

1 **The impact of a 3-D Earth structure on glacial isostatic adjustment in Southeast**
2 **Alaska following the Little Ice Age**

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11 **Key Points:**

- 12
- 13 • The regional 3-D upper mantle viscosity structure is constrained using glacial isostatic
14 adjustment, seismic tomography and mineral physics.
 - 15 • Lateral viscosity variations have no significant impact on the glacial isostatic model predictions
16 in the region.
 - 17 • Non-thermal effects play a dominant role in variations in seismic anomalies in the shallow upper
mantle in this region.

18 **Abstract**

19 In Southeast Alaska, extreme uplift rates are primarily caused by glacial isostatic adjustment (GIA), as a
20 result of ice thickness changes from the Little Ice Age to the present combined with a low-viscosity
21 asthenosphere. Previous GIA models adopted a 1-D Earth structure. However, the actual Earth structure is
22 likely more complex due to the long history of subduction and tectonism and the transition from a
23 continental to an oceanic plate. Seismic evidence shows a laterally heterogeneous Earth structure. In this
24 study a numerical model is constructed for Southeast Alaska, which allows for the inclusion of lateral
25 viscosity variations. The viscosity follows from scaling relationships between seismic velocity anomalies
26 and viscosity variations. We use this scaling relationship to constrain the thermal effect on seismic
27 variations and investigate the importance of lateral viscosity variations. We find that a thermal contribution
28 to seismic anomalies of 10% is required to explain the GIA observations. This implies that non-thermal
29 effects control seismic anomaly variations in the shallow upper mantle. Due to the regional geologic history,
30 it is likely that hydration of the mantle impact both viscosity and seismic velocity. The best-fit model has a
31 background viscosity of 5.0×10^{19} Pa-s, and viscosities at ~ 80 km depth range from 1.8×10^{19} to 4.5×10^{19} Pa-
32 s. A 1-D averaged version of the 3-D model performed slightly better, however, the two models were
33 statistically equivalent within a 2σ measurement uncertainty. Thus, lateral viscosity variations do not
34 contribute significantly to the uplift rates measured with the current accuracy and distribution of sites.

35 **1 Introduction**

36 In Southeast Alaska, Glacial Isostatic Adjustment (GIA) is the dominant process causing present-day
37 vertical crustal motions. Since the early 2000s, GPS observations have highlighted two major uplift areas
38 centered at the Yakutat Icefields (YK) and Glacier Bay (GB), exceeding 30 mm/yr (Hu & Freymueller,
39 2019), as shown in Figure 1. Past studies, beginning with Larsen et al. (2005), have shown that these uplift
40 rates are the result of the Earth's viscoelastic response to the decline of glaciers after the Little Ice Age
41 (LIA), and elastic rebound to present-day ice melt (PDIM). Raised shorelines indicate that uplift started at
42 the end of the 18th century in the Glacier Bay area, and at the end of the 19th century near the Yakutat
43 Icefields (Larsen et al., 2005).

44 To model the Earth's viscoelastic response to changes in ice loading, previous GIA studies (e.g.
45 Elliott et al., 2010; Hu & Freymueller, 2019; Larsen et al., 2005; Sato et al., 2011) have made use of GPS
46 uplift rates to optimize three key parameters in a 1-D Earth model: the effective elastic lithospheric
47 thickness, the asthenospheric viscosity, and its thickness. Larsen et al. (2005) modelled a vertically
48 stratified, non-rotating, self-gravitating, spherical, incompressible Maxwell Earth using the TABOO
49 software (Spada, 2003; Spada et al., 2003). They found that a thin elastic lithosphere (60-70 km) combined

50 with a low-viscosity asthenosphere ($2.5 - 4.0 \times 10^{18}$ Pa-s) were required to fit the observations. Subsequent
 51 studies (Elliott et al., 2010; Hu & Freymueller, 2019; Larsen et al., 2005) added new GPS data, increased
 52 the spatial resolution of the GIA model and updated the ice load model to account for the increasing PDIM.
 53 An overview of their optimal Earth model parameters is seen in Table 1. The maximum spherical harmonic
 54 components (degree and order) in Larsen et al. (2005) are 1024, whereas Elliott et al. (2010) and subsequent
 55 studies increased the spherical harmonics up to degree and order 2048. This way, small ice load changes
 56 and their effects could be resolved with higher spatial detail. In addition, Elliott et al. (2010) used a refined
 57 GPS dataset with higher accuracy and density. This mainly constrained a thinner lithosphere. Sato et al.
 58 (2011) investigated increased PDIM rates using the ice-rate model by Larsen et al. (2007), which resulted
 59 in a larger asthenospheric thickness being required to fit the data. Hu and Freymueller (2019) modeled
 60 higher PDIM rates, derived from the ice-rate model from Berthier et al. (2010). Overall, a thin lithosphere
 61 (50-70 km) underlain with a low-viscosity asthenosphere ($2.5 \times 10^{18} - 3 \times 10^{19}$ Pa-s) is preferred. However,
 62 some areas remain either underpredicted (e.g., the Yakutat Icefields) or overpredicted (e.g., Haines to
 63 Juneau) (Hu & Freymueller, 2019). These discrepancies are likely due to systematic errors in the ice load
 64 model or the Earth model.

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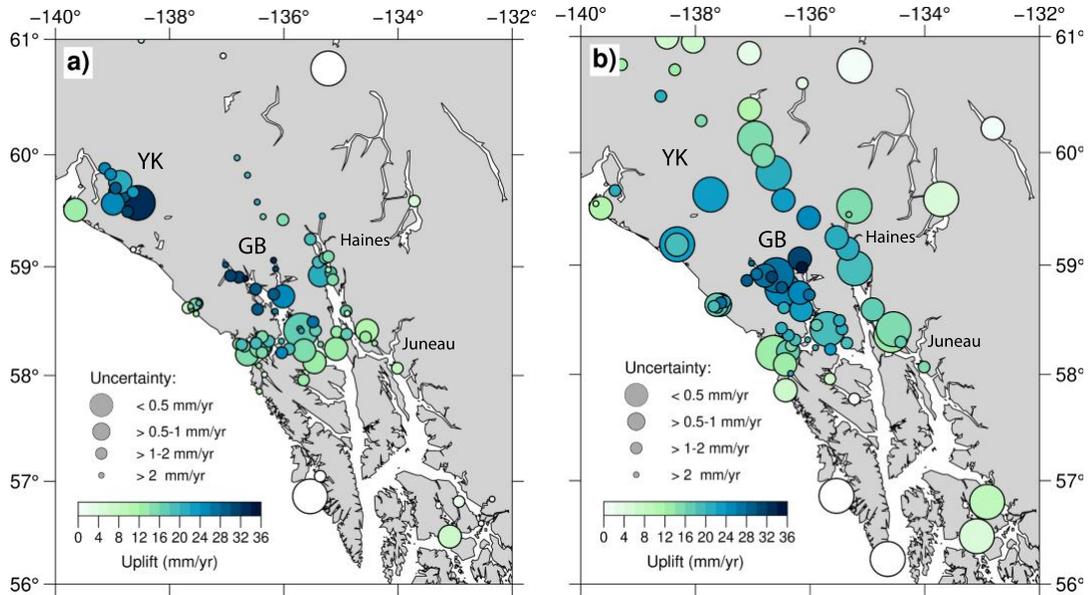
66 **Table 1.** Earth model parameters from various GIA studies in Southeastern Alaska.

Study	Lithosph. thickn. (km)	Asthenosp. thickn. (km)	Asthenosp. visc. (Pa-s)	Upper mantle visc. (Pa-s)	Largest update
Larsen et al. (2005)	60-70	110	$2.5 - 4.0 \times 10^{18}$	4.0×10^{20}	-
Elliott et al. (2010)	50	110	3.7×10^{19}	4.0×10^{20}	GPS
Sato et al. (2011) (2-layer mantle)	54	110	5.6×10^{19}	4.0×10^{20}	GPS + ice model
Sato et al. (2011) (4-layer mantle)	60	160	1×10^{19}	*	GPS + ice model
Hu and Freymueller (2019)	55	230	3×10^{19}	**	GPS + ice model

67 * 4-layer model: viscosities 3, 3, and 4 in units of 10^{20} Pa-s for the upper and lower parts of the upper
 68 mantle, and the lower mantle, respectively. ** set values of the VM5a model: upper part and lower parts
 69 have viscosities of 2.4×10^{21} and 5×10^{21} Pa-s, respectively.

