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1 **A data-driven model for constraint of present-day glacial isostatic adjustment in North America**

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11 12 **Abstract**

13 Geodetic measurements of vertical land motion and gravity change are incorporated into an *a priori*
14 model of present-day glacial isostatic adjustment (GIA) in North America via least-squares adjustment.
15 The result is an updated GIA model wherein the final predicted signal is informed by both
16 observational data, and prior knowledge (or intuition) of GIA inferred from models. The data-driven
17 method allows calculation of the uncertainties of predicted GIA fields, and thus offers a significant
18 advantage over predictions from purely forward GIA models. In order to assess the influence each
19 dataset has on the final GIA prediction, the vertical land motion and GRACE-measured gravity data
20 are incorporated into the model first independently (i.e., one dataset only), then simultaneously. The
21 relative weighting of the datasets and the prior input is iteratively determined by variance component
22 estimation in order to achieve the most statistically appropriate fit to the data. The best-fit model is
23 obtained when both datasets are inverted and gives respective RMS misfits to the GPS and GRACE
24 data of 1.3 mm/yr and 0.8 mm/yr equivalent water layer change. Non-GIA signals (e.g., hydrology) are
25 removed from the datasets prior to inversion. The post-fit residuals between the model predictions and
26 the vertical motion and gravity datasets, however, suggest particular regions where significant non-
27 GIA signals may still be present in the data, including unmodelled hydrological changes in the central
28 Prairies west of Lake Winnipeg. Outside of these regions of misfit, the posterior uncertainty of the
29 predicted model provides a measure of the formal uncertainty associated with the GIA process; results
30 indicate that this quantity is sensitive to the uncertainty and spatial distribution of the input data as well
31 as that of the prior model information. In the study area, the predicted uncertainty of the present-day
32 GIA signal ranges from ~0.2-1.2 mm/yr for rates of vertical land motion, and from ~3-4 mm/yr of
33 equivalent water layer change for gravity variations.

34
35 **Keywords:** Glacial isostatic adjustment, GRACE, GPS, North America

36 1. Introduction and Previous Modelling Studies

37 Glacial isostatic adjustment (GIA) is the Earth's ongoing long-term (kyr-scale) viscoelastic response to
38 surface loading and unloading by the ice sheets that existed during past glacial cycles. GIA causes
39 deformation of the Earth's solid surface and gravitational potential field, and these deformations in turn
40 result in sea-level changes via the redistribution of water in the global ocean (e.g., Peltier 1974, Farrell
41 and Clark 1976, Peltier and Andrews 1976, Clark et al. 1978, Mitrovica and Peltier 1991). The
42 absolute magnitude of the long-term GIA contribution to present-day observables (crustal deformation,
43 gravity field perturbations, sea-level change) is largest in regions proximal to the former ice sheets.
44 However, at all locations on the globe, ongoing GIA from the last glacial cycle can represent a
45 significant fraction of the total value of observed present-day change. Consequently, constraining the
46 contribution of shorter time-scale processes (contemporary ice mass loss, continental hydrology
47 variations, oceanographic changes) to total present-day rates of crustal deformation, gravity change,
48 and sea-level variation, requires an estimate of the GIA response at present day (e.g., Peltier and
49 Tushingham 1989, Tamisiea and Mitrovica 2011).

50

51 Because glacial isostatic adjustment can seldom be measured directly, the present-day GIA response
52 is often estimated by forward models (e.g., Lambeck et al. 1998, Peltier 2004, Spada et al. 2006,
53 Peltier et al. 2015). Forward modelled GIA is sensitive to several poorly constrained variables,
54 including ice sheet history, elastic lithospheric thickness, the magnitude and parameterization of
55 mantle viscosity, and the effects of lateral changes in Earth structure (e.g., Tushingham and Peltier,
56 1991, Lambeck et al. 1998, Latychev et al. 2005, van der Wal et al. 2010, Tamisiea 2011, Peltier et al.
57 2015), although forward model predictions typically lack any formal quantification of the uncertainties
58 associated with the input model combinations. However, variation of input model parameters within a
59 reasonable range of values can result in significant changes to the magnitude (and sometimes the
60 sign) of the predicted GIA response, indicating GIA uncertainty is large. This observation holds even in
61 far-field regions, which are characterized by much smaller GIA signals than near-field regions. For
62 example, Mitrovica and Davis (1995) found that estimates of the GIA contribution to far-field sea-level
63 change varied by as much as $\sim 0.3\text{-}0.5$ mm/yr for a range of GIA models, a value which represents

64 ~10-20% uncertainty in the GIA contribution relative to their mean total far-field sea-level rate
65 estimated from tide gauge measurements (~1.4 mm/yr).

66

67 As datasets from satellite geodesy missions have increased both in quantity and duration of
68 observation, increasing emphasis has been placed on the use of data-driven methods to constrain
69 better the individual components of total measured present-day change rates (Riva et al. 2009, Hill et
70 al. 2010, Wu et al. 2010a, Rietbroek et al. 2012, Sasgen et al. 2012, Lambert et al. 2013, Wang et al.
71 2013, Zhao 2013, Gunter et al. 2014, Wang et al. 2015). A main limitation of these types of models is
72 that they typically focus on present-day GIA signals, and therefore offer little insight into the time-
73 varying GIA response or ice sheet evolution. However, while the method, study area, and quantity of
74 primary interest vary by study, all of these studies either eliminate or reduce the uncertainty associated
75 with forward modelled GIA (through the use of separation approaches, or data-driven inversion
76 approaches, respectively).

77

78 In North America, separation approaches that use a combination of GPS measurements and
79 observations from the Gravity Recovery and Climate Experiment (GRACE) have been employed to
80 estimate recent continental hydrology changes (Lambert et al. 2013, Wang et al. 2013, 2015).
81 Although Lambert et al. (2013) and Wang et al. (2013, 2015) use different methodologies, both
82 methods assume a relationship between GIA-induced changes to vertical land motion and gravity
83 change that can be used to separate and remove the GIA effect from total measured rates (e.g., Wahr
84 et al. 1995), and thus avoid the use of forward modelled GIA predictions. Data-model combination
85 approaches involving the simultaneous adjustment of geodetic measurements with *a priori* forward
86 modelled GIA information have been applied globally (Wu et al. 2010a), in North America (Sasgen et
87 al. 2012, Zhao 2013), Antarctica (Sasgen et al. 2013), and Fennoscandia (Hill et al. 2010). These data
88 combination approaches yield updated models of present-day GIA informed by both observational
89 data and prior expectation of GIA motions derived from models, although in the North American
90 studies, the focus was not placed on quantifying GIA uncertainty. The methodology of Hill et al. (2010)
91 was also used to obtain the GIA model used for the Stable North American Reference Frame
92 (SNARF) (<https://www.unavco.org/projects/past-projects/snarf/snarf.html>).

93

94 In this study, we extend the data-driven combination method of Hill et al. (2010) to obtain a prediction
95 of present-day GIA in North America. Relative to the SNARF project, which used a similar
96 methodology, we include GRACE data, as well as use updated vertical land motion data and an
97 updated North American ice sheet reconstruction to generate the prior GIA information. We also use
98 variance component estimation to weight the contributions of the data and prior input to the final model
99 prediction. Our goal is to obtain a present-day GIA solution for the study region that adequately
100 predicts available observational constraints, minimizes the uncertainty associated with the forward
101 modelled GIA inputs, and includes a realistic estimation of formal model error.

