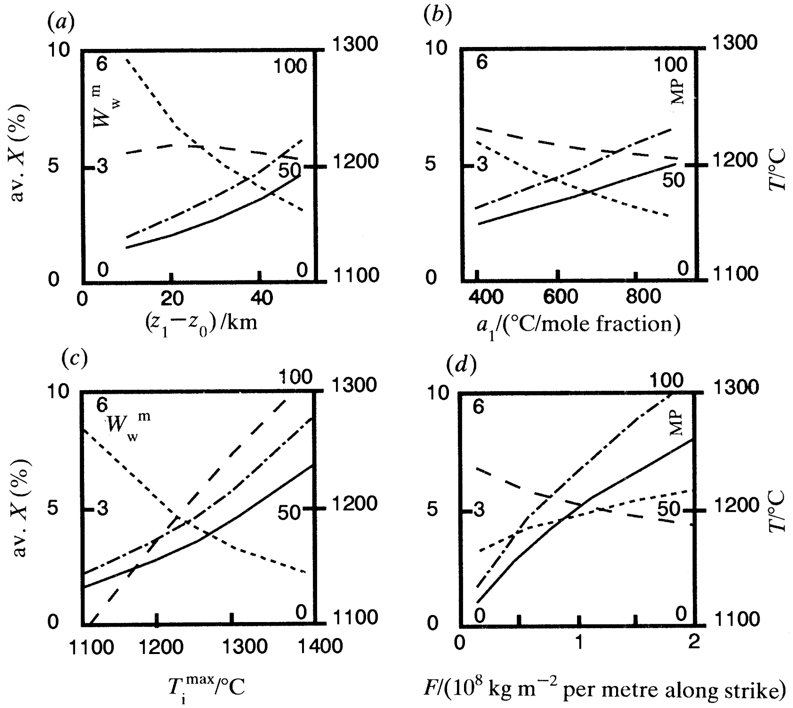

Figure 2. This shows how (i) the average degree of melting, whose scale is on the outside of the left hand side of the figures; (ii) the weight percent water in the primary melt just before segregation, ($W_w^m(z_1)$), whose scale is on the inside of the left hand side of the figures; (iii) the temperature of the primary melt just before segregation ($T_l(z_1)$), whose scale is on the outside of the right hand side of the figures; and (iv) rate of melting (km^3 Ma^{-1} km^{-1} along arc), whose scale is on the inside of the right-hand side of the figures; change as we vary (*a*) the height of the melting column i.e. (z_1-z_0), (*b*) the liquidus depression constant a_1, (*c*) the maximum temperature T_i^{max}, and (*d*) the water flux, F. All the other parameters are the standard parameters found in table 1. ——, Average degree of melting (%); ------, wt% water in melt; —·—·—, magma production (MP) (km^2 Ma^{-1}); – – – – –, primary magma temperature (°C).

T_i^{min} when a_i was decreased to allow melting. In figure 2*a*, it is seen that decreasing the column height decreases the rate of melting. A decrease in the rate of melting is also observed in figure 2*c* with decreasing maximum temperature. Thicker lithospheres would lead to both shorter melting columns and cooler wedges. Hence with thicker lithospheres we would expect less melting, with the melts being wetter and cooler, and the result of lower degrees of melting. Notice that as we decrease the water flux (figure 2*d*), the rate of melting decreases, as do the water contents of the melts. Decreasing the velocity of the advected solid has a similar effect to increasing the water flux, with the exception that the rate of melting is increasing rather than decreasing with velocity. It was found that varying the latent heat of melting had little influence on the system! The predicted rates of melting are within the wide range of estimates of arc additions quoted by Ito *et al.* (1983), but are generally higher than the estimates of Reymer & Schubert (1984) (33 km^3 km^{-1} Ma^{-1}) and Brown & Mussett (1981) (17 km^3 km^{-1} Ma^{-1}). The lower volumes suggest low water fluxes and/or that all the melt produced does not reach the arc crust. The average degrees of melting (2–8%) are generally lower than mid-ocean ridge basalts (MORB) (greater than 10%).

The temperatures of melt at segregation (maximum temperature) are generally lower than those suggested by the equilibration studies of Tatsumi *et al*. (1983). This could be due to the fact that the lavas equilibrated were actually an integration of magmas from a range of pressures, and hence finding the multisaturation point at a single depth might be misleading. They are slightly lower than the liquidus of the high Mg basalt (ID16) of Nye & Reid (1986), 1280 °C. The range of temperatures predicted by the models are not unreasonable given the eruption temperatures of subduction zone lavas (Gill 1981, p. 64).

The water contents of the segregating magmas at 1.6–6 wt % are at the high end of the range (0–5 wt %) suggested by the numerous lines of evidence listed by Gill (1981, p. 116). The predominance of relatively dry magmas suggests either a low water flux, or very near surface degassing. The water contents are sufficiently low that without crystallization they will not exsolve their water until the melts reach the crust.