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72

73 **Figure 1.** Average GPS uplift rates (mm/yr) over the periods (a) 1992-2003 and (b) 2003-2012 derived by
 74 Hu & Freymueller (2019). Two uplift peaks with rates >30 mm/yr can be seen at the Yakutat (YK) and
 75 Glacier Bay (GB) icefields.

76

77 The 1-D parameterization in previous GIA studies may not be representative which may affect the
 78 GIA predictions. Studies of other regions have shown that a 3-D structure has a large impact on the GIA
 79 predictions (e.g. Li et al., 2020; Spada et al., 2006; van der Wal et al., 2013). Global tomography studies
 80 (e.g. Schaeffer & Lebedev, 2013) and a regional tomography study for Alaska by Jiang et al. (2018) show
 81 that lateral variations in seismic velocities exist in the region, which likely correspond to variations in
 82 temperature and composition. The actual 3-D structure in Southeast Alaska is more complex due to the
 83 geologic history of tectonism. Before 40-60 million years ago, depending on the plate reconstruction
 84 considered, this region was a subduction zone (Engebretson et al., 1984; Fuston & Wu, 2020; Haeussler et
 85 al., 2003). Since that time, the region has been subjected to shear deformation and substantial margin-
 86 parallel (northward) transport of material via strike-slip faulting (DeMets & Merkouriev, 2016). Offshore
 87 of Southeast Alaska, there is an abrupt transition from continental to oceanic lithosphere across the Queen
 Charlotte fault.

88

89 Seismic anomalies can be converted to viscosity variations. For example, Ivins and Sammis (1995)
 90 used a scaling relationship for the conversion, by relating density anomalies to temperature anomalies based
 91 on the notion that seismic anomalies are caused by temperature variations alone. In reality, non-thermal
 92 effects such as compositional heterogeneity can also affect seismic anomalies. Non-thermal effects can also
 93 play a role in continental regions characterized by iron depletion or in tectonically active regions (i.e.
 characterized by partial melt and/or high water content) (Artemieva et al., 2004). In the GIA studies by

94 Wang et al. (2008) and Wu et al. (2013), the scaling relationship of Ivins and Sammis (1995) is multiplied
95 by a scaling factor, which represents the fractional thermal contribution to seismic anomalies. Wu et al.
96 (2013) found a best fit to GIA observations for the case that thermal effects have a dominant control over
97 seismic anomalies beneath Fennoscandia, with 65% control in the upper mantle which increases with depth
98 into the lower mantle. However, uncertainties related to the relative thermal contributions increase with
99 depth because the GIA process in Scandinavia is mostly sensitive to the upper mantle. Other methods to
100 determine viscosity rely on flow laws and an estimate of mantle temperature (e.g. van der Wal et al., 2013).

101 In this work we will mainly use the method by Wu et al. (2013) to constrain the thermal effect on
102 seismic anomalies by using seismic tomography and GIA due to post-LIA ice load changes. Because of the
103 short wavelength of the ice load, we restrict the investigation to parameters in the shallow upper mantle (<
104 400 km), since crustal velocities are not very sensitive to viscoelastic relaxation in deeper mantle layers
105 (Hu & Freymueller, 2019). A secondary aim of this research is to determine the importance of 3-D viscosity
106 structure for this regional loading problem, considering the high computational cost of 3-D models. The
107 scaling factor, which represents the fractional thermal contribution, is also a measure for the magnitude of
108 viscosity variations. Therefore, the obtained scaling value will reveal to what extent 3-D viscosities play a
109 role in the area. Predictions for 3-D models are compared to those of 1-D models.

110 To summarize, we aim to answer the following research questions:

111

- 112 I. How do shear wave velocities and GIA models constrain viscosity?
- 113 II. What is the thermal contribution to shear wave anomalies in the region?
- 114 III. What is the effect of lateral viscosity variations on GIA model predictions?

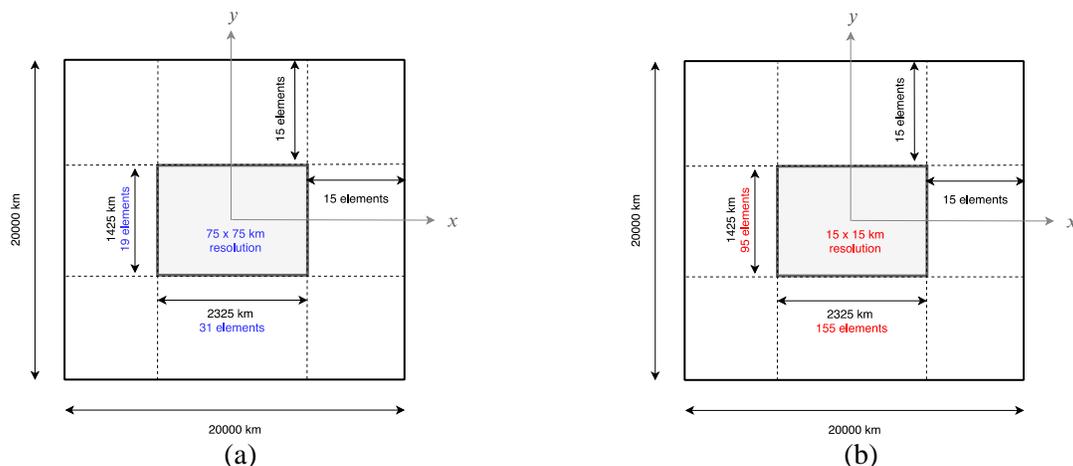
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116 The GIA modelling was performed with a Finite Element (FE) model which allows lateral heterogeneity.
117 Section 2.1 of this paper explains the FE model setup and the Earth model parameters. Section 2.2 briefly
118 describes the ice load model and Section 2.3 describes the method to retrieve viscosities from seismic
119 tomography. The range of possible viscosity values are presented in Section 3.1. In Section 3.2 the model
120 misfits are evaluated using a chi-square test. The role of 3-D viscosity variations is evaluated in Section
121 3.3. An alternative approach to retrieving a 3-D viscosity distribution is through flow laws for olivine,
122 where 3-D variations result from variations in temperature. We evaluate this approach in Section 3.4. To
123 compare our results with earlier studies, which are all based on incompressible earth models, we address
124 the role of material compressibility in Section 3.5. Implications on the Earth structure are discussed in
125 Section 3.6. Model limitations are described in Section 3.7, followed by the conclusions.

126 **2 Methods**127 *2.1 Model setup and geometry*

128 In this research the finite element (FE) method is used to model deformation and stress in the Earth as
 129 implemented in the commercial FE package ABAQUS FEA (Hibbitt et al., 2016), following the approach
 130 by Wu (2004). The GIA model in this research adopts a flat-Earth approximation. The validity of the flat-
 131 Earth approximation was shown in Wu and Johnston (1998) for loads up to the size of the Fennoscandian
 132 ice sheet. Hence, the flat-Earth assumption is reasonable considering the smaller extent of the ice load in
 133 Alaska since the LIA. In addition, material compressibility is assumed and the effects of density
 134 perturbations are neglected. From now on we refer to a compressible model where only material
 135 compressibility is implemented. Moreover, self-gravitation is neglected. Amelung and Wolf (1994) showed
 136 that the effect of neglecting self-gravitation is partly counteracted by the flat-Earth approximation, which
 137 was also confirmed in Spada et al. (2011) and Schotman et al. (2008).

