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- Rheological modeling of upper mantle under Iceland
- Low upper mantle viscosity and lithospheric thickness
- Modeling shows large lateral variations in viscosity and lithospheric thickness
- Indications for a dry upper mantle under Iceland
- Increase in lithospheric thickness away from the mid-oceanic ridge axis

Accepted Manuscript

Upper mantle viscosity and lithospheric thickness under Iceland

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Abstract

Deglaciation during the Holocene on Iceland caused uplift due to glacial isostatic adjustment. Relatively low estimates for the upper mantle viscosity and lithospheric thickness result in rapid uplift responses to the deglaciation cycles on Iceland. The relatively high temperatures of the upper mantle under the newly formed mid-ocean ridge under Iceland are responsible for the low upper mantle viscosity values. In this study, estimates for lithospheric thickness and upper mantle viscosity under Iceland from glacial isostatic adjustment studies are complemented by a microphysical modelling approach using the theoretical temperature distribution under mid-ocean ridges combined with olivine diffusion and dislocation creep flow laws. The lithospheric thickness (27-40 km) and upper mantle viscosity (2×10^{18} – 10^{19} Pas) outcomes for the upper mantle recent glaciation under the Vatnajökull glacier are consistent with previous reports of viscosity and lithospheric thickness from glacial isostatic adjustment studies. A combination of a 40 km thick elastic lithosphere and an average upper mantle viscosity of 5×10^{18} Pas would suggest that the upper mantle under Iceland is most likely dry. The earlier and larger Weichselian glaciation event (~10000 BP) on Iceland is predicted to have had a slightly larger upper mantle viscosity $\sim 10^{19}$ Pas and a lithospheric

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25 thickness of ~100 km. Large lateral variations in upper mantle viscosity and especially
26 lithospheric thickness are expected for Iceland perpendicular to the ridge axis.

27

28 **Keywords**

29 Rheology, deformation mechanism, creep, water content, melt content, mid-ocean ridge,
30 plume

31

32 **Introduction**

33 The Vatnajökull glacier, located in the south-east of Iceland is the largest ice cap of Iceland
34 having a mean radius of 50.7 km covering an area of ~8100 km² (Björnsson, 1998). Its
35 estimated mass loss since 1890 is 435 km³ (Árnadóttir et al., 2009), much more than the
36 earlier estimate by Sigmundsson and Einarsson (1992) where the mass loss between 1890 and
37 1978 is estimated at only 182 km³. The Vatnajökull glacier is situated directly on top of the
38 spreading axis in the eastern volcanic zone (EVZ) of the Icelandic mid-ocean ridge and near
39 the inferred centre of the Icelandic hotspot (e.g. Wolfe et al., 1997). Due to the vicinity of the
40 glacier to the active tectonic area, the response of the solid earth to melting of the ice cap is
41 strongly controlled by the properties of the newly formed upper mantle underneath the mid-
42 ocean ridge. The relatively high temperatures in the mantle during rifting result in relatively
43 low upper mantle viscosities and fast relaxation times in comparison with tectonically stable
44 areas such as Fennoscandia. Upper mantle viscosity estimates from glacial isostatic
45 adjustment (GIA) studies for Iceland range between 5×10^{17} Pas and 5×10^{19} Pas (e.g.
46 Sigmundsson, 1991; Sigmundsson and Einarsson, 1992; Thoma and Wolf, 2001; Sjöberg et
47 al., 2004; Fleming et al., 2007; Pagli et al., 2007; Biessy et al., 2008; Árnadóttir et al., 2009;
48 Le Breton et al., 2010) compared to 10^{20} - 10^{21} Pas in Fennoscandia (e.g. Lambeck et al., 1998;
49 Milne et al., 2001). Elastic thickness estimates in those studies including estimates from

50 seismological studies (e.g. Kaban et al., 2002; Kumar et al., 2005) range between a very thin
51 lithospheric thickness of 10 km up to 80 km. Recent glacial isostatic adjustment studies by
52 Pagli et al. (2007) and Árnadóttir et al. (2009) provides estimates for both the lithospheric
53 thickness and the upper mantle viscosity. It resulted in a lithospheric thickness of 10-20 km
54 and a viscosity of $4-10 \times 10^{18}$ Pas for Pagli et al. (2007) and a lithospheric thickness of 40 km
55 and a viscosity of 10^{19} Pas for Árnadóttir et al. (2009).

56 In this study, we present a microphysical approach to determine the lithospheric
57 thickness (elastic layer) and upper mantle viscosity under the Icelandic mid-ocean ridge and
58 specifically under the Vatnajökull glacier. Information on the rheological behaviour of the
59 main upper mantle mineral olivine is combined with information on the temperature
60 distribution under mid-ocean ridges, information on the microstructural state of the upper
61 mantle under mid-ocean ridges (grain size, water content) and stress levels within the upper
62 mantle due to glaciation events to determine the viscosity of the upper mantle. This approach
63 provides detailed estimates on the lateral and radial variations in upper mantle viscosity and
64 lithospheric thickness that can be tested in geodynamical models and compared with glacial
65 isostatic adjustment studies that so far neglected lateral variability in viscosity and
66 lithospheric thickness. In this study, we have based estimates of viscosity and elastic
67 thickness for Iceland on data from the temperature distribution of the mantle under oceanic
68 lithosphere calculated from a half-space cooling model for oceanic lithosphere (Turcotte and
69 Schubert, 2002) and the olivine flow laws from Hirth and Kohlstedt (2003) for linear
70 diffusion creep and non-linear dislocation creep in a wet and dry upper mantle. The half-space
71 cooling model simulates a normal mid-ocean ridge setting and a simplified approximation of
72 plume activity under Iceland. Temperatures of the plume at the center of the ridge axis are
73 $100\text{ }^{\circ}\text{C}$ and $200\text{ }^{\circ}\text{C}$ higher than temperatures at the ridge in a mid-ocean ridge setting, based
74 on estimates of excess temperatures of $75-200\text{ }^{\circ}\text{C}$ due to plume activity under Iceland (e.g. Ito