5. Predicted compositions

There are very few published experimental studies on the compositions of wet melts and there is much controversy (Green 1976). As discussed in the literature there are problems with quench phases, and also iron loss; these problems are illustrated by the lack of agreement among the various experiments (Green 1976; Kushiro 1990; Mysen & Boettcher 1975; Nehru & Wyllie 1975). We have used the three water-saturated experiments of Green (1976) to constrain predictions of compositions from our simple physical model of the source region. To extend the results of Green to parameters other than those of his experiments we have used the compositional parametrization derived by Watson & McKenzie (1991) (an enhancement of McKenzie & Bickle (1988)) for dry melting. We repeat the point made in McKenzie & Bickle (1988) that what we really want are Rayleigh type melting compositions, but we use batch melting compositions since it is unclear how to extract the requisite information on Rayleigh melting from the current experiments. We have evaluated a single shift in melt fraction and pressure that minimizes the sum of the absolute values of the relative misfits between the displaced dry parametrization and Green's three experiments. This is crude, in that we are hoping that wet melting has trends like dry melting. Given the lack of wet melting experiments with degrees of melting, we have no reasonable alternatives. We find that the best fit for the surface is with $\mathrm{d}P_{\mathrm{sat}} = -1.55$ GPa, and $\mathrm{d}X_{\mathrm{sat}} = 0.033$. We ignored Ti, Na and K, since the pyrolite used by Green was based on a Hawaiian basalt which has incompatible element concentrations inappropriate for the mantle. The actual fits to the experiments were reasonable, but given more data on hydrous melting it would be advisable to develop a separate parametrization for the saturated experiments, since the current fit suggests, that the compositional behaviour of wet melting is not identical to the compositional behaviour of dry melting. The average misfit was 5.5 % while the largest misfit was 20 %.

For water-undersaturated conditions we linearly decrease the shifts found for the water-saturated fits in proportion to their molar water contents towards zero for dry conditions; i.e. the compositions are given by

$$C_{\mathrm{wet}}(X, P) = C_{\mathrm{dry}}(X + \mathrm{d}X_{\mathrm{shift}}, P + \mathrm{d}P_{\mathrm{shift}}), \tag{8}$$

where
$$\mathrm{d}X_{\mathrm{shift}} = \mathrm{d}X_{\mathrm{sat}}(X_{\mathrm{m}}^{\mathrm{w}}/X_{\mathrm{m\,sat}}^{\mathrm{w}}) \tag{9}$$

and
$$\mathrm{d}P_{\mathrm{shift}} = \mathrm{d}P_{\mathrm{sat}}(X_{\mathrm{m}}^{\mathrm{w}}/X_{\mathrm{m\,sat}}^{\mathrm{w}}), \tag{10}$$

Table 2

(AOB, HAB, from Tatsumi *et al.* (1983), ID16 from Nye & Reid (1986), 25% and 10%, are from Plank & Langmuir (1988), while OTB, New Georgia, JOR-44, and AT-1 are all from compilation by Gust & Perfit (1987).)

	compositions								
	element (wt %)								
model	Si	Ti	Al	Fe	Mg	Ca	Na	K	the rest
basic model	47.6	1.5	15.5	9.3	11.9	11.2	2.5	0.3	0.4
AOB	49.1	1.0	15.5	9.4	11.6	9.7	2.5	1.1	—
HAB	49.4	0.9	15.7	9.8	12.1	9.4	2.3	0.3	—
OTB	49.7	0.7	15.0	10.6	13.0	9.0	1.6	0.3	—
ID16	48.9	0.7	16.0	9.0	11.4	10.9	2.2	0.5	—
New Georgia	49.8	0.5	12.2	10.1	12.7	10.9	2.0	0.25	—
JOR-44	52.6	0.8	16.6	7.5	9.4	8.5	3.5	0.1	—
AT-1	49.1	0.9	18.1	10.0	8.7	3.3	—	—	—
25%	49.8	0.6	13.3	9.4	13.6	12.0	1.3	—	—
10%	50.3	1.1	17.1	9.4	10.6	8.4	3.0	—	—

where $\mathrm{d}P_{\mathrm{sat}}$ and $\mathrm{d}X_{\mathrm{sat}}$ are the shifts that best fit Green's water saturated experiments to the dry parametrization of Watson & McKenzie (1991), and $X^{\mathrm{w}}_{\mathrm{m\,sat}}$ is the molar fraction of water in the melt at saturation.

We predict the compositions for primary melts at segregation using

$$C = \int_{z0}^{z1} XC\,\mathrm{d}z \Big/ \int_{z0}^{z1} X\,\mathrm{d}z, \tag{11}$$

i.e. the point and depth averaged composition (McKenzie & Bickle 1988). The equivalent anhydrous composition for the basic model (i.e. using the parameters listed in table 1) and a few estimated primary magmas are given in table 2.

The primary compositions discovered in the models are high magnesium low silica basalts. It is seen from table 2 that the predicted primary composition is in the middle of the range of estimates of island arc primary magmas. Remember that Ti, Na and K were not used to constrain the parametrization, hence these values have little meaning. It was found that even though the models covered a wide range of degrees of melting, little variation was found in the compositions, this was largely the result of the higher degree of melting trading off with lower $\mathrm{d}X_{\mathrm{shift}}$, due to lower water contents. The largest change in composition was found when the segregation height was made very shallow, i.e. a variation in pressure rather than in degree of melting. Given the crudeness of the parametrization, and the complexity of the experiments (see start of this section), the correspondence is heartening. It is clear though, that the experimental work needs to be extended to higher pressures (2–3 GPa) and lower degrees of melting (0–0.1); as well as the water-undersaturated régime.

6. Conclusion

We have developed a simple model for melting by hydrous fluxing in a subduction zone. If the varying estimates of arc additions and water contents are correct then our model suggests that the source region is fluxed by a range of water flux. Since estimated magma production rates are larger than arc additions rates this suggest either low water fluxes (1 wt % H_2O in upper 1 km of subducting crust only), or that not all the melt reaches the crust. Predicted degrees of melting are generally lower than for MORB. The poorly constrained compositions of the primary magmas predicted by the model are in the middle of previous estimates of primary magmas. The uncertainty in estimating the model's input temperature highlights the inadequacy of a one-dimensional model, and combined with the probable influence of horizontal variations in melting on compositions, point to the need to solve the problem self-consistently in two dimensions.

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