138 The incompressible flat-Earth FE model has been benchmarked against the normal-mode (NM)
 139 model of Hu and Freymueller (2019) (see Supplementary Information). The FE and NM models show good
 140 agreement, where most of the difference are smaller than 1 mm/yr. The largest differences (up to 2.5 mm/yr)
 141 are near the Yakutat Icefields, which are likely due to smoothing of the ice load model in the FE grid and
 142 approximation in the FE models. In addition, tests were performed on the FE model resolution with 10 and
 143 15 km. The NM model uses spherical harmonics with maximum order and degree of 2048 (~10 km). The
 144 10 km FE resolution model did not yield significantly better results than the 15 km resolution test
 145 (differences less than 0.5 mm/yr) and resulted in much longer computational times. For that reason, the 15
 146 km resolution was used in further simulations as it was adequate to represent the observed deformation. For
 147 further details on the model the reader is referred to the Supplementary Information.



148 **Figure 2.** Horizontal geometry. The upper 6 layers have a higher horizontal resolution of 15 km (b), whereas
 149 the lower layers have a lower horizontal resolution of 75 km (a).

150 The model geometry is based on work by Schotman et al. (2009) and Barnhoorn et al. (2011). The
151 loading area consists of 155 x 95 elements with the above-mentioned resolution of 15 x 15 km. Deeper
152 layers (starting from 90 km) have a coarser resolution: 31 x 19 elements of 75 x 75 km. Figure 2 shows the
153 model surface geometry. The total surface area of the model is 20,000 x 20,000 km and the model extends
154 to a depth of 10,000 km in order to minimize boundary effects. In total, 26 finite element layers are created,
155 which gives a total of 198,530 elements. The bottom and vertical edges are prescribed with boundary
156 conditions such that the bottom edge is fixed, and the sides are limited to vertical translation. Winkler
157 foundations (Wu, 2004) are applied at the Earth's surface and internal boundaries where density jumps
158 occur to simulate buoyancy forces.

159 The shallower layers of the model have a higher resolution, whereas the deeper layers have lower
160 resolution. The resolution of these two parts is chosen such that an even number of bricks of the higher
161 resolution part fit exactly on the top surface of the lower resolution part. Considering the mismatch in nodes
162 and the fact that ABAQUS does not provide a standard element to model this, tie constraints are applied at
163 the two surfaces where resolution jumps occur. The tie constraints allow for all active degrees of freedom
164 to be equal for both surfaces. The two surfaces are defined by the upper and lower element surfaces of the
165 two layers, respectively. The outer vertical edge elements are not taken into account as these element nodes
166 already have a fixed constraint.

167 The Earth model parameters are described in Table 2. The density and shear modulus are derived
168 from volume-averaging the PREM model (Dziewonski & Anderson, 1981). The Young's modulus is
169 required as input by ABAQUS and is computed using $E = 2G(1 + \nu)$, where ν is the Poisson's ratio.
170 Durkin et al. (2019) used the LITHO1.0 lithosphere model (Pasyanos et al., 2014) to infer variations in the
171 density and elastic structure and showed that lateral variations in these parameters have a small effect on
172 the elastic uplift. We therefore do not include a heterogeneous density and elastic structure. Material
173 compressibility is incorporated, and the Poisson's ratio varies with depth between 0.26 and 0.30.
174 Compressibility effects on the buoyancy force are not taken into account. This is expected to mainly affect
175 the horizontal displacement (Tanaka et al., 2011), which we do not consider in this study.

176 The upper three layers (depth <40 km) are considered fully elastic. Below 400 km depth, the
177 rheology is only varying with depth. Sato et al. (2011) showed that GIA is less sensitive to deeper viscosities
178 due to the short wavelength of the ice load involved. Therefore, we consider it reasonable to infer viscosities
179 from a global reference model below 400 km depth and the VM5a model (Peltier et al., 2015) was adopted

180 for this purpose. Below 40 km and above 400 km depth, each individual element within a layer is assigned
 181 to an individual viscosity value.

182

183 **Table 2.** Model layers and Earth parameters. Density and Young's modulus are derived from volume-averaged
 184 PREM values. Viscosity below 400 km depth follows the VM5a rheology model. The rheology between 40
 185 and 400 km depth are determined through scaling the seismic model. T.B.D. = to be determined.

Top of layer radius (km)	Layer thickness (km)	Density (kg/m ³)	Rigidity (GPa)	Youngs modulus (GPa)	Poisson's ratio	Viscosity (Pa-s)
6371	12	2171.5	26.6	68.1	0.28	-
6359	14	2885.1	42.8	109.6	0.28	-
6345	14	3380.3	68.1	174.4	0.28	-
6331	15	3378.0	67.9	174.2	0.28	T.B.D.
6316	15	3376.6	67.7	173.5	0.28	T.B.D.
6301	20	3375.2	67.5	172.9	0.28	T.B.D.
6281	20	3372.5	67.1	171.8	0.28	T.B.D.
6261	20	3370.9	66.9	171.0	0.28	T.B.D.
6241	20	3369.2	66.7	170.1	0.27	T.B.D.
6221	20	3365.8	66.3	168.8	0.27	T.B.D.
6201	20	3372.9	67.1	170.4	0.27	T.B.D.
6181	20	3380.0	67.9	172.0	0.27	T.B.D.
6161	20	3416.2	71.5	180.8	0.26	T.B.D.
6141	20	3452.4	75.1	189.5	0.26	T.B.D.
6121	40	3463.7	75.8	190.7	0.26	T.B.D.
6081	50	3486.1	77.1	199.3	0.29	T.B.D.
6031	60	3706.4	92.4	239.5	0.30	T.B.D.
5971	135	3781.5	116.4	302.3	0.30	5.0×10 ²⁰
5836	135	3950.7	117.9	304.6	0.29	5.0×10 ²⁰
5701	250	4443.9	170.1	439.2	0.29	1.6×10 ²¹
5451	250	4590.3	188.5	479.4	0.27	1.6×10 ²¹
5201	430	4780.0	208.8	533.8	0.28	3.2×10 ²¹
4771	430	5008.7	233.7	601.6	0.29	3.2×10 ²¹
4341	430	5227.8	258.4	668.5	0.29	3.2×10 ²¹
3911	431	5444.1	283.4	736.6	0.30	3.2×10 ²¹
3480	3480	10925.0	-	-	-	-

186 *2.2 Ice load model*

187 Upon ice removal, the Earth responds with an instantaneous elastic response and a time-delayed viscous
188 response. The timescale associated with the viscous flow is related to the characteristic relaxation time of
189 the mantle. Due to the presence of low viscosities in the asthenosphere, the associated relaxation times are
190 decades to years, which are comparable to the timescales of recent ice loss and make it difficult to separate
191 elastic and viscous processes. For that reason, both LIA and PDIM glacier variations are modelled together
192 in the ice model.

193 The ice load model is adopted from Hu and Freymueller (2019). We shortly summarize its main
194 characteristics here, but the reader is referred to Hu and Freymueller (2019) for further details. The ice load
195 model is assumed to be a function of space and time, where the ice evolution spans the last 2 ka. In essence,
196 the ice load model consists of three sub-models defined for selected areas: the whole region, Glacier Bay
197 (GB) and the Yakutat Icefield (YK). The last two are necessary because ice mass loss was asynchronous
198 with respect to the regional model. The Glacier Bay ice field experienced a large ice volume loss about 200
199 years ago and the Yakutat Icefields experienced an accelerated ice mass loss during the last two decades.
200 The reader is referred to Figure 5b in Hu and Freymueller (2019) for the time history of the ice loads. The
201 glacier evolution is essentially the same as in Larsen et al. (2005), except for the adoption of the late 20th
202 century ice rate map from Berthier et al. (2010) (here referred to as the Berthier model). The data used for
203 the Berthier model covers the period between 1962 and 2006. In their ice model, Hu and Freymueller (2019)
204 assume these rates to represent the average ice wastages between 1900 and 1995. For PDIM rates (1995-
205 2012) the Berthier model is extrapolated and ice wastage is enhanced by a factor of 1.8 and 2.2 for the
206 periods 1995-2003 and 2003-2012, respectively. PDIM has a contribution to the current uplift rates of up
207 to 45% and 25% for YK and GB, respectively (see Supplementary Figure S6), which is larger than the
208 values found in Larsen et al. (2005). This is largely attributed to the higher PDIM rates modeled.