75 et al., 1996; 1999; Ruedas et al., 2004; Schmeling and Marquart, 2008). The static
 76 temperature distribution in the upper mantle, based on the half-space cooling model, used
 77 here (Fig. 1) is less complicated than dynamic models for plume activity and mantle melt
 78 generation (e.g. Ito et al., 1996; 1999; Ruedas et al., 2004; Marquart, 2001; Schmeling and
 79 Marquart, 2008). However, those models only contain a relatively simple dependency of
 80 viscosity on pressure and temperature, whereas the viscosity dependence of this study
 81 includes also the influence of water, melt fraction and grain size. By using the flow laws of
 82 Hirth and Kohlstedt (2003) for olivine rheology, solid-state viscosity variations (both radially
 83 and laterally) under an active mid-ocean ridge are calculated. The study shows that large
 84 variations in elastic thickness and viscosity are expected under the mid-ocean ridge setting of
 85 Iceland due to cooling-related temperature variations.

86

87 **Temperature dataset and olivine viscosity modeling**

88 Geotherms for the oceanic lithosphere under Iceland are constructed using a time-dependent
 89 half-space cooling model (Turcotte and Schubert, 2002):

$$90 \quad T(z) = T_s + (T_m - T_s) \operatorname{erf} \left(\frac{z}{2\sqrt{\kappa\tau_c}} \right) \quad (1)$$

91 where T_s is the surface temperature, T_m the mantle temperature at a depth of 410 km (in this
 92 study we used $T_m = 1400, 1500$ or 1600 °C), τ_c the age of the oceanic lithosphere and κ the
 93 thermal diffusivity (taken as $1 \text{ mm}^2\text{s}^{-1}$; Turcotte and Schubert, 2002). Estimates of upper
 94 mantle temperatures from Goes et al. (2000) using P- and S-wave tomography data for
 95 Iceland (Fig. 1A) give temperatures in accordance with a time-dependent cooling model of
 96 the oceanic lithosphere. The oldest volcanic rocks outcropping on Iceland have an age of
 97 approximately 16 Ma (Foulger, 2006). The steady-state spreading rate of Iceland in plate
 98 motion models is 19-20 mm/year (e.g. Árnadóttir et al., 2009). Average growth rate of each
 99 plate on both sides of the spreading ridge is thus around $1 \text{ cm/year} = 10 \text{ km/Ma}$. Based on the

100 age constraint and the spreading rate, geotherms for the oceanic lithosphere between 0 and 20
 101 Ma (Fig. 1A) and a two-dimensional cross-section of the temperature distribution under the
 102 Icelandic mid-ocean ridge (Fig. 1B) are constructed. The cross-section encompasses the first
 103 200 km on either side of the ridge axis corresponding to the first 20 Ma of oceanic lithosphere
 104 generation. Mantle temperatures at the base of the modelled range are set at 1400, 1500 and
 105 1600 °C to reflect the temperature distribution in a normal mid-ocean ridge setting ($T_m = 1400$
 106 °C; Turcotte and Schubert, 2002) and settings where a plume with excess temperatures for
 107 Iceland of 100-200 °C (e.g. Ito et al., 1996; 1999; MacLennan et al., 2001; Ruedas et al., 2004;
 108 Putirka, 2005; Schmeling and Marquart, 2008) is imposed on a mid-ocean ridge setting ($T_m =$
 109 1500 and 1600 °C). The olivine-wadsleyite transition at ~410 km is taken as the lower
 110 boundary of the vertical temperature distribution in the upper mantle (Fig. 1B). The dry
 111 solidus of for an anhydrous mantle and the solidus at 100 ppm H₂O for a wet upper mantle of
 112 Hirschmann et al. (2009) are taken as melting curves. When the temperatures along the
 113 geotherms of Fig. 1A exceed the solidus temperature a partially molten upper mantle is
 114 assumed to be present (Fig. 1C).

115

116 *Olivine viscosity modelling*

117 Diffusion and dislocation flow laws for olivine from Hirth and Kohlstedt (2003) are used to
 118 calculate the upper mantle viscosity distribution for a dry/wet and melt-free upper mantle and
 119 a dry/wet and melt-bearing upper mantle. The diffusion and dislocation creep flow laws
 120 describe the dependence of the strain rate ($\dot{\epsilon}$) on temperature (T), grain size (d), deviatoric
 121 stress (σ), pressure (P), water content (fH_2O) and melt content (ϕ):

$$122 \quad \dot{\epsilon} = A \sigma^n d^{-p} fH_2O^r \exp(\alpha \phi) \exp\left(-\frac{E + PV}{RT}\right) \quad (2)$$

123 where A and α are constants, n , p and r are the stress, grain size and water fugacity exponents,
 124 E the activation energy and V the activation volume. For the modeling of dry/wet, melt-