102

103 **2. Methodology**

104 The GIA response is solved for by least-squares adjustment, following the methods described by Hill
105 et al. (2010). The final predicted GIA model response is represented by vector m^* , where the response
106 represents the GIA-related deformation type(s) of interest. Here, the predicted deformation types are
107 rates of vertical crustal motion and gravity change. A solution for m^* is obtained by minimizing the
108 objective function of the data misfits and the *a priori* model misfits (e.g., Tarantola 2005)

$$109 \quad \varphi(m^*) = (d - A \cdot m^*)^T C_d^{-1} (d - A \cdot m^*) + (m^* - m)^T C_m^{-1} (m^* - m), \quad (2.1)$$

110 where d is a vector of GIA-induced observations, A is the design matrix, C_d is the data covariance
111 matrix, m is a vector of *a priori* GIA predictions, and C_m is the prior model covariance matrix.

112

113 *2.1 Observational Inputs*

114 The observation vector d contains N measurements of GIA-related observations. In this study,
115 depending on the combination of data that is inverted, d contains observed vertical land motion rates,
116 GRACE-measured gravity change rates, or both. N is the total number of input observations used to
117 constrain the solution. For example, if n_{GPS} vertical land motion data and n_{GRACE} GRACE data points
118 are inverted simultaneously, then $N = n_{GPS} + n_{GRACE}$. The data covariance matrix C_d is an $N \times N$

119 matrix containing the covariances associated with the observations. The component of C_d associated
 120 with the vertical land motion data is assumed to be diagonal (variances only), while the component of
 121 C_d associated with the GRACE gravity data includes the full covariance matrix of the trend. The data
 122 are described further in Sections 3.1-3.3.

123

124 2.2 Model Inputs

125 The *a priori* model vector m contains the mean of a suite of forward-modelled GIA predictions. Each
 126 model deformation type is predicted at each of the input observation sites, as well as on a grid of the
 127 study area. The length of vector m is thus the sum of the total number of predictions at observation
 128 sites and the total number of predictions at grid locations, or $M = M_{obs} + M_{grid}$. For two model
 129 deformation quantities (vertical motion and gravity change) and a grid of n_{grid} locations, $M_{grid} = 2n_{grid}$
 130 and $M_{obs} = 2N$.

131

132 There are no formal uncertainties associated with forward GIA models. However, for a suite of GIA
 133 models that spans a reasonable range of parameter space, an input model covariance matrix can be
 134 constructed using

$$135 \quad C_m^{ij} = \frac{1}{\Omega} \sum_{k=1}^{\Omega} (m_i^k - m_i) (m_j^k - m_j) \quad i, j = 1, \dots, M, \quad (2.2)$$

136 where $k = 1, \dots, \Omega$ represents the suite of Ω forward models, m_i^k is the model prediction at the i^{th} spatial
 137 location in the k^{th} GIA model, m_i is the average prediction of Ω models at the i^{th} location, and here i
 138 and j are the indices of the model covariance matrix. The *a priori* model averages and the associated
 139 suite of GIA model predictions are discussed further in Section 3.4.

140

141 2.3 Design Matrix

142 The $N \times M$ design matrix A consists of the partial derivatives of the observations with respect to the
 143 model parameters according to

144
$$A_{ij} = \left. \frac{\partial d_i}{\partial m_j^*} \right|_{m^*=m} \quad i = 1, \dots, N; j = 1, \dots, M. \quad (2.3)$$

145 If the location of the i^{th} observation coincides with the location of the j^{th} model prediction, and the
 146 observation type is equal to the model prediction deformation type, then the design matrix has a value
 147 of 1. The design matrix has a value of 0 at all other locations (that is, all locations in i and j where the
 148 site location and deformation type are not coincident). In this study, the deformation type is either
 149 vertical motion or gravity change.

150

151 *2.4 Model Solution*

152 Minimizing the objective function in equation 2.1 with respect to the model solution by setting $\frac{\partial \varphi}{\partial m^*} = 0$
 153 yields the model solution (e.g., Tarantola 2005)

154
$$m^* = m + (A^T C_d^{-1} A + C_m^{-1})^{-1} A^T C_d^{-1} (d - Am), \quad (2.4)$$

155 with a posterior covariance matrix of

156
$$C_{m^*} = (A^T C_d^{-1} A + C_m^{-1})^{-1}. \quad (2.5)$$

157 Both the GPS and GRACE datasets are incorporated into the model solution together, as well as
 158 independently, for a total of three scenarios. Each dataset is weighted relative to the prior information
 159 (and in the case of the combined solution, relative to each other) using variance component estimation
 160 (VCE) following the description of Koch and Kusche (2002). When weighting is applied to the solution,
 161 equation 2.4 takes the form

162
$$m^* = m + \left(\frac{1}{\sigma_d^2} A^T C_d^{-1} A + \frac{1}{\sigma_\lambda^2} C_m^{-1} \right)^{-1} A^T \frac{1}{\sigma_d^2} C_d^{-1} (d - Am) \quad (2.6)$$

163 where σ_d^2 is the variance factor applied to the data and σ_λ^2 is the variance factor applied to the prior. If
 164 both datasets are incorporated into the model solution, then each dataset has its own variance factor.
 165 The influence of the prior model on the model solution is inversely proportional to the variance factor,
 166 σ_λ^2 ; that is, the influence of the prior information decreases as σ_λ^2 increases. Further discussion of the
 167 variance component estimation is found in Section 4.3 and the supplementary material.

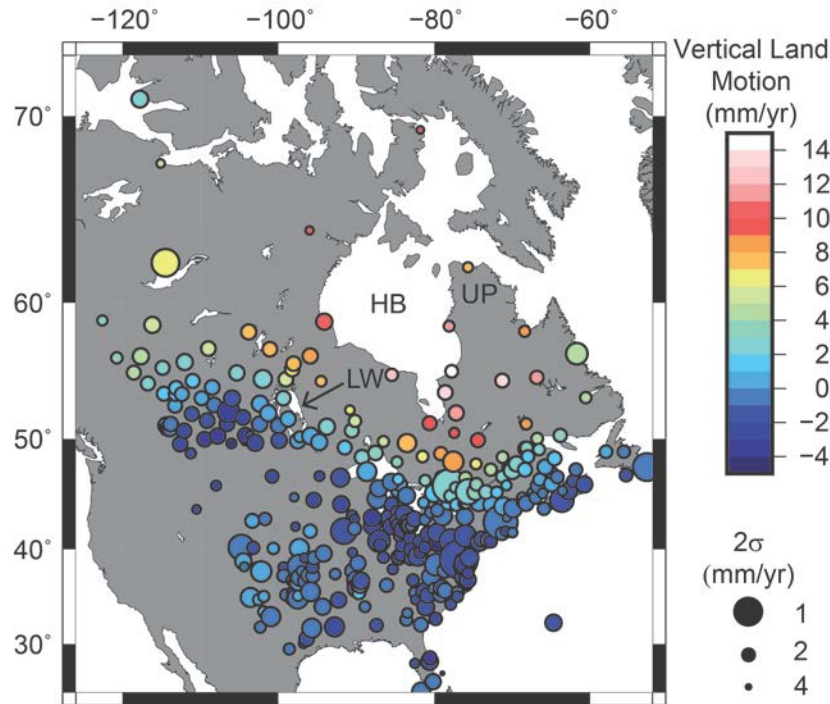
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169