209 The regional ice load model is given by 677 disks with radii between 10 and 11 km, whereas the
210 GB ice load model is represented by 5 additional disks with radii between 13 and 19.5 km. The disks around
211 YK, which are given by the regional model, are subjected to ice loss rates three times larger than the regional
212 model. This loading is interpolated to the FE grid, while conserving total mass change at each time step.
213 The fraction of each disk covered by a rectangular FE grid element is assigned to a rectangular grid cell.
214 The mesh size needs to be small enough so that errors around the ice load edges are minimized. Our

215 benchmark analysis (see Supplementary Information) showed that a resolution of 15 km was sufficient to
 216 this end.

217 *2.3 3-D viscosity calculations from seismic velocity anomalies*

218 The 3-D viscosity structure is estimated directly through seismic tomography as described in the approach
 219 by Wu et al. (2013). In the crust seismic anomalies are mainly controlled by composition, whereas in the
 220 upper mantle seismic wave anomalies are for a large portion controlled by temperature (e.g., Cammarano
 221 et al., 2003; Goes et al., 2000; Hyndman, 2017). Assuming that temperature variations are responsible for
 222 viscosity anomalies, Ivins and Sammis (1995) introduced a scaling relationship between seismic velocity
 223 anomaly and viscosity that computes viscosity anomalies based on the effect of temperature and mineral
 224 physics. Wu et al. (2013) slightly modified this relationship and included a parameter to scale the viscosity
 225 anomaly based on the thermal contribution. The scaling relationship is given by (Wu et al., 2013):

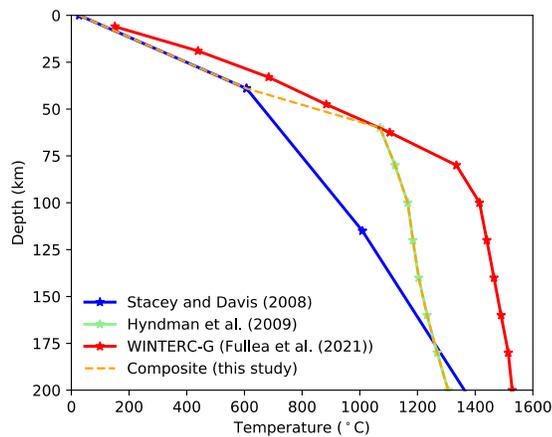
$$226 \quad \log_{10}(\Delta\eta) = \frac{-\log_{10}(e) \beta E + PV}{[\partial \ln v_s / \partial T]_{tot}} \frac{\partial v_s}{v_{s,0}}, \quad (1)$$

227 in which β is a scaling factor ($\beta \in [0,1]$) representing the thermal contribution to shear wave anomalies, T_0
 228 is the background temperature as a function of depth (assumed to be 1-D), E is the activation energy, P is
 229 the pressure, V is the activation volume, R is the gas constant, $\frac{\partial v_s}{v_{s,0}}$ is the fractional shear wave anomaly
 230 computed with respect to the reference seismic anomaly profile $v_{s,0}$ and $[\partial \ln v_s / \partial T]_{tot}$ is the velocity
 231 derivative with respect to temperature accounting for both anharmonic and anelastic effects. Thus 3-D
 232 temperature variations are determined from the seismic velocity variations (scaled by the parameter β), and
 233 viscosity variations are inferred from the temperature variations. The absolute viscosity is then related to
 234 the background viscosity and the viscosity anomalies with $\Delta\eta \equiv \eta/\eta_0$.

235 The seismic anomalies are taken from the global shear wave velocity model SL2013sv (Schaeffer
 236 & Lebedev, 2013). This model defines lateral variations in velocity with respect to a 1-D velocity profile
 237 for the mantle (<1000 km). Although uncertainties associated with the input velocity model can influence
 238 the results (e.g. Yousefi et al., 2021), we consider such uncertainties to be small in comparison to other
 239 uncertainties in the scaling relationship. Shear wave velocity anomalies in Southeast Alaska are dominantly
 240 negative (see Figure 4, upper panels), which result in lower viscosities and thus a weakening effect on the
 241 upper mantle rheology. The velocity derivatives are taken from Table 20.2 in Karato (2008), which
 242 represent global averages and may introduce a bias for Southeast Alaska; anelasticity is expected to play a
 243 larger role in this area due to the higher temperatures involved (Hyndman et al., 2009). Also, if indeed the
 244 mantle is substantially hydrated, the increased water content will enhance the anelasticity effects (Hyndman
 245 et al., 2009). If anelastic effects are not taken into account (or not enough), temperatures could be

246 overestimated and in turn result in lower viscosities. Uncertainties related to the effect of the composition,
 247 water content and partial melt, are not considered here and may also affect the β parameter.

248 The 1-D background temperature profile (T_0) used in Equation 1 is a composite of the globally
 249 averaged profile by Stacey and Davis (2008) and the regional study by Hyndman et al. (2009). Between 0
 250 and 60 km depth, temperatures are taken from Stacey and Davis (2008), between 60 and 200 km the
 251 temperatures from Hyndman et al. (2009) (model C8 in their appendix) are taken and below 200 km the
 252 temperature follows the adiabatic gradient of 0.4 K/km, as shown in Figure 3. A globally averaged profile,
 253 such as the Stacey and Davis (2008) temperature profile results in a too thick elastic lithosphere, whereas
 254 previous GIA studies have shown that a thin elastic lid (50-70 km) is required to fit the crustal deformations
 255 (Hu & Freymueller, 2019; Larsen et al., 2005; Sato et al., 2011). Therefore, the globally averaged profile
 256 is not considered suitable.



257

258 **Figure 3.** Averaged temperature profiles by selected studies. Hyndman et al. (2009) focus on the Northern
 259 Cordillera, whereas WINTERC-G (Fullea et al., 2021) represents a global 3-D temperature and Stacey &
 260 Davis (2008) represent global 1-D temperatures. The WINTERC-G temperature are averaged within the
 261 area bounded by latitudes 58°N-60°N, and longitudes 135°W-141°W.

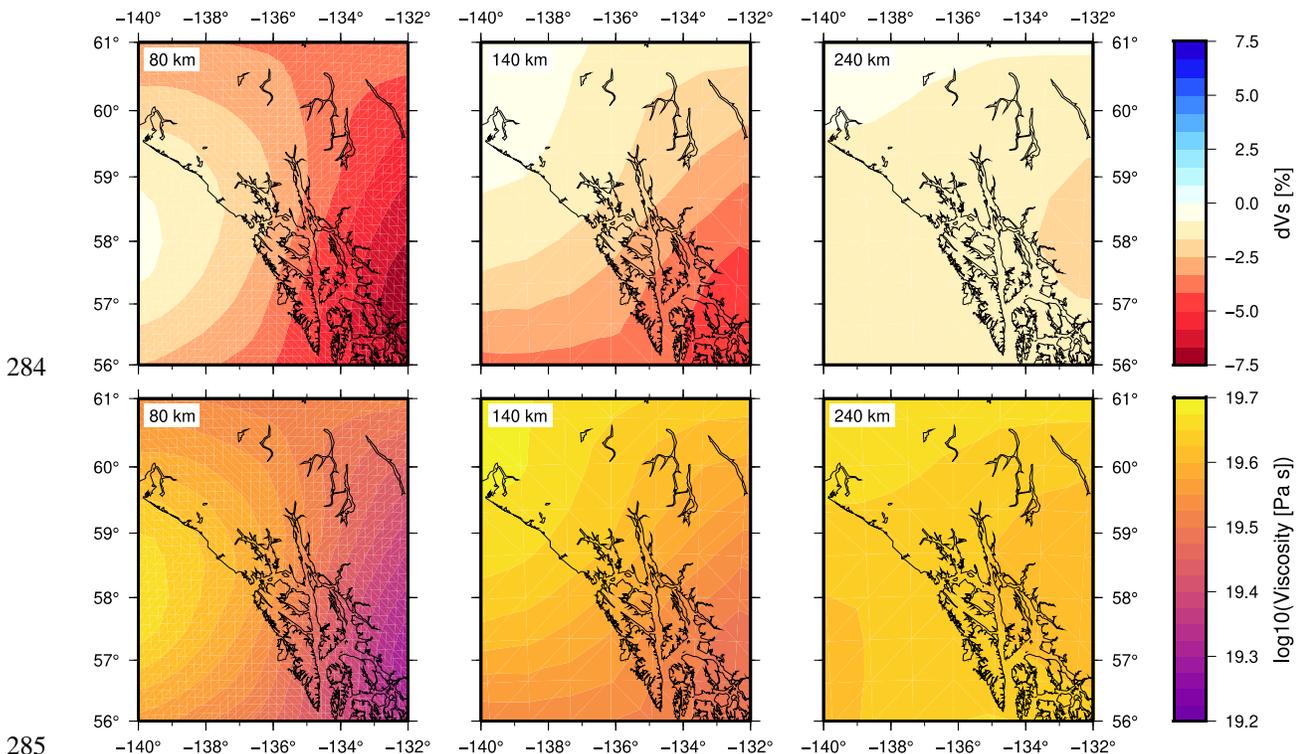
262 The search space consists of the background viscosity profile and the β parameter in the upper
 263 mantle. We limit the search grid to the upper mantle as GIA does not constrain deeper viscosities due to the
 264 short wavelength of the regional deglaciation (e.g., Hu & Freymueller, 2019; Larsen et al., 2005; Sato et
 265 al., 2011). The shear wave anomalies beneath Southeast Alaska are dominantly negative and, consequently,
 266 viscosities will always be lower than the background viscosity profile. Choosing the background viscosity
 267 should be done carefully. The VM5a viscosity structure is not suitable for this area due to its relatively high
 268 viscosity in the upper mantle: 10.0×10^{21} (between 60 and 100 km) and 0.5×10^{22} (between 100 and 420 km)
 269 Pa-s, as opposed to the range ($2.5 \times 10^{18} - 3.0 \times 10^{19}$ Pa-s) for the asthenosphere found in regional GIA studies
 270 (e.g., Hu & Freymueller, 2019; Larsen et al., 2005; Sato et al., 2011). We selected the best-fit earth model