125 bearing upper mantle a melt fraction (ϕ) of $\phi = 0.04$ is used. A melt content of 4% is just
126 above the upper limit for estimates for melt that can remain in a peridotite (up to 2-3%; Faul
127 (1997), 1-3%; Ruedas et al. (2004)). We have used a higher value of 4% to illustrate that the
128 presence of melt has only a relatively small reducing effect on the overall viscosity within the
129 upper mantle under Iceland. A study of the Oman ophiolite (Dijkstra et al., 2002) shows that
130 olivine grain sizes in mantle rocks of mid-ocean ridges have grain sizes between 0.8 and 3.3
131 mm. An average grain size of 2 mm is used in the modelling as a representative grain size for
132 the upper mantle in mid-ocean ridge settings. Stress profiles within the upper mantle due to
133 glacial loading have been calculated for Iceland using a three-dimensional finite element
134 model for glacial isostatic adjustment based on the commercial program Abaqus. We use the
135 geometry of Schotman et al. (2009) who studied Scandinavia, and scaled down the spatial
136 dimensions by a factor of 10 because the Icelandic glaciations cover a much smaller area. The
137 model has a total of 97x97 elements in the horizontal direction and 7 element layers in the
138 vertical direction. Elements in the inner surface area of 292 x 292 km are 8 x 8 km wide.
139 Density and Young's modulus parameters are obtained by volume averaging from PREM.
140 The rheology is a combination of dislocation and diffusion creep (Van der Wal et al., 2010)
141 with parameters taken from Hirth and Kohlstedt (2003) and a temperature-depth profile for a
142 cooling oceanic lithosphere of 4 million years old (Turcotte and Schubert, 2002). The top 15
143 kilometer is taken to be fully elastic, which is reasonably close to the 10 km elastic thickness
144 of Sigmundsson (1991). Note however, that the real elastic thickness in the model for glacial
145 loading depends on the effective viscosity which is a function of stress.

146 The model is loaded separately by two glaciations. (i) the Weichselian (or Younger
147 Dryas) glaciation is modelled following Sigmundsson (1991), approximating the ice cap by
148 four discontinuities in height. The surface extent of the ice cap is 160 km and the maximum
149 height is 1800 m. We have included a 100000 year growth phase of the ice cap ; unloading

150 takes place at different time steps between 10000 and 9500 years before present according to
 151 Sigmundsson (1991). (ii) A model for the recent growing and melting of the Vatnajökull ice
 152 cap is taken from Fleming et al. (2007). A perfectly plastic ice cap (discretised on our 8 x 8
 153 km elements) grows from a radius of 45 km and central thickness of 800 m at 900 AD to a
 154 radius of 52 km and a central thickness of 915 m at 1890 AD. After that melting takes place in
 155 two different rates to end up with a radius of 51 km and 904 m in 2001 AD. The GIA model
 156 provides the von Mises equivalent stress as output:

$$157 \quad q = \sqrt{\frac{3}{2} \sigma_{ij} \sigma_{ij}} \quad (3)$$

158 where σ_{ij} is an element of the stress tensor. Von Mises equivalent stresses are calculated at
 159 the center of the layers (3.75, 11.25, 20, 37.5, 70, 120, 185) and averaged (Fig. 2) over a
 160 horizontal area of 200 km radius and also averaged over time (from 20 ka BP up to 9 ka B.P.
 161 for the Weichselian ice cap; from 10 ka B.P. up to 2001 for the recent melting of the
 162 Vatnajökull ice cap). The stresses are interpolated for every kilometre in the upper mantle to
 163 produce Fig. 2. Using a faster melt rate between 1890 and 2003 (Árnadóttir et al. , 2009) than
 164 reported by Fleming et al. (2007) resulted in average stress values that differed by less than
 165 7% in areas where the stress is high enough to influence the effective viscosity. The resulting
 166 difference in stress using both melt rate estimates is too small to significantly influence the
 167 outcomes of this microphy modelling study. The Von Mises stress is a measure of deviatoric
 168 stress and can be used directly in flow laws derived from uni-axial experiments (Ranalli,
 169 1995). We assume that the equivalent stress can be used as the stress parameter in Eq. (2).
 170 The von Mises stresses is the equivalent stress multiplied by $\sqrt{3}$. The maximum equivalent
 171 stress value of 1 MPa at shallow depths for the Weichselian glaciation (Fig. 2) can be
 172 compared to estimates of an average stress in the lithosphere and upper mantle due to a simple
 173 model of glacial unloading in Scandinavia which are in the range of ~0.5 to at maximum 30

174 MPa (Barnhoorn et al., in press). The smaller ice cap of Iceland in comparison to Scandinavia
 175 causes the average stresses due to glacial loading to be lower for this study. A stress of 1 MPa
 176 is also the lower stress estimate for convective processes in the upper mantle (Ranalli, 1995).
 177 Pressure (P) is calculated as $P = 0.033 \text{ GPa/km}$.