170 **3. Observational Data and Model Inputs**

171 *3.1 Vertical Land Motion Data*

172 Rates of vertical land motion are obtained from the solution in the supplementary material of Peltier et
173 al. (2015). Only rates from within the North American study area are selected. Peltier et al. (2015) did
174 not include sites from westernmost North America in their supplement, possibly due to the potentially
175 significant contribution to the measured rate from tectonic deformation. Several studies have for
176 example suggested that in addition to GIA, west coast sea-level data may include signals from tectonic
177 deformation (e.g., Clague et al. 1982, James et al. 2009a, Roy and Peltier 2015) and that regional GIA
178 models may require decreased upper mantle viscosity values to best fit sea-level data there (e.g.,
179 James et al. 2009b). Therefore, to avoid the influence of active tectonics, and similar to other studies
180 (Sella et al. 2007), GPS data from the western part of the continent are excluded here. We also
181 exclude locations where there may be contributions to vertical motion from present-day mass loss of
182 glaciers and ice caps (Alaska, and islands within the Canadian Arctic Archipelago) (Arendt et al. 2002,
183 Gardner et al. 2011, Jacob et al. 2012), and sedimentation and subsurface fluid withdrawal (Gulf of
184 Mexico coastline) (Ivins et al. 2007, Kolker et al. 2011, Letetrel et al. 2015). 340 measurements
185 comprise the dataset of vertical land motion rates (Figure 1). We additionally remove the effect of
186 hydrological loading from the vertical velocities (Section 3.3).



187

188 **Figure 1.** Vertical land motion rates in North America used as input for the inversion, from the solution of Peltier et
 189 al. (2015). Measurements from sites significantly influenced by non-GIA processes are removed, as discussed in
 190 Section 3.1. Uncertainties are shown as 95% confidence limits (2σ); symbol size is inversely proportional to
 191 uncertainty size. HB – Hudson Bay, LW – Lake Winnipeg, UP – Ungava Peninsula.

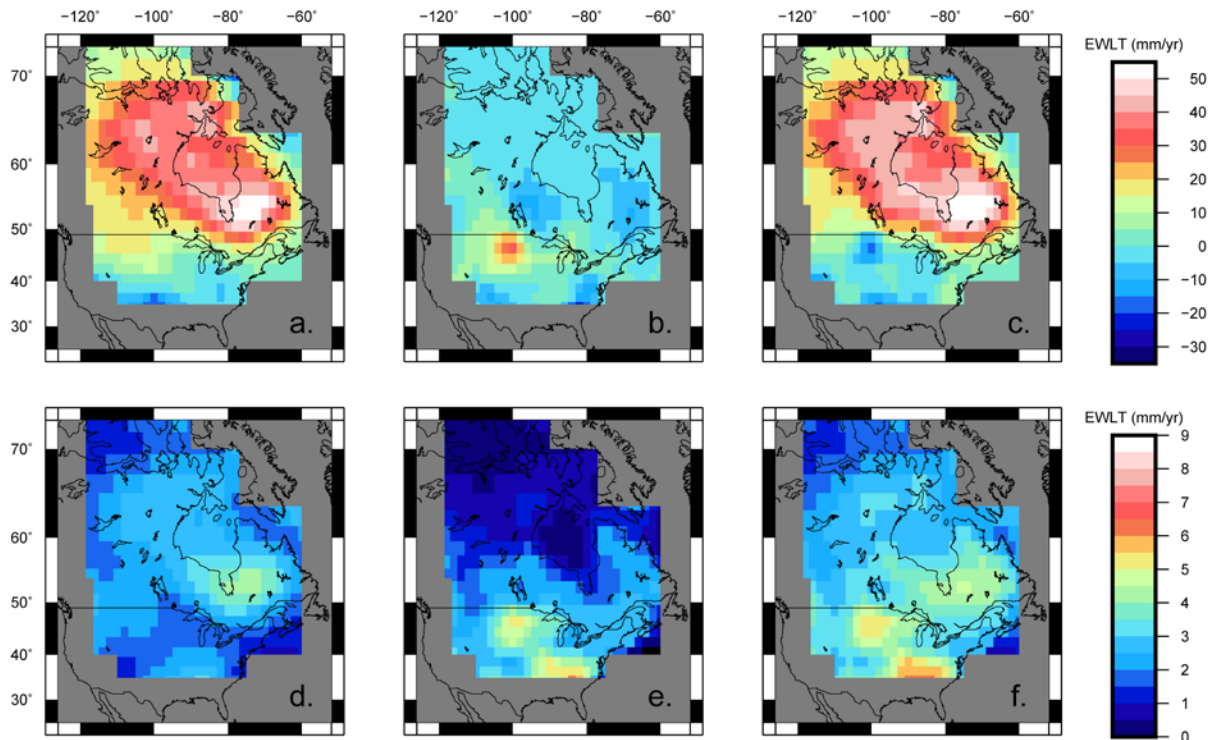
192

193 The vertical motion data consist primarily of measurements from continuous GPS sites, as well as
 194 rates from Canadian Base Network (CBN) episodic GPS sites, although a limited number (<5%) of
 195 velocities are also estimated using VLBI, SLR and DORIS observations (Peltier et al. 2015). The
 196 continuous GPS data span a time period from 1994 to 2012, with the lengths of the individual time
 197 series ranging from approximately 4 to 16 years, while the episodic CBN rates are derived from four
 198 occupations between 1996 and 2011. The data are aligned in the International GNSS Service (IGS)
 199 reference frame IGS08, which is a realization of the International Terrestrial Reference Frame solution
 200 ITRF2008 (Altamimi et al. 2011). The vertical velocities are represented in the centre of mass (CM)
 201 frame. The data processing for the vertical motion solution is discussed further in Peltier et al. (2015)
 202 and Argus et al. (2014). In Peltier et al. (2015), the uncertainties of the vertical motion rates are
 203 expressed as 95% confidence limits (2σ). The input data covariance matrix C_d of the vertical motion
 204 measurements is represented by a diagonal matrix of the associated variances (σ^2).

205

206 3.2 GRACE Data

207 Gravity change rates are estimated using GRACE Release-05 (RL05) monthly solutions from the
208 University of Texas at Austin Center for Space Research (CSR). The coefficients are truncated at
209 degree and order 96, consistent with a spatial resolution of ~200 km. The C_{20} coefficients are replaced
210 with values estimated from Satellite Laser Ranging (Cheng et al. 2013). The GRACE data are
211 naturally in the CM frame, thus degree 1 coefficients are not necessary. The monthly fields are filtered
212 with a statistically optimal Wiener filter following Klees et al. (2008). The optimal filter incorporates the
213 full variance-covariance information of the monthly solutions, and is designed to filter less aggressively
214 in regions where signal is stronger (e.g., over regions with a significant hydrological cycle). Using 113
215 monthly solutions that span a time period of ~10.5 years (2004.02 to 2014.06), a mass trend is
216 estimated that accounts for bias, as well as annual and semi-annual variations. The estimated mass
217 trend incorporates the full error covariance matrices of the monthly solutions, and the covariance of
218 the resulting trend coefficients is represented by the full covariance matrix of the linear parameter
219 estimate. The trend uncertainties thus differ from the more zonal patterns estimated by filters that do
220 not include information about the signal and noise covariances. The trend coefficients and the
221 associated covariance terms are represented on a uniform $2^\circ \times 2^\circ$ grid over North America.