271 of Hu and Freymueller (2019) as the baseline for comparison with our analysis: the elastic lithosphere is
 272 55-km thick and the asthenosphere is 230-km thick with a viscosity of 3.0×10^{19} Pa-s.

273 3 Results and Discussion

274 3.1 Viscosity variations and 1-D averaged profiles

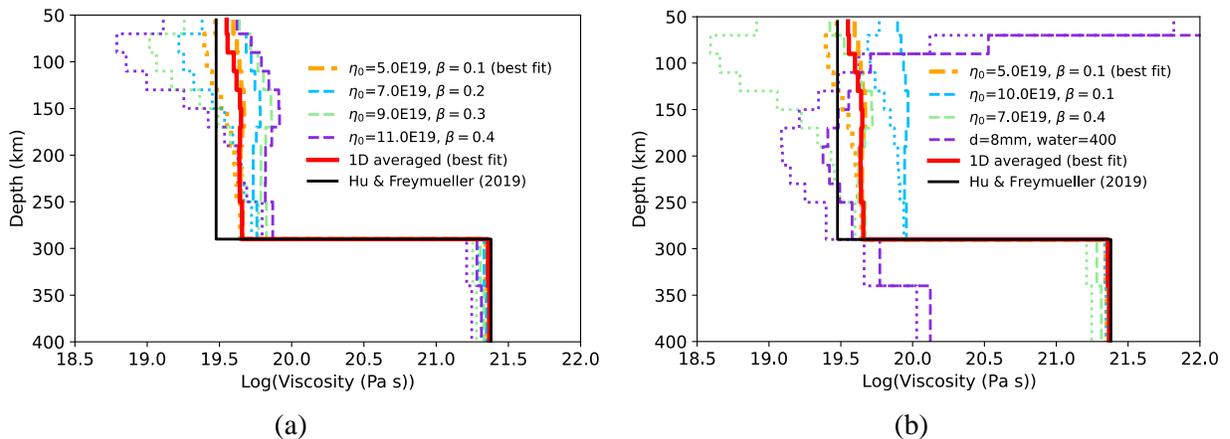
275 The mantle viscosity throughout the volume was computed following Equation 1. Lateral viscosities at
 276 selected depths for the best fit model (described in sections 3.2 and 3.3) are depicted in Figure 4 (lower
 277 panels). The seismic anomalies are directly related to the viscosities; low shear wave velocities correspond
 278 to higher viscosities and vice versa. The largest variations are seen at approximately 80 km and variations
 279 decrease with depth. In deeper layers the viscosity variations decrease as seismic velocity variations
 280 decrease, because we use a constant scaling relation throughout the mantle. In addition, we ran a number
 281 of models using an alternative approach, which entails inferring viscosities from temperature variations
 282 from the global temperature model WINTERC-G (Fullea et al., 2021) through flow laws for olivine (Hirth
 283 & Kohlstedt, 2003).



285
 286 **Figure 4.** Shear wave anomaly and viscosity maps at depths 80 km, 140 km and 240 km. Shear wave
 287 anomalies were linearly interpolated from the SL2013sv model (Schaeffer & Lebedev, 2013). Negative
 288 shear wave anomalies indicate larger temperatures and thus lower viscosities. The largest contrasts are at a
 289 depth of 80 km and lateral variations reduce with depth.

290 Upper and lower bounds on the 3-D viscosity variations with depth are shown in Figure 5a for
 291 several combinations of the adjustable parameters η_0 and β . These profiles illustrate a tradeoff between the
 292 two parameters values for models providing a good fit to the GPS observations, such that the scaling factor
 293 β needs to be larger for models with a higher background viscosity in order to fit the data well. All of these
 294 models have relatively similar 1-D average viscosity profiles, with the lowest viscosities always located at
 295 shallow depth in the southeastern part of the model domain (Figure 4, left panels). However, a lower
 296 background viscosity combined with small lateral viscosity variations results in the best fit.

297 For the profiles shown in Figure 5b, the averaged viscosity is either too low or too high, resulting
 298 in poor prediction of the uplift rates. For example, the green line in Figure 5b has, on average, a lower
 299 viscosity than the best fit model, and the blue line has higher averaged viscosities. All of the models based
 300 on the WINTERC-G model performed poorly. The best-fit profile from this approach is indicated in Figure
 301 5b (purple lines). This approach did not yield good fits because the thick lithosphere imposed by the global
 302 temperature model WINTERC-G is incompatible with the thin lithosphere required in this region (Figure
 303 3). Further details on the results of this approach are discussed in Section 3.4.



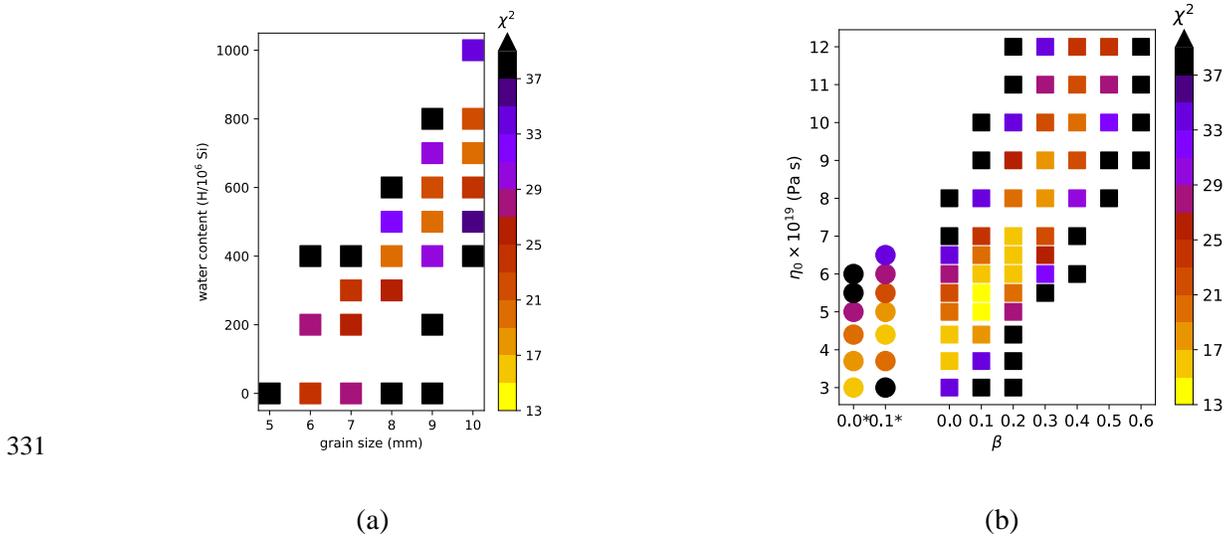
304 **Figure 5.** Viscosity profiles derived from shear wave velocity anomalies for selected background viscosity
 305 (η_0) and scaling factor (β). The viscosity profiles are taken at two points; 133.6°W, 57.3°N (dotted) and
 306 138.8°W, 59.3°N (dashed) representing the minimum and maximum profiles of the 3-D structure. The
 307 profiles in plot (a) represent the combinations along the diagonal in Figure 6b with relatively low χ^2 values,
 308 which have similar average viscosities. The profiles in (b) result in large χ^2 values (except for the orange,
 309 red and black lines). The orange lines show the viscosity profiles of the 3-D best fit model and the red lines
 310 show the averaged 1-D viscosity of the latter model. The purple line in plot b represent the best fit 3-D
 311 model obtained through flow laws for olivine by varying the grain size and water content. The black line
 312 was the result of the best fit 1-D incompressible model in Hu and Freymueller (2019).