178 A composite flow law (e.g. Ranalli, 1995) in which both competing mechanisms
 179 dislocation creep and diffusion creep contribute to the deformation is used to calculate the
 180 total viscosity where:

$$181 \quad \dot{\epsilon}_{total} = \dot{\epsilon}_{diff} + \dot{\epsilon}_{dist} \quad (4)$$

182 and

$$183 \quad \eta_{dist} = \sigma / 2\dot{\epsilon}_{dist} ; \quad \eta_{diff} = \sigma / 2\dot{\epsilon}_{diff} ; \quad \eta_{total} = \sigma / 2\dot{\epsilon}_{total} \quad (5)$$

184 Two melt contents have been modelled (0% and 4%) for a dry ($C_{OH} = 0 \text{ H}/10^6 \text{ Si}$) and
 185 wet upper mantle ($C_{OH} = 100 \text{ H}/10^6 \text{ Si}$). The lateral extent of the melting zone in the 4% melt
 186 scenario is controlled by the crossing of the geotherm with the solidus of the upper mantle of
 187 Hirschmann et al (2009). Melt-controlled rheology is in this scenario concentrated under the
 188 present-day ridge axis and extends $\sim 120 \text{ km}$ for $T_m = 1400 \text{ }^\circ\text{C}$ away from the ridge axis.
 189 Outside of this melting zone, the upper mantle deforms by a melt-free rheology ($\varphi = 0$). The
 190 lithospheric thickness is defined as the layer in which the viscosity is high enough so that no
 191 measurable influence on the relaxation process is exerted. Analyses of the viscosity cut-off
 192 value for Scandinavia (Barnhoorn et al., in press) shows that viscous deformation starts to
 193 affect predicted sea-level changes by around one meter for an average viscosity in the
 194 lithosphere below 10^{25} Pas . Since relaxation times for the Icelandic ice caps are much smaller
 195 than the relaxation times for Scandinavia (hundreds of years compared to tens of thousands of
 196 years), we have selected a viscosity cut-off value of 10^{23} Pas to represent the maximum
 197 viscosity value at which viscous deformation contributes to glacial isostatic adjustment due to
 198 recent melting on Iceland. We have determined the lithospheric thickness for any distance

199 away from the ridge axis in two ways: 1) as the depth at which the average viscosity from the
200 top surface of the Earth to that depth is less than 10^{23} Pas, 2) as the depth at which the
201 viscosity falls below 10^{20} Pas. We have chosen a maximum viscosity value of 10^{20} Pas since
202 Árnadóttir et al. (2009) state that in their GIA model the lithosphere behaves essentially
203 elastic above 10^{20} Pas. The lithospheric thickness varies in both cases perpendicular to the
204 ridge axis. Below the lithosphere, the upper mantle is deforming viscously and the average
205 viscosity values reported here are the averages of the viscosity values below that lithospheric
206 thickness. The radius of ~50 km for the Vatnajökull glacier present nowadays (Sigmundsson,
207 1991) indicates that roughly the first vertical 100 km of the upper mantle is affected by recent
208 glaciation, whereas the radius of ~160 km present during the Weichselian glaciation
209 (Sigmundsson, 1991) indicates that in that period the first ~300 km of the upper mantle is
210 affected by the Weichselian glaciations. In addition, our stress distribution profiles calculated
211 for both glaciation events show that at depths of 100 km for the recent glaciation event and at
212 300 km for the Weichselian glaciation event the upper mantle is still subjected to some
213 loading induced stresses. At larger depths for both events the stresses decrease towards zero.
214 We therefore have calculated average viscosities from depths below the lithospheric thickness
215 to a depth of 100 km for recent glacial isostatic adjustments on Iceland and from depths below
216 the lithospheric thickness to a depth of 300 km for the glacial isostatic adjustment during the
217 Weichselian period. Maximum depths of the upper mantle affected by the glaciation events
218 are approximations. Establishing the exact depth-sensitivity requires more detailed
219 computations using for example Frechet kernels (Mitrovica and Peltier 1991). Horizontally
220 averaged viscosity profiles and vertically averaged viscosity trends have been produced for
221 both glaciation events. Single values representing the average viscosities and lithospheric
222 thicknesses under both icecaps have been determined for comparison with glacial isostatic

223 adjustment studies on Iceland. Those studies usually assume that the upper mantle under
224 Iceland only has one viscosity and lithospheric thickness without any lateral variations.

225

226 **Results**

227 The rheological modelling for the Iceland mid-ocean ridge shows that diffusion creep
228 dominates the deformation in the upper mantle (Fig. 3). Only at shallow depths under the
229 mid-ocean ridge a small region can occur where dislocation creep dominates the deformation.
230 This is only the case for the Weichselian glaciation event. All deformation during the recent
231 glaciation event is controlled by diffusion creep according to our model. As a result of the
232 dominance of diffusion creep under Iceland most of the upper mantle experiences stress-
233 independent viscosities and therefore the Weichselian and recent glaciation events produced
234 almost identical viscosity fields within the upper mantle (Fig. 4). Significant variations in
235 upper mantle viscosity occur during both glaciation events both with depth and with
236 increasing distance from the ridge axis (Fig. 4). The radial variations in viscosity are
237 characterised by a decrease in viscosity up to depths of ~70 km caused by the rapid increase
238 in temperature with depth and the presence of a partially molten region to depths of ~70 km
239 (Fig.2b). Below that the radially averaged viscosity (Fig. 5) increases from depths >70 km to
240 410 km caused by the increased importance of the pressure dependence of the rheology (Fig.
241 4 and 5), while temperatures are relatively constant at those depths (Fig. 1).

