Extended Methods for: The Importance of ice sheet growth and retreat on magmatism and mantle CO_2 flux

John J. Armitage[1], David J. Ferguson[2], Kenni D. Petersen[3], Timothy Creyts[4]

1. Dynamique des Fluides Géologiques, Institut de Physique du Globe de Paris, Paris, France
2. School of Earth and Environment, University of Leeds, Leeds, U.K.
3. Department of Geoscience, University of Aarhus, Aarhus, Denmark
4. Lamont-Doherty Earth Observatory, Columbia University, U.S.A

Generating an Ice sheet History

Our goal is to create realizations of ice loading over the Iceland rift zone. The ice volume history of Iceland through the entire last glacial period is fragmentary, and poorly constrained prior to the Last Glacial Maximum (26-110 ka)[1]. This lack of information is a result of the growth of the ice sheet to the LGM (~19-26.5ka)[1,2], because it reworks prior deposits and remobilizes sediments, obscuring the earlier record[3]. Additionally, we are mainly interested in the loading history and less concerned with the extent, glacial features, or ice rafted debris, that are specific to the extent and configuration of the Icelandic Ice Sheet. We therefore only require a realization from which to run the melting model. Hence we believe simplified ice sheet history is sufficient to understand the magmatism in Iceland.

We develop a simple method of reconstructing the volume using published ice sheet model results. We use the results of Patton et al.[4] to identify the ice thickness over the rift zone. Their model results are designed to understand the configuration of the ice sheet during the deglaciation. This time period is also the best constrained in terms of the sea level history[5,6] and deglacial climate[7]. Recognizing that the Patton et al. model[4] is driven by the NGRIP climate record[7] and also that the volume of ice sheets is also contained in sea level histories[5], we use these to constrain ice sheet history prior to the LGM.

We create rift zone loading histories that extend back through the last glacial to the previous interglacial period. We perform a parametric regression on the rift ice load, the NGRIP $\delta^{18}O$ record, and the deglacial sea level record through the deglaciation. We then use the coefficients of the deglacial history to extend the ice sheet load back in time using published sea level records[8-10].

We regress the ice sheet results against the ice core $\delta^{18}O$ [7] and the relative sea level curve [5]:

$$v_{ind} = a_1 \delta_{NGRIP} + a_2 \Delta\zeta_{esl} + c \qquad (1)$$

where we seek an index for the volume of the Icelandic Ice Sheet v_{ind} by solving for the coefficients a_1 and a_2 as well as the offset c. We use the results of Patton et al.[4] from 10 ka to 23.5 ka so that a_1 = -21.94, a_2 = -4.993. Fig. 1 shows how each of the input time series yields the scaled time series. We then apply these coefficients to the remainder of the NGRIP and sea level histories to get the loading from 120,000 years to the present (Fig. 2). Because sea level is not well-constrained through the last glacial period, we utilize three different sea level curves[8-10] to produce the relative ice sheet volume (Fig. 2). We force the melting model with the two extreme ice sheet histories, the M1 model based on the sea level curve ICE-5G[8] and the model M2 based on the sea level curve of Pico et al.[10].

Numerical Melting Model Description and Methods

We solve for the conservation of energy and the vertical percolation of melt within a one-dimensional vertical column. We define a average velocity of the solid and liquid phases[11],

$$\bar{v}=(1-\varphi)v_s+\varphi v_l, \tag{2}$$

where φ is porosity, v_s is the solid mantle velocity, v_l is the melt velocity. All model parameters are listed in Table 1. The change in surface load is assumed to impact the average velocity. We calculate the displacement due to a ice sheet where the change in surface displacement, w_0, with time is given by [12],

$$N\frac{\partial^4}{\partial x^4}\left(\frac{\partial w_0}{\partial t}\right)=\frac{P_{ice}}{\tau_e}, \tag{3}$$

where N is the elastic flexural rigidity, P_{ice} is the load die to the ice sheet, and τ_e is the viscoelastic decay time. The elastic flexural rigidity is given by,

$$N=\frac{ET_e}{12(1-\mu^2)}, \tag{4}$$

where E is Young's modulus, T_e is the effective elastic thickness, and μ is Poisson's ratio. The viscoelastic decay time is defined as [12],

$$\tau_e=\frac{3\eta_s}{E}, \tag{5}$$

where η_s is the viscosity of the lithosphere. Equation 3 is solved using a simple finite element model with linear weighting functions, to solve for w_0 as a function of time due to the change in surface load. The vertical velocity of the mantle below is then updated by the rate of change in displacement due to the surface load, assuming that the displacement decays with depth as a function of the wavelength (width), λ, of the ice load[13],

$$w=w_0\exp\left(-\sqrt{3}\pi\frac{z}{\lambda}\right). \tag{6}$$

Therefore the vertical velocity is given by,

$$\bar{v}=u+\frac{\partial w}{\partial t}, \tag{7}$$

where u is the constant upwelling rate.