313 *3.2 Model performance*

314 The GIA model performance is tested against the GPS rates from Hu and Freymueller (2019), shown in
 315 Figure 1. The vertical uplift is due to the combined effect of GIA due to post-LIA, PDIM, and Pleistocene
 316 glaciations. The impacts of Pleistocene glaciations include contributions of the Laurentide ice sheet (ICE-
 317 3G), glaciers in southern Alaska (Wheeler, 2013) and glaciations of southern British Columbia and
 318 Cascadia (James et al., 2009), and are very small in this region (on the order of 1 mm/yr). Tectonic effects
 319 are expected to be small because of the largely strike-slip tectonics and are not taken into consideration
 320 here, as in previous studies. The uplift rates were computed by Hu and Freymueller (2019) using a normal-
 321 mode GIA model, which may not represent the effects with a FE model with different Earth parameters.
 322 However, the Last Glacial Maximum (LGM) effects are sufficiently small that the overall difference is
 323 assumed to be negligible. The misfits between the observed and predicted GIA rates are evaluated using
 324 the χ^2 test:

$$325 \quad \chi^2 = \frac{1}{N} \sum_{i=1}^N \left(\frac{o_i - p_i}{\sigma_i} \right)^2, \quad (2)$$

326 where N is the number of observations, o_i is the observed GPS rate, p_i is the predicted uplift rate (including
 327 LGM, LIA and PDIM effects) and σ_i the GPS error. GPS observations are available for two periods: 1992-
 328 2003 and 2003-2012. Observations of both periods are combined to compute the best-fit value. Note that
 329 we do not include the error in the LGM, LIA and PDIM models as there is no good error information
 330 available for these models.



332 **Figure 6.** (a) Scatter plot of χ^2 values obtained through using flow laws for olivine. (b) Scatter plot of χ^2
 333 values where the viscosity distribution is obtained directly from shear wave velocity anomalies. The models

334 are compressible with an exception made for (b) where round symbols indicate results from incompressible
 335 models. The best fit model ($\chi^2 = 13.7$) is obtained for the scaled seismic anomalies approach with scaling
 336 factor 0.1 and background viscosity 5.0×10^{19} Pa-s.

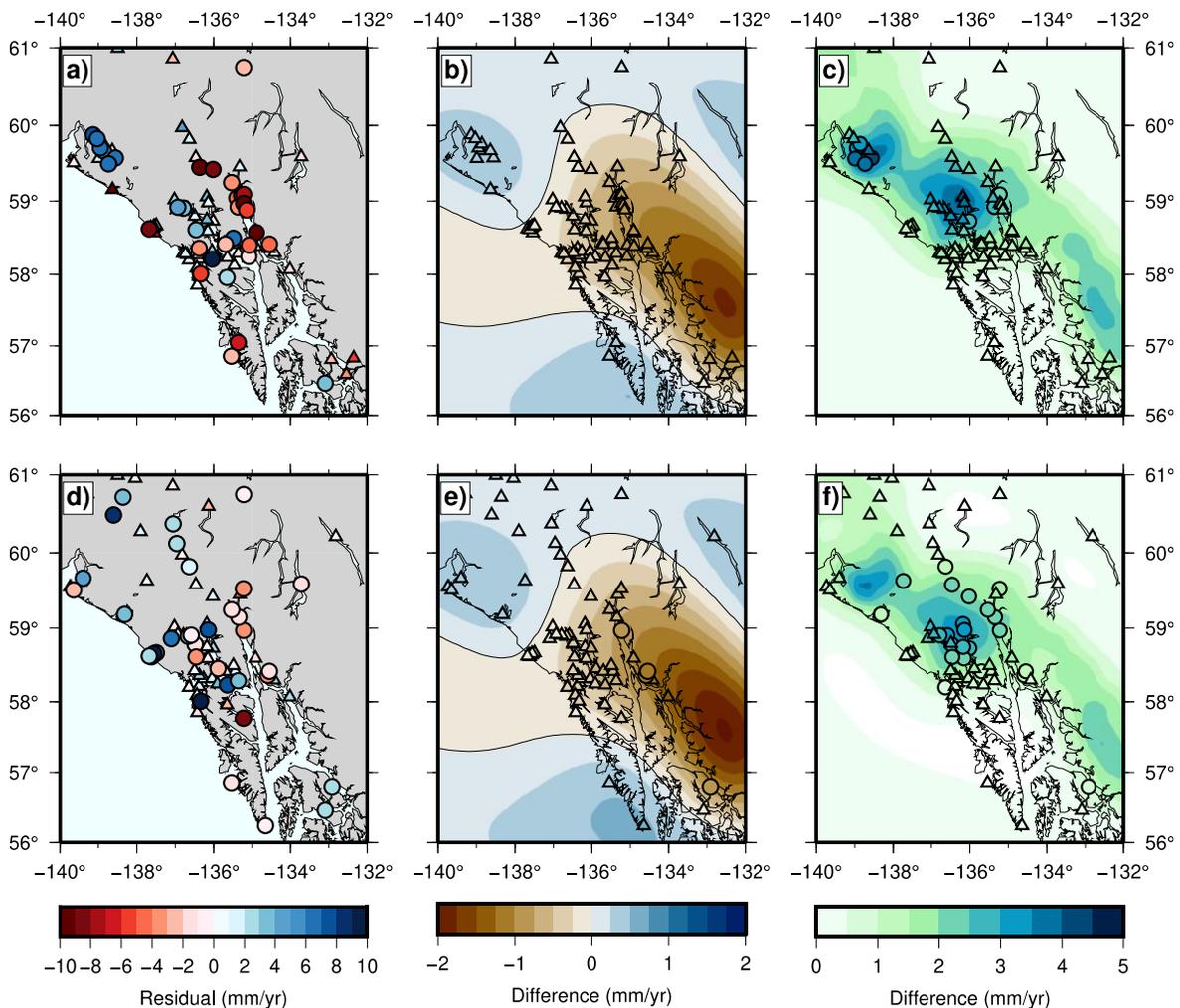
337 The viscosities of the best fit model vary between 1.8×10^{19} and 4.5×10^{19} Pa-s. Viscosity maps at selected
 338 depths are shown in Figure 4 (lower panels). Viscosity variations are largest at ~ 80 km and decrease with
 339 depth. The 3-D viscosities are in the same range found in previous 1-D GIA models (e.g., Hu &
 340 Freymueller, 2019; Larsen et al., 2005; Sato et al., 2011). However, a large portion of the 3-D viscosities is
 341 larger than the previous 1-D GIA models, which is attributed to the compressibility (see Section 3.4). Figure
 342 6b shows that incompressible models prefer a lower background viscosity than compressible models with
 343 the same scaling parameter. Present-day velocities increase for a compressible model in comparison to an
 344 incompressible model with the same earth model parameters. Thus, an incompressible model would require
 345 lower viscosities to achieve similar uplift rates as compressible model. Tanaka et al. (2011) showed that
 346 incompressible models underestimate the viscosity, which agrees with our results. Next, we investigate if
 347 the lateral variations for our best fit model are significant enough to be differentiated from a 1-D model
 348 with the GPS uplift rate data.