242 The microphysical modelling shows a strong, but narrow low-viscosity layer for the
243 Weichselian glaciation at depths of ~30-70 km (Fig. 6) due to the influence of the melt
244 content on viscosity and the increase in temperature close to reaching the adiabatic
245 temperature gradient. The decrease in viscosity in this layer due to the presence of the melt is
246 ~1-2 orders of magnitude in comparison with the viscosities in a melt-free upper mantle. The
247 low viscosity zone for the recent glaciation starts even shallower (~15 km) due to closer

248 proximity of the Vatnajökull glacier to the mid-ocean ridge axis where higher temperatures
249 prevail (Fig. 6). A wet upper mantle ($C_{OH} = 100 \text{ H}/10^6 \text{ Si}$) results in a reduction of the
250 viscosity of around 1 order of magnitude in comparison with a dry upper mantle (Fig. 5B and
251 5D). The melting temperatures for a wet upper mantle solidus are lower than those for a dry
252 upper mantle (Fig. 1) causing the partially-molten region, and hence melt-induced lowering of
253 the viscosity, to extend to larger distances away from the mid-ocean ridge axis within a wet
254 upper mantle (Fig. 1 and Fig. 5). Plume activity within the upper mantle under Iceland may
255 have resulted in lower upper mantle viscosities than in a normal mid-ocean ridge scenario.
256 Higher temperatures in the range of ΔT of 100-200 °C have been postulated for Iceland (e.g.
257 Ito et al., 1996; 1999; MacLennan et al., 2001; Ruedas et al., 2004; Putirka, 2005; Schmelung
258 and Marquart, 2008). We have simulated the effect of additional plume activity on Iceland by
259 increasing the mantle temperatures at the bottom of time-dependent half-space cooling model
260 (Turcotte and Schubert, 2002) with 100 and 200 °C to $T_m = 1500 \text{ °C}$ or 1600 °C (Fig. 7). The
261 higher temperatures with the mantle due to the plume lower the viscosities by ~ 0.6 orders of
262 magnitude per 100 °C excess temperatures (Fig. 6 and 8).

263 The vertical temperature gradients and resulting viscosity gradients underneath a mid-
264 ocean ridge result in a large increase in lithospheric thickness (Fig. 5A and 5C) with an
265 increasing distance from the ridge axis irrespective of the method used to determine the
266 lithospheric thickness (the depth at which the average viscosity between the top and bottom of
267 the elastic layer falls below 10^{23} Pas or the depth at which the viscosity is smaller than 10^{20}
268 Pas). The temperature gradients due to cooling of the lithosphere away from the ridge axis
269 cause a rapid increase in viscosity in cooled upper mantle material thereby increasing the
270 lithospheric thickness. Melt content in the partially molten region only slightly lowers the
271 lithospheric thickness (at maximum a couple of kilometres, Fig. 5A and 5C). The stress-

272 independent viscosities for Iceland due to the dominance of diffusion creep results in similar
273 lithospheric thicknesses for the Weichselian and recent glaciation events (Fig. 5A and 5C),
274 Fig. 5 shows that defining the lithosphere as the depth at which the average viscosity is
275 lower than 10^{23} Pas produces thicknesses significantly larger than defining the lithosphere as
276 the depth at which the viscosity is lower than 10^{20} Pas (~130-150 km at a distance of 200 km
277 (20 Ma upper mantle from the mid-ocean ridge axis, Fig. 5) versus ~50 km). Close to the
278 ridge axis the viscosity gradient is so high that both methods result in more similar
279 lithospheric thicknesses (differences smaller than 50 km). Excess temperatures due to plume
280 activity lower the viscosity estimates for the upper mantle under Iceland and thus also lower
281 the lithospheric thickness. Lithospheric thicknesses 200 km away from the mid-ocean ridge
282 axis are reduced by ~20 km per 100 °C excess temperatures using the lithospheric thickness
283 definition of average viscosity below 10^{23} Pas and only a reduction of a few kilometres per
284 100 °C excess temperatures (Fig. 8) using the lithospheric thickness definition of a viscosity
285 below 10^{20} Pas.

286 Vertically averaged upper mantle viscosity estimates below the elastic lithosphere in
287 sections from the ridge axis to a distance of 200 km from the ridge axis are fairly constant
288 (Fig. 5B and 5D) for both the Weichselian and recent glaciation events with the lowest
289 average viscosities at the ridge axis. However, averaged viscosities for the recent glaciation
290 event are considerably lower than the averaged viscosities for the Weichselian glaciation
291 event (Fig. 5B and 5D). This is primarily caused by the fact that the upper mantle viscosities
292 for the Weichselian glaciation are averaged from the bottom of the elastic layer to a depth of
293 300 km (maximum depth affected by the Weichselian glaciation), whereas for the recent
294 glaciation event the viscosities are averaged between the bottom of the elastic layer to a depth
295 of 100 km. Thus, Weichselian viscosity averages are dominated by the larger viscosity values
296 of the deeper parts of the upper mantle with only a relatively minor influence of the shallow

297 low-viscosity zone, whereas recent viscosity averages are primarily dominated by the
298 viscosities of the shallow low-viscosity zone (Fig. 6). For a dry upper mantle, average upper
299 mantle viscosities are predicted to be between $\sim 6 \times 10^{18}$ Pas and 1×10^{19} Pas for the
300 Weichselian glaciation event and $\sim 1 \times 10^{18}$ Pas and 8×10^{18} Pas for the recent glaciation event
301 (Fig. 5). In the case of a wet upper mantle, average upper mantle viscosities range are
302 predicted to be between 3×10^{18} Pas and 7×10^{18} Pas during the Weichselian glaciation event
303 and 7×10^{17} Pas and 7×10^{18} Pas for the recent glaciation event (Fig. 5).