222

223 **Figure 2.** GRACE mass trend, uncorrected for hydrology changes (a), hydrology changes derived from model
 224 PCR-GLOBWB (b), GRACE mass trend corrected for hydrology (c). (d)-(f): The 2σ uncertainty associated with
 225 each of (a)-(c). Time span of all trends is 2004.02 – 2014.06. Units are mm/yr of change to equivalent water layer
 226 thickness (EWLT).

227 As with the vertical motion data, the gravity signal over North America is not everywhere dominated by
 228 the long-term GIA response that we wish to model. Some regions are significantly influenced by non-
 229 GIA effects, such as present-day mass loss of glaciers and ice caps, and thus these locations are not
 230 included in the input dataset. The final dataset for the study area consists of 468 gravity change
 231 estimates on a grid over North America, and a full covariance matrix, C_d . The derived gravity trend
 232 and its associated uncertainty are shown in Figures 2a,d. The total gravity trend is corrected for
 233 changes in terrestrial water storage as described in the next section (3.3).

234

235

236 *3.3 Hydrology Correction for GRACE and GPS Data*

237 We estimate terrestrial water storage changes using the global hydrology model PCRaster Global
238 Water Balance (PCR-GLOBWB) (Wada et al. 2010, 2014). In the supplementary material, it is shown
239 that the results are not strongly sensitive to the choice of hydrology model, but that including a
240 correction for hydrology can yield a slightly improved result overall. In PCR-GLOBWB, the total water
241 storage is the sum of five contributions: groundwater, snow, soil moisture, surface water, and other
242 processes (canopy interception, irrigation, and river channels). The model predictions are expressed
243 as averaged monthly model solutions on a $0.5^\circ \times 0.5^\circ$ global grid. Using the same methodology and
244 time span as was used with the GRACE data, we fit a trend to the monthly model predictions. In order
245 to use the same spatial resolution as the GRACE data, the gridded trend prediction is transformed into
246 a spherical harmonic expansion truncated at degree and order 96. A fan filter with maximum degree
247 96 is then applied to the trend coefficients in order to suppress ringing artefacts (e.g., Siemes et al.
248 2013, Zhang et al. 2009). The degree 1 component was not removed from the hydrological signal at
249 this stage, but we find that its inclusion has only a small effect on the computed trend (in fact, inclusion
250 of the hydrological correction itself has only a small impact on the results). Finally, the trend
251 coefficients are transformed back to the spatial domain onto a $2^\circ \times 2^\circ$ grid and removed from the total
252 GRACE-derived gravity trend (Figure 2c). The uncertainties in the hydrological signal are estimated as
253 the uncertainty in the linear parameter estimate (Figure 2e); the GRACE uncertainties and the
254 hydrological uncertainties are added in quadrature for the total corrected signal (Figure 2f).

255

256 The vertical motion data are also corrected for the effect of hydrological loading. The vertical elastic
257 response associated with the hydrology trend derived from PCR-GLOBWB is estimated using elastic
258 Earth parameters averaged from the Preliminary Reference Earth Model (Dziewonksi and Anderson
259 1981), and then removed from the total measured vertical land motion rates prior to the inversion.
260 Within the study area, the model-predicted effect of hydrological loading on vertical velocities is
261 generally small (on average, less than 0.2 mm/yr), and is less than the 1σ uncertainty of the
262 observations at all sites.

263

264

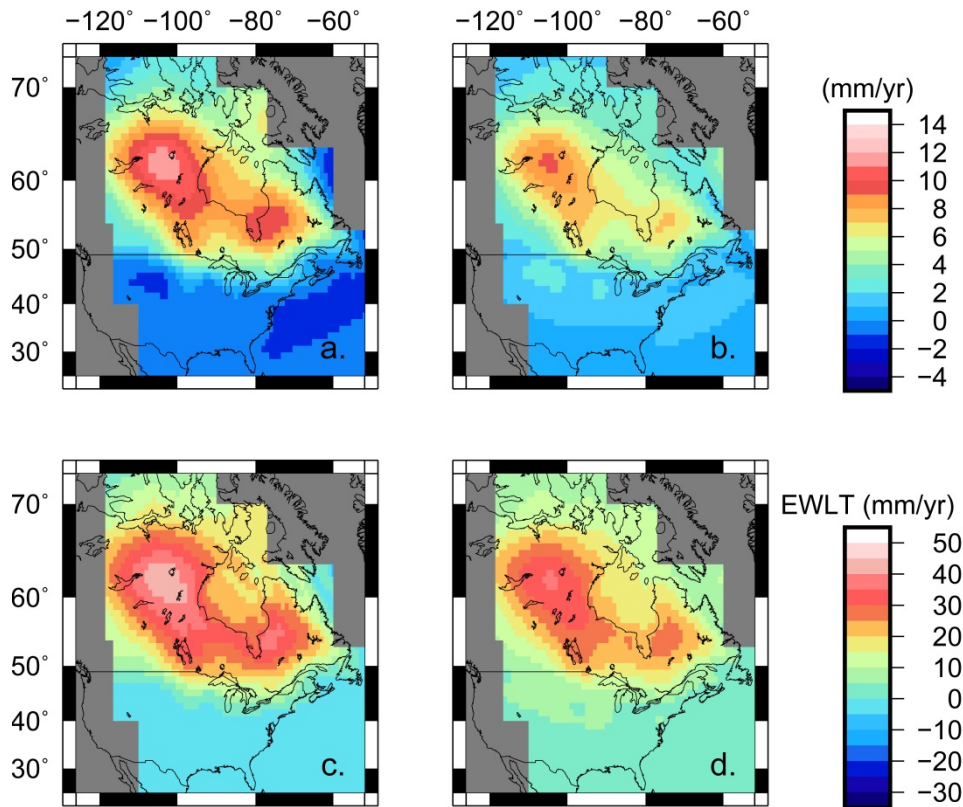
265

266 3.4 Prior GIA Models

267 The *a priori* GIA model (m in equation 2.4) consists of predictions of present-day vertical land motion
268 change and gravity change, and is the average of multiple predictions from a suite of GIA models that
269 varies the Earth's rheological structure and ice sheet history. In total, 150 model combinations are
270 used (75 Earth models and 2 ice sheet histories). The Earth models have a 90 km thick elastic
271 lithosphere and an underlying 2-layer compressible viscoelastic mantle. The upper mantle (>660 km
272 depth) viscosity values span 1×10^{19} Pa s – 5×10^{21} Pa s, and the lower mantle viscosity values span
273 1×10^{20} Pa s – 1×10^{23} Pa s. Consistent with the data constraints, the deformations of the GIA
274 models are calculated in the CM frame.

275

276 The first ice sheet scenario uses the SJD15 and Laur16 models for the Innuitian and Laurentide ice
277 sheets, respectively (Simon et al. 2015, 2016). These models feature a North American ice sheet
278 complex that has been modified relative to the global ICE-5G model (Peltier 2004) to provide an
279 improved fit of GIA model predictions to relative sea-level measurements and GPS-measured vertical
280 land motions in north-central Canada; ICE-5G is used for the rest of the globe in this version of the ice
281 sheet history. The second ice sheet scenario uses the global ICE-5G reconstruction of Peltier (2004).
282 The North American ice thickness history of ICE-5G differs significantly from Laur16 and SJD15. In
283 particular, relative to ICE-5G, Laur16's ice cover is thinner west of Hudson Bay by ~30%, and thicker
284 east of Hudson Bay by ~20-25%. These modifications to the North American component of ICE-5G
285 address ICE-5G's significant overprediction and underprediction of vertical land motion and gravity
286 rates west and east of Hudson Bay, respectively (e.g., Lambert et al. 2006, Argus and Peltier 2010,
287 Peltier et al. 2015).