To calculate the vertical flow of melt as function of variations in the decompression rate, the conservation of energy is given by,

$$mL+\frac{\partial T}{\partial t}+\bar{v}\frac{\partial T}{\partial z}-\kappa\frac{\partial^2 T}{\partial z^2}=0, \tag{8}$$

where m is the melt production rate, L is the latent heat of fusion, T is temperature, and κ is the thermal diffusion coefficient. The latent heat is given by, $L=T\Delta S/C_p$, where ΔS is the entropy change due to melting and C_p is the heat capacity. The conservation of mass for the liquid phase is given by,

$$\frac{\partial\varphi}{\partial t}+\frac{\partial}{\partial z}(\varphi v_l)=m. \tag{9}$$

To relate the solid velocity to the liquid velocity we turn to Darcy's law,

$$\varphi(v_l-v_s)=\frac{k_0\varphi^n}{\eta_l}\left(\Delta\rho g+\frac{\partial P}{\partial z}\right) \tag{10}$$

where k_0 is the permeability coefficient, n is the exponent on the assumed power law relation between permeability and porosity, $\Delta\rho$ is the density difference between melt and the solid mantle, g is gravity, and P is the pore pressure. We simplify equation 10 by assuming that the compaction length scale is very small, the zero-compaction length approximation[14]. This means that Darcy's law becomes,

$$\varphi(v_l - v_s) = \frac{k_0 \varphi^n}{\eta_l} \Delta\rho g \qquad (11)$$

Furthermore, we substitute v_s with the average velocity (equation 2) to get,

$$v_l - \bar{v} = (1-\varphi)\frac{k_0 \varphi^n - 1}{\eta_l} \Delta\rho g \qquad (12)$$

This form of Darcy's law allows us to substitute for v_l within the conservation of the liquid phase to give,

$$\frac{\partial\varphi}{\partial t} + \bar{v}\frac{\partial\varphi}{\partial z} + \frac{3 k_0 \Delta\rho}{\eta_l}\varphi^2\left(1 - \frac{4}{3}\varphi\right)\frac{\partial\varphi}{\partial z} = m. \qquad (13)$$

Here we assume that $\partial\bar{v}/\partial z \sim 0$.

Equations 8 and 13 are coupled by the melt production rate. We calculate the melting rate from the temperature difference above the solidus. The solidus is a function of water content, depletion, temperature and pressure. In the energy balance we have ignored the loss of heat as the mantle decompresses, but the adiabatic gradient needs to be accounted for within the thermodynamic balance for the melting equations. Assuming the mantle is dehydrated it is calculated as,

$$T_{Sdry} = T_{S0} + \left.\frac{\partial T_S}{\partial F}\right|_P F + \left(\left.\frac{\partial T_S}{\partial P}\right|_F + \frac{\alpha T}{\rho C_p}\right)P, \qquad (14)$$

where $\left.\partial T_S/\partial F\right|_P$ is the solidus depletion gradient, F is depletion, $\left.\partial T_S/\partial P\right|_F$ is the solidus pressure gradient, α is the coefficient of thermal expansion, C_p is the heat capacity, and P is pressure. The solidus is assumed to deepen in the presence of water,

$$T_{Swet} = T_{Sdry} + K\left(D_{H_sO} C_{H_2O}\right)^{\gamma}, \qquad (15)$$

where the coefficients K and γ are from the parameterisation of Katz et al.[15] (their equation 16), D_{H20} is the partition coefficient for water, and C_{H2O} is the concentration of water within the solid mantle. Therefore the melt productivity is given by,

$$m = \frac{\Delta T}{L + \left.\frac{\partial T_S}{\partial F}\right|_P + \left.\frac{\partial T_S}{\partial F}\right|_{H_2O}}, \qquad (16)$$

where $\Delta T = T - T_{Swet}$ is the temperature difference between the mantle and the wet solidus, and $\left.\partial T_S/\partial F\right|_{H_2O}$ is the solidus depletion gradient during the melting of a hydrated mantle. This is calculated using the chain rule,

$$\left.\frac{\partial T_S}{\partial F}\right|_{H_2O} = \frac{\partial T_S}{\partial C_{H_2O}}\frac{\partial C_{H_2O}}{\partial F}. \qquad (17)$$