304 To compare the results of this microphysical modelling with estimates of constant
305 viscosity and constant lithospheric thickness from glacial isostatic adjustment studies for
306 Iceland we have produced a single upper mantle viscosity value and lithospheric thickness
307 value for each of the two glaciation events (averaging out all lateral variations in viscosity and
308 thickness). We have used the range in lateral extent for the Weichselian glaciation event
309 underneath all of the land mass of Iceland (Sigmundsson, 1991) and the lateral extent of the
310 Vatnajökull glacier for the recent glaciation event to determine the average viscosities and
311 lithospheric thicknesses. Maximum viscosity values in a dry upper mantle under the recent
312 Vatnajökull glacier are around 5×10^{18} Pas and would have been around 10^{19} Pas during the
313 Weichselian glaciation. Plume activity and/or a wet upper mantle scenario reduce average
314 viscosities to be around 10^{18} Pas or lower (Fig. 9). The average lithospheric thicknesses for
315 both glaciations differ depending on which method is used to define the lithospheric
316 thickness. For the Weichselian glaciation, the average lithospheric thickness in a dry upper
317 mantle are either 100 km or ~ 30 km depending on the definition used and ~ 40 km or ~ 10 km
318 for the recent glaciation event (Fig. 9). Additional plume activity or a wet upper mantle
319 scenario reduce the lithospheric thickness estimates considerably.

320

321 **Discussion**

322 The microphysical modelling in this study predicts the lithospheric thickness and upper
323 mantle viscosities under Iceland using a theoretical temperature distribution under mid-ocean
324 ridges combined with the rheological behaviour of olivine. It results in estimates that are in
325 line with estimates from glacial isostatic adjustment studies and seismological studies. All
326 studies show smaller elastic thicknesses (<100 km) and relatively low viscosity values ($< 10^{20}$
327 Pas) in comparison with studies of continental areas such as Fennoscandia. This is due to the
328 relatively high temperature conditions under the newly formed oceanic lithosphere of Iceland.
329 There is a significant spread in upper mantle viscosity estimates for Iceland between 10^{17} Pas
330 and 10^{20} Pas and in elastic lithospheric thickness estimates between 10 and 80 km amongst
331 the different studies (Fig. 10) over the past two decades (Sigmundsson, 1991; Sigmundsson
332 and Einarsson, 1992; Thoma and Wolf, 2001; Kaban et al., 2002; Sjöberg et al., 2004; Kumar
333 et al., 2005; Pagli et al, 2007; Biessy et al., 2008; Árnadóttir et al., 2009; Le Breton et al.,
334 2010). The most recent glacial isostatic adjustment studies that provide estimates of both
335 upper mantle viscosity and lithospheric thickness for Iceland (Pagli et al., 2007; Árnadóttir et
336 al., 2009) are compared with estimates from our study. The best correspondence for the recent
337 glaciation event of both the elastic thickness and average upper mantle viscosity estimates of
338 this study is with the study of Árnadóttir et al. (2009). An average viscosity in the range of
339 6×10^{18} Pas to $\sim 2 \times 10^{19}$ Pas and an elastic thickness of 40 km (Árnadóttir et al., 2009) is only
340 approached by the results of this study for a dry upper mantle scenario without any additional
341 plume activity (average viscosity $\sim 5 \times 10^{18}$ Pas, Fig. 9). All other scenarios produce too low
342 average viscosity values for the recent glaciation event (Fig. 9). Using the depth at which the
343 viscosity falls below 10^{20} Pas as the definition for the lithospheric thickness produces a
344 thickness of ~ 10 km (Fig. 9), which is considerably below the thickness preferred by
345 Árnadóttir et al. (2009). This suggests that using the definition of the depth at which the
346 average viscosity is lower than 10^{23} Pas results in more realistic estimates of lithospheric

347 thicknesses in this microphysical study. Using the preferred dry upper mantle scenario with a
348 lithospheric thickness definition of an average viscosity smaller than 10^{23} Pas results in an
349 average viscosity of 10^{19} Pas and an lithospheric thickness of ~ 100 km for the Weichselian
350 glaciation event (Fig. 9). Our viscosity estimates are in line with estimates of glacial isostatic
351 adjustment studies for the Weichselian deglaciation (Sigmundsson 1991, Biessy et al 2008;
352 Le Breton et al. 2010), although here significantly larger lithospheric thickness values (~ 100
353 km) are predicted for the Weichselian glaciation event than the ~ 15 km lithospheric thickness
354 predicted by Sigmundsson (1991). A correspondence of the upper mantle viscosities estimates
355 of this study can also be achieved with the study for the recent glaciation event of Pagli et al.
356 (2007). A mean upper mantle viscosity of $\sim 10^{18}$ Pas of Pagli et al. (2007) would imply that
357 the upper mantle under Iceland is most likely wet (Fig. 9), as a dry upper mantle produces a
358 higher average viscosity of $\sim 5 \times 10^{18}$ Pas. However, the lithospheric thickness estimates of 10-
359 20 km of Pagli et al. (2007) do not correspond to our estimates of elastic thickness in a wet
360 upper mantle of ~ 30 km. Upper mantle viscosity values in the range of 10^{18} Pas can also be
361 achieved by the presence of a plume under Iceland affecting the viscosities of the mantle. This
362 possibility can not be excluded in the comparison with Pagli et al. (2007), but a plume can not
363 explain the viscosity estimates of Árnadóttir et al. (2009). Upper mantle viscosity estimates in
364 the range of 10^{17} Pas (e.g. Sjöberg et al., 2004) or as high as 10^{20} Pas are not preferred by
365 Árnadóttir et al (2009) since they do not fit with recent GPS uplift data for Iceland. Using our
366 modeling approach, very low viscosities values around 10^{17} Pas can only be produced with
367 either a very wet upper mantle ($C_{OH} \geq 1000$ H/ 10^6 Si) or with very high deep mantle
368 temperatures due to additional plume activity under Iceland ($T_m > 1600$ °C). Upper mantle
369 viscosity estimates around 10^{20} Pas or even higher can on the other hand only be achieved by
370 having very large olivine grain sizes around 5-10 mm in a dry upper mantle, which is much
371 larger than the average grain size present at mid-ocean ridges (e.g. Dijkstra et al., 2002). Since