288

289 **Figure 3.** *A priori* GIA model inputs. Shown are vertical land motion rates averaged over a suite of 150 models (a)
 290 and the associated standard deviation of the vertical motion rates for the entire suite of models (b). (c) and (d) are
 291 the same as (a) and (b), except rates of gravity change are shown and are expressed as change to equivalent
 292 water layer thickness (EWLT). Uncertainty rates are not VCE-scaled.

293

294 The input model covariance matrix for the suite of GIA models is constructed using equation 2.2 with
 295 small perturbations added to the main diagonal to yield a full-ranked matrix. Figure 3 shows the
 296 averaged *a priori* model predictions $\langle m_i \rangle$ and their associated uncertainty (i.e., $\sqrt{C_m^{ii}}$). The averaged
 297 predicted rates display characteristics of both ICE-5G and Laur16. For instance, west and east of
 298 Hudson Bay, two distinct domes of vertical uplift are clearly visible (Laur16), with the largest averaged
 299 rates present to the west of Hudson Bay (ICE-5G) (Figure 3). The regions with the largest ice load also
 300 have the largest uncertainty rates, consistent with the results of Wu et al. (2010b). The use of 75 Earth
 301 models and 2 different ice load histories allows the final model prediction to vary both as a function of
 302 mantle viscosity and ice thickness. The sensitivity of the final model prediction to variations in the a

303 *priori* information is examined in the supplementary material (Section S2). Characteristic behaviours
304 and misfits of selected models from the *a priori* set are also examined in the supplement (Section S4).

305

306 **4. Results**

307 The data are inverted simultaneously with the *a priori* information for the present-day GIA response
308 (equation 2.8). We discuss the prediction of both model deformation types (vertical land motion and
309 gravity change) for three scenarios: i) inversion of only the vertical motion data (termed 'Data-driven 1',
310 or D1), ii) inversion of only the gravity change data (D2), and iii) inversion of both datasets (D3).

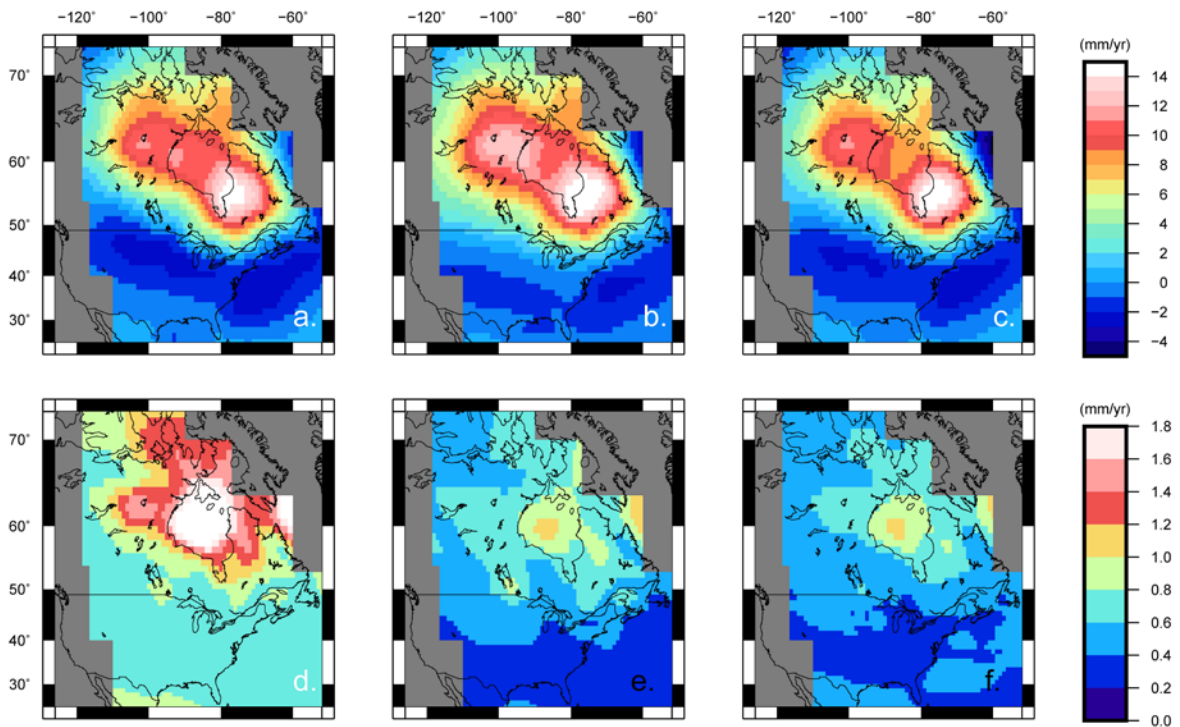
311

312 *4.1 Prediction of Vertical Land Motion*

313 When only the vertical land motion rates are incorporated into the prior model in the D1 scenario, the
314 peak predicted vertical uplift rate is 14.7 mm/yr to the southeast of Hudson Bay (Figure 4a). West of
315 Hudson Bay, a smaller secondary dome of uplift is predicted, with a maximum uplift rate of 11.3
316 mm/yr. The estimated 2σ rates of the predictions range from ~0.6-0.8 mm/yr in more southern portions
317 of the study area to 0.8-2 mm/yr in more northern regions of the study area (Figure 4d). The largest
318 predicted 2σ rates (~2 mm/yr) occur within Hudson Bay, where there is no constraint on the rate of
319 vertical land motion.

320

321 In the D2 scenario, the peak uplift rates southeast and west of Hudson Bay are 15.5 mm/yr, and 12.5
322 mm/yr, respectively (Figure 4b). Relative to D1, the D2 prediction lacks the northwest-southeast
323 trending band of subsidence that is observed to the south of the margin of the former ice sheet (Figure
324 4b). For D2, the 2σ uncertainties range from 0.2 mm/yr to 1.2 mm/yr (Figure 4e). When both datasets
325 are incorporated in the D3 scenario, the peak uplift rates east and west of Hudson Bay are 15.0 mm/yr
326 and 11.2 mm/yr, respectively (Figure 4c). The predicted vertical motion uncertainty rates for D3 are
327 between 0.2-0.6 mm/yr in the southern part of the study area, 0.6-1.2 mm/yr in the northern part of the
328 study area, and peak in Hudson Bay (Figure 4f).