The change in water composition as a function of depletion is calculated assuming a mass balance between the partitioning of water between the solid and melt phase,

$$\frac{\partial C_{H_2O}}{\partial F} = -C_{H_2O}\frac{1}{D_{H_2O}}(1-F)^{\frac{1}{D_{H_2O}}-2} \qquad (18)$$

and the gradient in solidus with water composition is from equation 15,

$$\frac{\partial T_S}{\partial C_{H_2O}} = \gamma K\left(D_{H_2O} C_{H_2O}\right)^{\gamma-1}. \qquad (19)$$

The kinetic calculation of melt production described in equations 14 to 19 can be unstable for small time steps as temperature is a function of the melt production and the melt production is a function of the temperature. The mantle composition also feeds back into the solidus and hence melt production. If the jump in temperature due to the movement of the solid mantle between time steps is too large the model can become unstable. To improve stability we therefore implemented a simple Runga Kutta scheme to solve equations 14 to 19 once the temperature solution was found.

To track the composition of the melt we assume disequilibrium melting, where the conservation of the solid composition, C_s, is given as[16],

$$\frac{\partial C_s}{\partial t}+v_s\frac{\partial C_s}{\partial z}=\left(\frac{1}{D}-1\right)\frac{C_s m}{1-\varphi}, \tag{20}$$

and the melt composition, C_l, can be written as follows[16],

$$\frac{\partial C_l}{\partial t}+v_l\frac{\partial C_f}{\partial z}=\left(\frac{C_s}{D}-C_l\right)\frac{m}{\varphi}. \tag{21}$$

The melt composition is calculated from the solid composition and knowledge of the partition coefficient D. The partition coefficient is a function of the mineral phase stability [17],

$$D=f_{ol}D_{ol\rightarrow melt}+f_{opx}D_{opx\rightarrow melt}+f_{cpx}D_{cpx\rightarrow melt}+f_X D_{X\rightarrow melt} \tag{22}$$

where f is the proportion of each mineral within plagiolcase, spinel, and garnet peridotite, D_{ol}, etc, are the partition coefficients for Nb and carbon for each phase into melt, and X represents plagioclase, spinel and garnet, respectively. The values for f are taken from Gibson & Geist[18]. For the trace element Nb we use the partition coefficients cited in Gurenk & Chaussidon[19], and for carbon we use the partition coefficients of Rosenthal et al.[20]. The initial mantle composition is assumed to be 60% primitive mantle and 40% MORB[21], The water composition of the solid mantle is advected using equation 20, assuming a partition coefficient of 0.01 for all phases. To account for the possible effects of fractional crystallisation on absolute Nb concentrations, we finally increase the Nb concentration in the erupted melts by 20%.

To convert the predicted concentration of carbon to a flux of CO_2 we assume that the flux of carbon is given by,

$$\Phi_{CO_2}=3.69\,C_C A\rho_l\varphi_{surface}(v_l-v_s) \tag{23}$$

where the factor 3.67 accounts for the conversion of carbon to CO_2, C_C is the carbon concentration in the melt at $\varphi_{surface}$, the upper boundary of the model, and A is the surface area of magmatism, which is assumed to be 30,000 km^2.

We first solve equations 8 and 13, decomposing the diffusion and advection components and using a standard second order accurate explicit finite difference scheme for the diffusion and a flux conservative total variance diminishing (TVD) scheme for the advection term. Once we have solved for porosity and temperature, we calculate the melt and solid velocity using equation 1, and update the solid and melt compositions (equations 20 and 21) using the the same TVD scheme. The code, written in MATLAB, is available at **https://bitbucket.com/johnjarmitage/melt1d-icesheet/**.