349 *3.3 Role of 3-D viscosities in the upper mantle*

350 The optimal background viscosity and the scaling parameter are $\eta_0 = 5.0 \times 10^{19}$ Pa-s and $\beta = 0.1$ (Figure 6b),
 351 respectively. The best-fit model has a fit value of $\chi^2 = 13.7$ with residual mean of 0.3 ± 3.92 mm/yr. The
 352 residuals of the 3-D model are shown in Figure 7a,d. This 3-D model was able to reduce the residuals with
 353 respect to the best-fit 1-D model in Hu and Freymueller (2019); most reductions are located inland and
 354 along the Lynn Canal, the inlet from Haines to Juneau. Because the 1-D model in Hu and Freymueller
 355 (2019) is incompressible, we compare the 3-D model with a "1-D averaged" model, where the viscosity
 356 structure is derived from the best fit 3-D model by averaging the viscosity in each layer over a certain area.
 357 The area is confined to grid cells with prediction rates larger than 15 mm/yr. The viscosity profile can be
 358 seen in Figure 5a, which shows that the 1-D viscosity profile is closer to the upper bound of the viscosity
 359 variations in the 3-D structure. The 1-D averaged model has a fit value of $\chi^2 = 12.1$. We also tested
 360 thresholds of 5, 10 and 20 mm/yr, which resulted in a change of +0.2, +0.3 and 0.0 in the χ^2 value,
 361 respectively. Thus, we conclude that the averaging is not very sensitive to the considered test areas.

362 The differences between the 3-D and 1-D averaged model uplift prediction rates are shown in
 363 Figure 7b,e. The largest differences in the uplift occur in the southeastern corner (up to -2 mm/yr), which
 364 is where the largest variations in seismic velocity and thus viscosity are located. However, there are few
 365 observations available where the largest differences occur. We do see that the 1-D model systematically

366 performs better than the 3-D model along Lynn Canal, the inlet from Haines to Juneau (Supplementary
 367 Figure S4). For the 3-D case, this corresponds to the zone where seismic velocity anomalies were most
 368 negative and hence the viscosity becomes lower than its surroundings at shallow depths (Figure 4). This
 369 results in too large uplift rates for the 3-D model, while the 1-D model predictions are smaller and closer to
 370 the observations. The 3-D model systematically performs better around Glacier Bay; however, the
 371 difference is within the GPS uncertainties. It is possible that errors in the seismic velocity model make the
 372 3-D model fit slightly worse than the 1-D model, or the contribution of temperature variation to viscosity
 373 might not be uniform over the area. Overall, most of the differences are smaller than 2σ , indicating that the
 374 3-D and 1-D models cannot be differentiated from each other (Figure 7b,e). This implies that lateral
 375 variations do not impact the predicted uplift rate significantly.



376

377 **Figure 7.** (a) residuals of the uplift predictions between 1992-2003 of the best fit 3-D model; (b) differences
 378 in uplift (1992-2003) between the best fit 3-D model and averaged 1-D model; (c) difference in uplift (1992-
 379 2003) between the best fit 3-D compressible model and 3-D incompressible model; (d) residuals of the

380 uplift predictions between 2003-2012 of the best fit 3-D model; (e) differences in uplift (2003-2012)
 381 between the best fit 3-D and averaged 1-D model; and (f) difference in uplift (2003-2012) between the best
 382 fit 3-D compressible model and 3-D incompressible model. Triangles indicate that the models cannot be
 383 differentiated within 2σ uncertainty, whereas circles show that they can be resolved. Negative (positive)
 384 values in plots b and e indicate larger (smaller) uplift predictions for the 3-D model. An incompressible
 385 model results in smaller uplift predictions than a compressible model, therefore, plots c and f only show
 386 positive values.

387 *3.4 Comparisons to 3-D viscosities derived from olivine flow laws*

388 An alternative approach to determine the 3-D viscosity variations is by using flow laws for creep olivine.
 389 Here, we use the same methods in van der Wal et al. (2013) for diffusion creep. For this approach,
 390 viscosities are inferred from lateral variations in temperature, water content and grain size are varied, and
 391 a best fit with the GPS data is searched. We use the global temperature model WINTERC-G by Fullea et
 392 al. (2021), of which the average temperature profile beneath Southeast Alaska is seen in Figure 3. Water
 393 content and grain size shift the viscosity profile (the purple line in Figure 5b), whereas the temperature
 394 model determines the shape of the profile and the magnitude of the viscosity variations. For details on the
 395 method and the data, the reader is referred to the Supplementary Information.

396 The fit results (χ^2) are seen in Figure 6a. The plot shows that similar fits can be obtained with
 397 certain combinations of water content and grain size: if we increase the water content (a weakening effect),
 398 then we need to increase the grain size (a strengthening effect). The best fit parameters of 8 mm grain size
 399 and 400 ppm H₂O result in a fit value of $\chi^2 = 20.1$. The best fit value of 20.1 is, however, more than 50%
 400 larger than the values obtained with the scaling of shear wave tomography approach. The uplift pattern (not
 401 shown) indicates that the predictions are flattened and the largest residuals are seen at the uplift peaks. The
 402 larger misfits in this model are due to the thicker elastic lithosphere, which was induced by the WINTERC-
 403 G temperatures (Figure 3) when translating to viscosity. As shown in Figure 5b, the sharp change in
 404 viscosity occurs at a depth of ~100 km, whereas the other models show a thinner lithosphere in which this
 405 viscosity contrast occurs at 55 km. This approach results in larger lateral viscosity variations, where
 406 viscosity contrasts are up to a factor of 4.0 (compared to 2.5 obtained with the shear wave tomography
 407 approach). Nevertheless, the range in lateral viscosities here is rather narrow, which also agrees with the
 408 results we obtained in the seismic approach.

409 *3.5 Role of material compressibility*

410 Previous models assumed incompressibility as well as laterally-homogeneous viscosity. Compressibility is
 411 shown to have an increasing effect on uplift rates in Southeast Alaska (Tanaka et al., 2015). However,

412 differences between compressible and incompressible models can be reduced (to <10%) by adjusting the
413 flexural rigidity of the elastic lithosphere (Tanaka et al., 2011). To isolate the role of compressibility we
414 analyze the outcome of the best fit compressible 3-D model with its incompressible version, where the latter
415 is obtained by adjusting Poisson's ratio to 0.4999. Note that this means that we only consider material
416 compressibility (Klemann et al., 2003) and ignore the effects of internal buoyancy forces. The differences
417 between the incompressible and compressible models can be seen in Figure 7c,f. The differences are up to
418 5 mm/yr, where the largest effects are seen in Glacier Bay and the Yakutat Ice fields, which correspond to
419 the largest uplift rates. The lithosphere in the compressible model can deform more easily than in the
420 incompressible model, leading to higher peak uplift predictions.

421 Figure 6b includes the misfit values of a select number of incompressible models. The best fitting
422 incompressible model has a misfit 11.7% larger than the best compressible model, so the inclusion of
423 compressibility is an important model element. The incompressible models require a lower background
424 viscosity than the compressible models. In other words, the compressibility weakens the material, and we
425 need to strengthen it to achieve the same uplift rate as the incompressible model by increasing the viscosity.
426 The best fit averaged profile (the red line in Figure 5a) shows somewhat higher viscosity values ($\sim 4.2 \times 10^{19}$
427 Pa-s) than the best fit model by Hu and Freymueller (2019) (3.0×10^{19} Pa-s) in the asthenosphere. This
428 implies that incompressible models slightly underestimate the viscosity, which agrees with the results from
429 Tanaka et al. (2015). In addition, the best fit values for the compressible models are lower than the
430 incompressible models, showing that compressibility also improves the fit for the lithosphere thickness
431 used here.