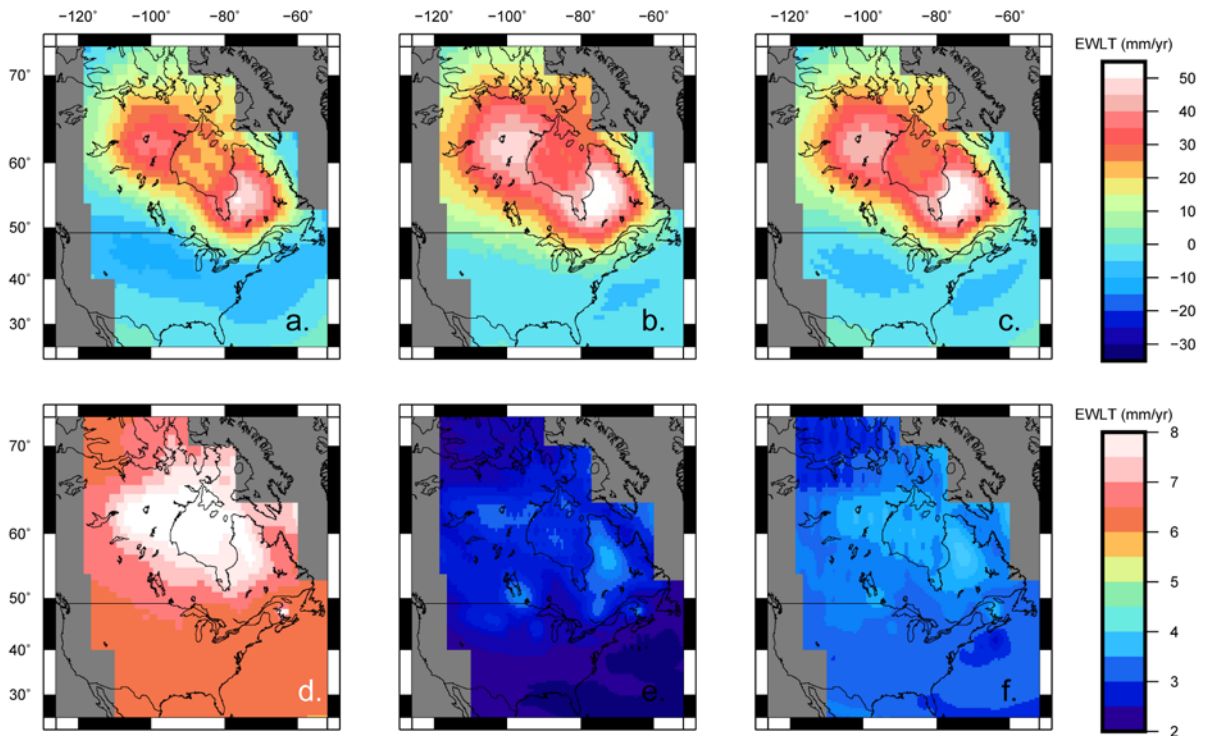
329

330 **Figure 4.** Results for prediction of rates of GIA-induced vertical land motion for three scenarios: a) inversion of the
 331 vertical motion data only (D1), b) inversion of the gravity change data only (D2), and c) inversion of both the
 332 vertical motion and gravity data (D3). (d-f) the predicted posterior 2σ uncertainties for each of (a-c). Uncertainties
 333 are not VCE scaled.

334

335 4.2 Prediction of Gravity Change

336 In general, the model predictions of gravity change are analogous to those for vertical motion (Figure
 337 5). All scenarios predict two distinct positive anomalies in the gravity trend west and east of Hudson
 338 Bay, and in all cases, the peak rate of the eastern anomaly is predicted to be somewhat larger than
 339 the western anomaly. The respective eastern and western peak rates for the combined D3 solution are
 340 58 mm/yr and 44 mm/yr of change in equivalent water layer thickness (EWLT). For D3, the 2σ
 341 uncertainties range from approximately 3-4 mm/yr EWLT. The most prominent difference between the
 342 three model predictions is in the region around the southern boundary of the former Laurentide Ice
 343 Sheet. Here, as with the vertical motion predictions, inclusion of the GPS data in models D1 and D3
 344 decreases the model-predicted gravity signal, and in the case of D1, a band of negative values is
 345 predicted that is not present in the GRACE data.



346

347 **Figure 5.** Results for prediction of rates of GIA-induced gravity change for three scenarios: a) inversion of the
 348 vertical motion data only (D1), b) inversion of the gravity change data only (D2), and c) inversion of both the
 349 vertical motion and gravity data (D3). (d-f) the predicted posterior 2σ uncertainties for each of (a-c). Shown in
 350 mm/yr of change to equivalent water layer thickness. Uncertainties are not VCE scaled.

351

352 4.3 Results of Variance Component Estimation

353 Variance component estimation is used to weight the datasets relative to each other and relative to the
 354 prior model information. Following Koch and Kusche (2002), the VCE iteratively determines the
 355 weighting factors applied to each input dataset (equation 2.8). The converged variance weighting
 356 factors and their ratios are given in Table 1. The ratios describe how each input is weighted relative to
 357 the other(s) in the model solution. In D3, the prior information is down-weighted by a factor of 1.1, and
 358 the gravity data are weighted relative to the vertical motion data by a factor of -0.13 . The gravity data
 359 are therefore weighted less heavily than the vertical motion data in the D3 solution; however, adding
 360 the gravity data significantly improves the fit of the model-predicted gravity rates to the observations
 361 (next section). In general, the predicted solutions with and without VCE are quite similar, although

362 down-weighting of the prior information moderately increases the predicted model uncertainties (see
363 supplementary material).

364

365 **5. Discussion**

366 *5.1 Evaluation of the Predicted Models: RMS Misfits and Post-fit Residuals*

367 Table 2 summarizes the RMS misfits between the datasets and the different model scenarios. When
368 only the vertical data are incorporated in the D1 scenario, the RMS misfit is 1.06 mm/yr. Not
369 surprisingly, the poorest fit to the vertical motion observations occurs when only the gravity data are
370 incorporated in the model (RMS = 2.06 mm/yr). The poorer fit of D2 to the vertical motion data is in
371 part due to the lack of prediction of the northwest-southeast trending band of subsidence observed
372 near the southern margin of the former ice sheet, which is likely due to the GIA effect of a collapsing
373 peripheral bulge in the mantle. There is no analogous signal present in the gravity data, which may
374 indicate either a limit to the spatial resolution of the GRACE data, or that the effect of mass
375 redistribution on gravity is small relative to its direct effect on vertical motion. D3's RMS value for
376 prediction of vertical motion is 1.25 mm/yr.