Model sensitivity

In order to understand the model sensitivity, we explore the effect of ice sheet width, the rheology of the lithosphere and the rate of decompression (Fig. 3). We explore how changing these five parameters impacts the peak melt production due to a single step change in ice sheet thickness from 2 km to 0 km. By varying the rate of up-welling and the mantle temperature while keeping the other three parameters fixed, we find that melt production increases with decreasing up-welling rate, as the displacement will have a larger effect (Fig. 3a). In 2-D simulations it has however been observed that the width of the zone of partial melting also changes with spreading rate, so countering this 1-D effect[22].

The increase in melt production due to the step removal of the ice sheet is much more sensitive to the mechanical properties of the lithosphere (Fig. 3b). If the elastic thickness and viscosity are low, then the quantity of melt generated is increased by a factor of 2500 % (Fig. 3b) because, for low elastic thickness and viscosity, there is a maximum in flexural displacement. Furthermore, by varying the wavelength of the ice sheet, it can be seen that at wavelengths less than 200 km the geometry of the ice sheet strongly modulates the magnitude of the response, as the displacement due to unloading decreases with depth down the melting column as a exponential function of the wavelength (Fig. 3c). Hence small ice sheets have a smaller impact on melt generation.

Conversion of model porosity to seismic velocity

A critical factor when investigating the effect of surface loading on the magma supply rate is the mantle permeability, with low permeability significantly dampening any signal caused by high-frequency pressure variations at the rift surface due to slow melt transport (Fig. 4). To constrain the permeability for Iceland we examined the effects of varying permeability on the seismic properties of the mantle in our model by converting our predicted 1D thermal structure and porosity to S-wave velocities, assuming that melt reduces the velocity by 7.9% per percent porosity[23] and including the effects of attenuation[24]. Recent joint inversion of teleseismic and ambient noise Rayleigh waves in Iceland would suggest that the S-wave velocity is between 4 and 3.8 kms^{-1} at depths of 50 to 150 km[25] (Fig. 4). We find a significantly better match to the seismic observations for the upper part of the mantle in models where permeability is higher (permeability coefficient of 10^{-5} m^2, equating to average melt transport rates of ~10 m yr^{-1}), which leads to very low melt retention in the zone of partial melting (Fig. 4b). If permeability is lower (permeability coefficient of 10^{-7} m^2, equating to average melt transport rates of ~1 m yr^{-1}), then the models predict a high quantity of melt is retained and the seismic wave speed is reduced to speeds lower than inversion models would suggest is reasonable (Fig. 4c).

Model evolution due to step changes in ice sheet thickness

We perturb the model with a periodic step function of glaciation and deglaciation that has a periodicity of 40 ka (Fig. 5a). We explore the sensitivity of the model to change in the permeability of the mantle, varying the permeability coefficient, k_0, from 10^{-7} to 10^{-5} m^2 (equation 12; red and black lines respectively in Fig. 5). It is found that deglaciation creates a peak in melt eruption rates. The permeability coefficient, which encompasses the effects of both the interconnectivity of the porous network and the viscosity of the melt, controls the velocity of melt percolation. For a permeability coefficient of 10^{-7} m^2 the average melt velocity is 1 myr^{-1}, while for 10^{-5} m^2 the average melt velocity is 10 myr^{-1}.

The peak in eruption rate is sensitive to the permeability (Fig. 5b). The peak occurs sooner and with greater intensity if the permeability is high (10^{-5} m^2; Fig. 5b). This is because the rate of melt percolation is much more rapid if the melt permeability is high[26]. The peak in melt production is diluted by the slower melt transport, and the signal of increased melt production reaches the surface with a delay of a few thousand years if the permeability is 10^{-7} m^2 (Fig. 5b). Interestingly, glaciation acts to gradually shut down melt generation as the signal of surface displacement travels down the 1-D model (Fig. 5b). Carbon fluxes mirror the trend in melt eruption rate (Fig. 5c), with sharp peaks upon deglaciation due to the increased flux of melt. We do not observe a significant delay between the CO_2 flux and eruption rate[26], as we find that carbon partitions into the melt not only at the on-set of melt production. If the permeability is high, 10^{-5} m^2, there is a broad bulge in CO_2 flux that arrives after the peak. This secondary pulse is due to the arrival of deep carbon rich melts percolating upwards as the lower regions of the zone of partial melting recover.

Correlation analysis of Nb observations, ice sheet models, and melt model predictions.