372 those very low or very high viscosity estimates do not fit with the most recent uplift rate
373 scenarios (Árnadóttir et al., 2009), a wet and/or hot upper mantle scenario or a very large
374 grain size in combination with a dry upper mantle scenario are not likely to be present in the
375 upper mantle under Iceland.

376 By comparing the glacial isostatic adjustments studies of Pagli et al. (2007) and
377 Árnadóttir et al. (2009) with this study using a different approach, similar and relatively low
378 average upper mantle viscosity estimates for the mid-ocean ridge under Iceland can be
379 obtained. However, this study shows that lateral variations in upper mantle viscosity and
380 lithospheric thickness are likely to occur under an active tectonic setting such as the Icelandic
381 mid-ocean ridge. A large heterogeneity in temperature distribution within the upper mantle
382 causes large lateral and radial variability in viscosity. So far, glacial isostatic adjustment
383 studies have neglected any lateral variability in their geophysical models of the upper mantle
384 (e.g. Sigmundsson, 1991; Sigmundsson and Einarsson, 1992; Sjöberg et al., 2004; Fleming et
385 al., 2007; Árnadóttir et al. 2009). This study shows that glacial isostatic adjustment studies of
386 Iceland, particularly concerning the Weichselian glaciation event, should in the future
387 incorporate such lateral and radial variability in viscosity and lithospheric thickness.

388 The modelling of a mid-ocean ridge setting assumes a relatively thin crust of <10 km
389 under Iceland. Different observations of the thickness of the crust under Iceland are reported
390 with thickness varying between a thin crust scenario of 10-15 km and a thick crust scenario of
391 20-40 km (summarised by Björnsson, 2008). If, in the thick crust scenario, the temperature
392 distribution of the thickened crust is the same as the temperature distribution would be in a
393 time-dependent cooling lithosphere, then the mechanically defined elastic lithosphere values
394 reported here would still be valid estimates. If, on the other hand, the thickened crust under
395 Iceland is also significantly cooler, then the lithospheric thicknesses of 25-40 km obtained in

396 this study will be too small. The thin crust scenario, also favoured by Björnsson (2008) for
397 Iceland, is in line with the mid-ocean ridge model used in this study.

398 In conclusion, based on the closest match of upper mantle viscosity and lithospheric
399 thickness between Árnadóttir et al. (2009) and this study, upper mantle viscosity estimates
400 around 5×10^{18} Pas of this study for the recent glaciation event are preferred, implying that a
401 dry upper mantle is most likely present under Iceland. A dry upper mantle scenario is in
402 contrast with the conclusion drawn from a similar rheological modelling study for
403 Scandinavia and where a wet upper mantle was concluded to be most likely present
404 (Barnhoorn et al., in press; Schotman et al., 2009). The different outcomes may point to
405 lateral variations in water content within the upper mantle e.g. between continental
406 lithosphere and oceanic lithosphere.

407

408 **Conclusions**

409 Microphysical modelling using the temperature distribution under mid-ocean ridges combined
410 with olivine diffusion and dislocation creep flow laws with an olivine grain size of 2 mm and
411 stress distributions for the upper mantle produced lithospheric thickness and upper mantle
412 viscosity estimates for the upper mantle under Iceland during the Weichselian glaciation event
413 and the recent glaciation event which nowadays forms the Vatnajökull glacier. Using the most
414 recent estimates of Árnadóttir et al. (2009) as best estimates for upper mantle viscosity and
415 lithospheric thickness, the results of this study that best match the estimates of Árnadóttir et
416 al. (2009) (viscosity $\sim 5 \times 10^{18}$ Pas and thickness of ~ 40 km) suggest that the upper mantle
417 under Iceland is most likely dry. Also, the presence of a plume under Iceland can not explain
418 the recent viscosity values reported for Iceland. Glaciation during the Weichselian glaciation
419 event should have produced viscosity values $\sim 10^{19}$ Pas and a lithospheric thickness of ~ 100
420 km for Iceland based on this microphysical modelling study. The Weichselian viscosity

421 estimate is in line with GIA studies (Sigmundsson 1991, Biessy et al 2008; Le Breton et al.
422 2010). However, this study predicts a significantly larger Weichselian lithospheric thickness
423 due to the lateral variability of the lithospheric thickness in the modelling. Because the
424 deviatoric stress level is mostly below 1 MPa, dislocation creep only dominates in a small
425 area underneath the ridge axis. However, viscosity between the two Weichselian and recent
426 glaciations still differs because of the depth-sensitivity which depends on the size of the ice
427 caps. Large lateral variations in upper mantle viscosity and lithospheric thickness are present
428 perpendicular to the ridge axis with a lithospheric thickness in excess of 100 km at 200 km
429 from the ridge axes.