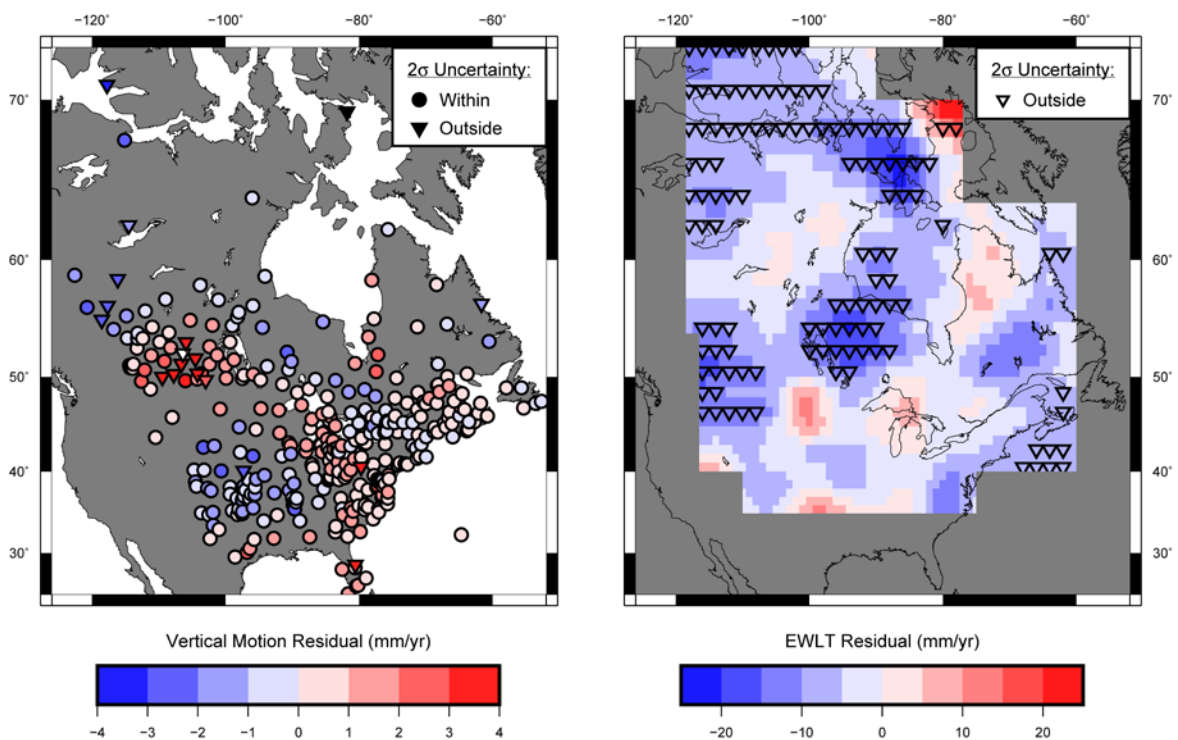
377

378 Incorporation of only the vertical motion data provides the poorest fit to the gravity data (RMS = 1.24
379 mm/yr). As expected, the best-fit to the gravity change observations occurs when the gravity data are
380 incorporated in the prior model (RMS = 0.57, and 0.76 mm/yr for D2 and D3, respectively). The
381 combined D3 model provides the best overall prediction of both vertical motion and gravity change
382 (Table 2).

383

384 We also calculate the corresponding RMS and χ^2 misfit values of the recent ICE-6G forward model
385 (Peltier et al. 2015). Relative to the vertical land motion and gravity change data, ICE-6G's respective
386 RMS values are 1.52 and 0.96 mm/yr (Table 2). The equivalent χ^2 values are 1.3 and 15.7, which
387 yields a total χ^2 value of 9.6 and indicates that the misfits to the GRACE data are largest in regions

388 where the error is lowest (i.e., the north). The reasonable fit (both RMS and χ^2) of ICE-6G to the
389 vertical motion data is expected since these data were used to constrain the model. GRACE gravity
390 data, in contrast, were not used to constrain ICE-6G, but rather used as an independent dataset for
391 comparison with model predictions. In their own comparison, Peltier et al. (2015) noted further
392 improvement of ICE-6G's fit to the gravity data may require modifications to the hydrology correction
393 or to ICE-6G itself. The RMS misfit values for ICE-6G are most similar to those of the D1 model, which
394 uses only the vertical land motion data as observational constraint.



395
396 **Figure 6.** Residuals of model predictions relative to observations for vertical motion (left) and gravity change
397 (right). Residuals that are within the 2σ uncertainty of the observations are plotted with a circle (left only);
398 residuals that exceed the 2σ uncertainty of the observations are plotted with an inverted triangle (left and right).
399 Model prediction is for the D3 combined data scenario (both vertical motion and gravity change data are
400 incorporated).

401

402 The individual post-fit residuals provide an additional evaluation of model fit (Figure 6). For most sites
403 within the study region, the D3 model predicts rates of vertical motion within the uncertainty of the
404 observations. However, there is a notable region of misfit centred on approximately 110° W (west of
405 Lake Winnipeg, Figure 1) where the model overpredicts the rate of vertical motion relative to the data

406 (Figure 6a). Comparison of ICE-6G predictions to vertical motion observations shows a similar misfit in
407 this region (e.g., Snay et al. 2016). The rates of gravity change predicted by D3 are also within the
408 data uncertainties for large parts of the study area, with coherent regions of misfit remaining west of
409 Lake Winnipeg, southwest of Hudson Bay, and in the northern part of the model domain (Figure 6b).

410

411 These regions of misfit may be partly explained by missing or inaccurate representation of non-GIA
412 processes within the model inputs. The separation approaches of both Lambert et al. (2013) and
413 Wang et al. (2013, 2015) indicate a large increase in water storage after 2005 in the upper Assiniboine
414 River watershed of the Canadian prairies west of Lake Winnipeg, an inference further supported by in-
415 situ hydrological measurements. These studies estimate that water accumulation in this region yields a
416 positive mass anomaly of between 20-34 mm/yr EWLT. This anomaly is not present in PCR-GLOBWB
417 (Figure 2b), and its location approximately corresponds with a region of significant misfit in our gravity
418 predictions (Figure 6b). Because the model underpredicts the gravity signal around Lake Winnipeg,
419 removal of a positive hydrological anomaly in this region may improve the fit of the model to the data.
420 Underprediction of the gravity signal by the model in the northern part of the study area may similarly
421 indicate modification to the hydrology model is needed in this region, as well as unmodelled influences
422 from mass loss of nearby glaciers (Figure 2b). The prediction of vertical motion will likewise be
423 affected by any unmodelled hydrology signals, although the correction for hydrological changes is
424 smaller for vertical motion data than it is for gravity change measurements (Section 3.3). West of Lake
425 Winnipeg, vertical land motion data may also be influenced by pronounced localized subsidence
426 caused by extensive potash mining in southern Saskatchewan (Samsonov et al. 2014), an effect not
427 considered prior to inversion of the data. Further effort to identify and remove any remaining non-GIA
428 signals from the vertical motion data may be needed to reconcile the GIA model with the data in this
429 region (Figure 6a). Use of PCR-GLOBWB, however, improves the fit of the D3 model prediction to
430 vertical motion observations west and east of Hudson Bay.

431

432

433

434 5.2 Predicted GIA Uncertainty

435 Because many studies use forward GIA models for a variety of applications, it is useful to have some
436 measure of the uncertainty associated with GIA predictions. GIA uncertainty estimates generated by
437 variation of forward model parameters tend to provide only broad constraint, as the variation in rates is
438 often of the same order of magnitude as the predicted signal (Figures 3b,d). The method presented
439 here provides an alternative means of estimating the uncertainty associated with the GIA process. An
440 important assumption implicit in this discussion of the posterior uncertainty is that the input data are
441 attributable to the GIA process only. As discussed in Section 5.1, analysis of the post-fit residuals has
442 identified specific regions where this assumption may not hold. However, in general, the posterior
443 uncertainty prediction quantifies how well we can expect to resolve the GIA signal, given the input data
444 constraints and prior information. For example, when only vertical motion data are inverted, we can
445 expect to predict GIA-induced vertical motion to within 0.6-0.8 mm/yr in the southern part of the study
446 area where GPS sites are spatially dense and are often characterized by relatively small uncertainties
447 (Figure 4d). The same example indicates an uncertainty of ~2 mm/yr associated with GIA vertical
448 motion in and around Hudson Bay (Figure 4d); this is a logical result, since here the GPS sites are
449 sparsely distributed and characterized by relatively large uncertainty, and the prior model information
450 also has large uncertainty (Figure 3b). Incorporation of the gravity data lowers the maximum posterior
451 uncertainty of the predicted vertical motion in D3 (Figure 4f). However, the spatial pattern of the
452 uncertainty is largely unchanged for each of D1-D3 (Figures 4d-f), indicating that the prior information
453 also contributes to the spatial characteristics of the posterior uncertainty. The results are similar for the
454 prediction of gravity change (Figures 5d-f) with higher posterior uncertainty predicted in regions with
455 higher uncertainty in the data and/or the prior information. Incorporation of the regularly gridded gravity
456 data reduces the predicted model uncertainty (Figures 5e,f) relative to the scenario in which only
457 vertical motion data are inverted (Figure 5d). Throughout the study area, the uncertainty associated
458 with GIA-induced gravity change is ~3-4 mm/yr EWLT for the D3 model (Figure 5f).