432 *3.6 Implications for Earth structure*

433 Upper mantle temperatures are widely considered responsible for most seismic velocity variations in the
434 upper mantle, whereas compositional effects are thought to have second-order effects (e.g., Cammarano et
435 al., 2003; Goes et al., 2000). However, Goes and van der Lee (2002) also point out that the Western U.S.
436 shows very low seismic velocities, which are likely due to fluids in the mantle introduced by the long history
437 of subduction. In addition, Trampert and van der Hilst (2005) argue that chemical heterogeneity can
438 introduce first-order uncertainties in the conversion from shear wave anomalies to temperature.

439 By using GIA and seismic tomography we were able to constrain the thermal effect on seismic
440 velocity variations. Wang et al. (2008) showed that the thermal effects in Laurentia and Fennoscandia due
441 to seismic anomalies are between 20% and 40%, by assuming a constant scaling factor throughout the
442 mantle and not taking anelasticity effects into account. Wu et al. (2013) found a thermal contribution of
443 65% in the upper mantle, where they applied different scaling factors for the upper and lower mantle and
444 anelasticity was taken into account.

445 We have found a low thermal contribution of 10% to the seismic anomalies, which implies that
446 non-thermal effects control variations in seismic velocities. Although model misfit is similar if both
447 background viscosity and thermal contribution are higher, such models lead to worsened fit relative to our
448 preferred model (Figure 6b). GIA observables in Southeast Alaska are insensitive to the lower mantle
449 structure and only weakly sensitive to deeper layers of the upper mantle (Hu & Freymueller, 2019).
450 Therefore, it is not likely that scaling factors for deeper layers will have a large impact on our results.
451 Uncertainty in $[\partial \ln v_s / \partial T]_{tot}$ can influence our findings. For the global averages we used Table 20.2 from
452 Karato (2008). Uncertainties in $[\partial \ln v_s / \partial T]_{tot}$ are stated to be between 10% and 20% (Karato, 2008). Thus,
453 the scaling factor may be underestimated between 10 and 20%. However, these uncertainties cannot explain
454 the difference found with the global studies that found larger scaling factors.

455 Thus, we are left with the conclusion that non-thermal contributions to seismic velocities are large,
456 and they are likely due to the presence of hydration and/or partial melt. If the mantle is substantially
457 hydrated, anelastic effects would be stronger (Hyndman et al., 2009), and $[\partial \ln v_s / \partial T]_{tot}$ in Equation 1
458 becomes larger, which results in a larger scaling factor. Considering that the region had a long history of
459 past subduction, it is indeed likely that the mantle is substantially hydrated (Dixon et al., 2004).

460 *3.7 Model limitations*

461 There are a number of limitations to the GIA model that are briefly discussed. First of all, the uncertainties
462 regarding the ice load model, both spatially and in time, influence the obtained earth model parameters.
463 These uncertainties are related to both historic and PDIM load changes. The PDIM rates (1992-2012) were
464 constrained by means of comparing GIA predictions with GPS observations in Hu and Freymueller (2019).
465 Scaling of the ice thinning rates (Section 2.2) may not be uniform within Southeast Alaska and select areas
466 may have an asynchronous ice load history with respect to the regional ice load model (such as YK and
467 GB). The ice loading history was optimized for Southeast Alaska and this may not hold for all of Alaska.
468 Moreover, the spatial loading history by Berthier et al. (2010) may be subjected to uncertainties and biases,
469 as discussed in more detail in that study.

470 A second uncertainty relates to limitations in the earth model. The seismic velocity model used
471 here (Schaeffer & Lebedev, 2013) contains uncertainties because of station distribution, and assumptions
472 in the seismic tomography. Seismic stations are sparse across this region, leading to limits in model
473 resolution and the likelihood that details of the velocity structure are not well constrained. Regional seismic
474 models exist (e.g. Jiang et al., 2018), but cover too small an area to have been used in this study. The
475 lithospheric thickness was fixed for our main approach. It is possible that a good fit could be obtained for a
476 different combination of earth model parameters, for example a larger lithosphere thickness, larger
477 background viscosity and larger scaling factor. We investigated an alternative approach based on an olivine

478 flow law, in which lithospheric thickness is not specified a priori and the 3-D variations are derived from a
479 global temperature model based on seismic and gravity data. However, it resulted in worse fits compared
480 to the scaling of seismic anomalies because it effectively imposed a lithosphere that was much too thick.
481 This is likely due to limitations in the data used in the global WINTERC-G model. Another limitation of
482 this alternative approach is that only diffusion creep was modelled. A number of GIA studies have shown
483 that a power-law rheology or composite rheology improved the overall fit to GIA observables (e.g. van der
484 Wal et al., 2013; van der Wal et al., 2010). However, this conclusion cannot be applied directly here, as
485 composite and power-law rheologies lead to lower viscosity which would further raise the modelled uplift
486 rates. Viscosity required for best fit cannot be reconciled with the presence of hydration or standard grain
487 sizes. Furthermore, background stresses could be significant here and other creep mechanisms, such as
488 grain boundary sliding and transient creep could play a role. The latter could play an important role
489 considering the timescales of the ice history. Transient creep has been shown to play a significant role in
490 post-seismic studies in Alaska on monthly to decadal timescales (Freed et al., 2012). However, it is
491 unknown how this plays out with past and current ice load changes.

492 **4 Conclusions**

493 In this study, the shallow upper mantle viscosity structure beneath Southeast Alaska is studied using shear
494 wave tomography and mineral physics in a GIA model for LIA and present-day ice thickness variations.
495 The model is constrained by GPS uplift rates. The role of thermal effects on shear wave velocity anomalies
496 is investigated by using an adjustable scaling factor, which determines what fraction of the seismic velocity
497 variations is due to temperature changes, as opposed to non-thermal causes. If the scaling factor is 0, then
498 there is no thermal contribution to variations in the seismic velocity. Contrarily, a scaling factor of 1
499 indicates that variations in the seismic velocity are only due to thermal effects. The viscosity values are
500 computed using a law that scales seismic velocity anomalies, and relies on a background viscosity. The
501 scaling factor also results in lateral variations in viscosity. By using this aspect, the role of lateral viscosity
502 variations that are expected in this tectonically active region is also investigated.

503 Our best fit model is obtained with a temperature scaling factor of 0.1 and background viscosity of
504 5.0×10^{19} Pa-s. Models with a higher background viscosity and a higher scaling factor gave similar, but
505 worse, misfit. This result implies that the contribution of thermal effects on shear wave velocity variations
506 is small, which implies that seismic anomalies in the shallow upper mantle are mainly controlled by non-
507 thermal effects such as hydration and/or partial melt. The presence of hydration and/or partial melt (Dixon
508 et al., 2004) is consistent with the tectonic history of the region. For the best fit model, the viscosities at a
509 depth of 80 km vary between 1.9×10^{19} and 4.5×10^{19} Pa-s and viscosity variations decrease within deeper

510 layers. The viscosities obtained here are in the same range determined by previous 1-D GIA studies focused
 511 on Southeast Alaska (Hu & Freymueller, 2019; Larsen et al., 2005; Sato et al., 2011).

512 To address the relevance of lateral variations in the viscosity, we have compared the 3-D results to
 513 a radially symmetric model. The 1-D viscosity profile is obtained by averaging the 3-D viscosities in each
 514 Earth layer within a predefined area. The outcome shows that the 1-D model has a slightly better fit,
 515 however, the residuals cannot be distinguished from each other within measurement uncertainties of 2σ .
 516 Therefore, we conclude that 3-D variations do not have significant impact on the predicted uplift, given the
 517 current accuracy and spatial distribution of measurements.

518 **Data availability statement**

519 GPS data are provided in Hu and Freymueller (2019). The shear wave velocity data we use are from the
 520 shear wave tomography model SL2013sv (Schaeffer & Lebedev, 2013). Most of the figures were
 521 prepared using Generic Mapping Tools (GMT) (<https://www.generic-mapping-tools.org/>).

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