430

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435

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529 **Figure captions**

530 **Figure captions**

531 Figure 1: A) Geotherms (black lines) for oceanic lithospheres with an age between 0 and 20
532 Ma (Turcotte and Schubert, 2002) using a mantle temperature of 1400 °C. Dry solidus (red
533 solid line) and wet solidus at 100 ppm H₂O (red dotted line) taken from Hirschmann et al.,
534 2009) and mean temperatures (blue symbols) for Iceland from the temperature dataset of Goes
535 et al. (2000; personal communications) are plotted as well. B) Two-dimensional temperature
536 distribution in the upper mantle under mid-ocean ridges between 0 and 200 km (equivalent to
537 0 and 20 Ma using a spreading rate of 1 cm/year) constructed using the geotherms of Fig. 1A.
538 Temperature distribution represents a simplified two-dimensional cross-section of the upper
539 mantle under Iceland. The lateral extent of the Vatnajökull glacier during the recent glaciation
540 event and during the Weichselian glaciation event are indicated on the cross-section. C)
541 Partially molten area (red area) in which the temperatures are above the dry solidus (dark red)
542 and wet solidus (light red) temperatures for the mantle.

543

544 Figure 2: Averaged von Mises equivalent stress for the upper mantle under Iceland at seven
545 different depths for the Weichselian glaciation (filled squares) and recent glaciation (open
546 squares) determined using a three-dimensional finite element model for glacial isostatic
547 adjustment for Iceland. Stresses between 0 and 410 km have been interpolated from the von
548 Mises stress determined in the modelling at the seven different depths.

549

550 Figure 3: Dominant deformation mechanism map for the Weichselian glaciation event in a
551 dry and upper mantle without excess temperatures due to plume activity. Red colours indicate
552 dominance of diffusion creep; blue colour dominance of dislocation creep and brown colours
553 indicate very high viscosities $> 10^{30}$ Pas). Lower stress values during the recent glaciation
554 event caused the absence of any dislocation creep dominance underneath the ridge axis.

555

556 Figure 4: Viscosity distribution under Iceland for a dry and wet ($C_{OH} = 100 \text{ H}/10^6 \text{ Si}$) upper
557 mantle without excess temperatures due to additional plume activity at melt contents of 0 and,
558 4% during the Weichselian glaciation event.

559

560 Figure 5: Elastic thickness (A and C) and mean upper mantle viscosity (B and D) for Iceland
561 for a dry and wet ($C_{OH} = 100 \text{ H}/10^6 \text{ Si}$) upper mantle at melt contents of 0 and 4% during the
562 Weichselian (solid lines) and recent glaciation events (broken lines). Elastic thickness at
563 every distance from the ridge axis is determined either as the depth at which the average of
564 the logarithm of the viscosity from the surface to that depth becomes smaller than 10^{23} Pas
565 (black lines) or as the depth at which the logarithm of the viscosity becomes smaller than 10^{20}
566 Pas (grey lines). Mean upper mantle viscosity at every distance from the ridge axis is the
567 average of the logarithm of the viscosities from the depth of the elastic thickness at that

568 distance to a depth of 300 km for the Weichselian glaciation event and 100 km for the recent
569 glaciation event.

570

571 Figure 6: Horizontally averaged upper mantle viscosity profiles of the upper mantle for the
572 Weichselian glaciation event and the recent glaciation event for melt contents of 0 and 4 % in
573 a dry upper mantle for a normal mid-ocean ridge scenario (A) and for excess temperatures of
574 100 °C and 200 °C (mantle temperature of 1500 °C (B) and 1600 °C (C)) due to additional
575 plume activity (averaged for each depth across the section of the upper mantle under the
576 Weichselian and recent glaciation). The profiles show averages for each depth of all the
577 viscosities present underneath the Weichselian and recent glaciated areas. D) Comparison of
578 viscosity profiles for the different temperature regimes in the absence of melt.

579

580 Figure 7: Viscosity distribution under Iceland for a dry, melt-free upper mantle at excess
581 temperatures of 100 °C and 200 °C (mantle temperatures of 1500 °C and 1600 °C) due to
582 additional plume activity.

583

584 Figure 8: Elastic thickness (A and C) and mean upper mantle viscosity (B and D) for Iceland
585 for a dry upper mantle during the Weichselian (A and B) and recent glaciation events (C and
586 D) for a normal mid-ocean ridge scenario and excess temperatures of 100 °C and 200 °C
587 (mantle temperatures of 1500 °C and 1600 °C) due to additional plume activity.

588

589 Figure 9: Mean lithospheric thickness (mean of all thicknesses under the glaciated areas) and
590 mean upper mantle viscosity (mean of all viscosities under the glaciated areas during the
591 Weichselian and recent glaciation events of Fig. 5 and 7) for various mantle temperatures at
592 the base of the modelled upper mantle in a dry and wet ($C_{OH} = 100 \text{ H}/10^6 \text{ Si}$) upper mantle.

593

594 Figure 10: Compilation of lithospheric thickness and upper mantle viscosity estimates for
595 Iceland from glacial isostatic adjustment studies, seismology and the rheological modelling of
596 this study.

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