The correlation between the Nb observations and the model ice sheet history is complicated by the irregular spacing of the Nb observations in time. To overcome irregularly spaced data requires some form of data smoothing, which will always improve the apparent correlation. However, it remains informative to attempt to quantify the visual association between the observed reduction in Nb at the end of the Pleistocene and the deglacation predicted to have occurred at 14 and 12 ka.

In order to test for correlation we first smooth and re-sample both the observations and models. We first smooth the data by fitting a generalized additive model (GAM) to data using the R package. This then gives us a smoothed best fit (Fig. 6a). We use the same approach to fit a curve to the ice sheet history M2[10], and the predicted Nb melt concentration assuming an upwelling rate of 20 mm/yr and a permeability coefficient of 10^{-5} m^2 (Fig. 6b and 7). We then test these the two model curves for their correlation to the observed Nb concentrations using Pearson's test. We find that the correlation coefficient between the Nb observations and the ice sheet history M2 is 0.376. For the Nb observations and predictions the correlation coefficient is 0.723.

Table Caption

Table 1: List of model parameters as used within the main text. An asterisk denotes the parameters that were varied in Supplementary Fig. 1, and within the main text. Our preferred values for the Icelandic model are displayed within the table.

Figure Captions

Fig. 1: We use the NGRIP $\delta^{18}O$, relative sea level history[5], and ice thickness results[4] to create a scaled thickness of the Icelandic Ice Sheet through the deglaciation. Our resultant history tends to preserve some high frequency characteristics of the $\delta^{18}O$ record but mutes larger changes such as decrease in ice thickness at the start of the Bølling-Allerød (~14.7ka).

Fig. 2: Reconstructed ice volumes for 120,000 years to present. Here, we use the differing relative sea level reconstructions not infer the relative volume of ice in Iceland. The curves match thought the regression period of 10000-23500 years.

Fig. 3: Plots of % change in melting from steady-state to peak as a function of the model parameters; mantle potential temperature, up-welling rate, lithosphere viscosity, elastic thickness, and the width of the ice sheet. (a) Lithosphere viscosity is set to $10^{21.5}$ Pa s, the elastic thickness is 30 km and the ice sheet width is 100 km. (b) The up-welling rate is 30 mm yr^{-1}, the mantle potential temperature is 1400 °C, and the ice sheet width, λ, is 100 km. (c) The the up-welling rate is 30 mm yr^{-1}, the mantle potential temperature is 1400 °C, and the lithosphere viscosity is $10^{21.5}$ Pa s.

Fig. 4: Profiles of porosity and S-wave seismic velocity for the two model permeabilities of $k_0=10^{-7}$ m^2 black line, and $k_0=10^{-5}$ m^2 red line. (a) Porosity plotted against depth at 0 ka (see Fig 2). (b) S-wave velocity from the joint inversion of teleseismic and ambient noise Rayleigh waves [25] and the predicted S-wave profile from the high permeability case. The model includes the addition of the adiabatic gradient, the anharmonic effects from the mineralogy and effects of attenuation[24]. The

dashed line assumes melt has no effect, the solid line includes a 7.9% reduction in V_S per percent melt[23]. (c) S-wave velocity predictions for the low permeability case.

Fig. 5: Model response to a repeated step change ice sheet thickness. (a) Periodic change in ice sheet thickness assuming a width, λ, of 200 km. (b) Model response in terms of eruption rate. Here the model elastic thickness is 10 km, the viscosity is 10^{21} Pa s, the upwelling rate is 10 mm/yr, and the mantle temperature is 1350°C. The black line is for a permeability coefficient of 10^{-7} m^2, and the red line is for a permeability coefficient of 10^{-5} m^2. (c) Model response in terms of CO_2 flux.

Fig. 6: Comparison between Nb observations and ice sheet history. (a) Observed Nb compositions for the Reykjanes Peninsula[27]; and the Western Volcanic Zone[28]. (b) Predicted ice sheet history for model M2[10]. The observations and ice sheet model have a correlation coefficient of 0.4.

Fig. 7: Comparison between Nb observations and predictions. (a) Observed Nb compositions for the Reykjanes Peninsula[27]; and the Western Volcanic Zone[28]. (b) Predicted Nb concentrations for the model M2[10] ice sheet history assuming a permeability coefficient of 10^{-5} m^2 and upwelling rates of 20 mmyr^{-1}. The observations and model predictions have a correlation coefficient of 0.7.

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