459

460 5.3 A priori GIA information

461 The incorporation of prior GIA model information into the model solution introduces dependence of the
462 final prediction on uncertain forward model information. The use, however, of *a priori* information

463 permits prediction of the GIA response everywhere including in regions of poor data coverage, and
464 provides a numerical way to relate vertical crustal motion to gravity change (i.e., vertical motion data
465 can be used to predict gravity change, and vice versa, through their relationship in the model
466 covariance matrix, C_m). The use of prior GIA model information may also help to identify any
467 significant and spatially coherent non-GIA features that remain in the datasets. For example, the
468 vertical motion data show a region of subsidence west of Lake Winnipeg, centred on approximately
469 110° W longitude and surrounded by larger uplift rates to the east and northwest (Figure 1). This
470 feature is likely not GIA-induced (at least, it is unlikely that this pattern of deformation would arise from
471 any of the forward model combinations assumed here), and it subsequently appears as a high misfit
472 region in the final model prediction (Figure 6a), suggesting the need to account for an additional
473 process here (Section 5.1). At minimum, this result is informative in the sense that we actually do not
474 wish our final GIA model to succeed in the prediction of non-GIA features.

475

476 The supplementary material examines the influence of the prior information on the predicted solution.
477 We consider three representations for the prior information in which the set of Earth models remains
478 the same but is coupled to each of ICE-5G only, Laur16/SJD15 only, and both ice sheet histories
479 together. Results show that the final prediction can indeed be affected by the choice of prior ice sheet
480 history, particularly when ICE-5G is the only ice sheet history used (Figures S6-S11). This issue,
481 however, can be largely mitigated by either the use of more data or the use of other ice histories in the
482 prior information. We find that similar model predictions are achieved as long as either: i) sufficiently
483 large datasets are used, or ii) sufficient variation is included in the prior Earth model and ice sheet
484 combinations. The second point may be particularly important for study areas where constraint on the
485 size and geometry of the past ice sheet is limited.

486

487 Although the model solution is only weakly sensitive to variations in the prior input, it is still possible
488 that simplifications in the suite of Earth models may influence the results. Within the Earth model set, a
489 two-layer mantle viscosity profile is varied, although mantle viscosity profiles employed in forward GIA
490 studies often have three or more layers (e.g., Peltier et al. 2015). A narrower viscosity range than that
491 employed would be more consistent with the VM2/VM5a viscosity profiles to which the ice sheet

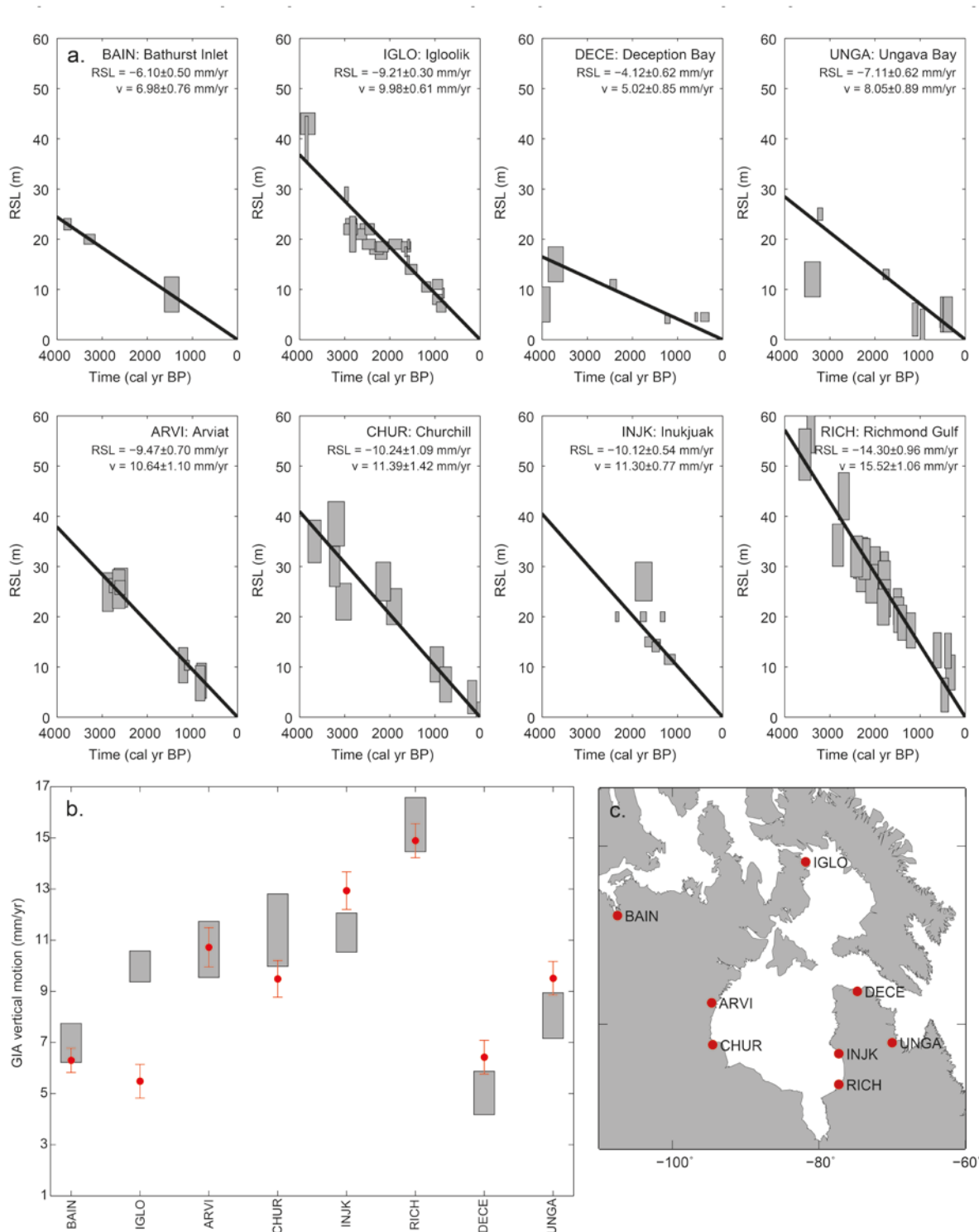
492 histories were tuned and may be more realistic, but would not capture the wider range of possible GIA
493 deformation, including variations due to potentially heterogeneous Earth structure. As is, the
494 constructed covariance matrix may contain a few very end-member forward models; however, the
495 overall behaviour of the model set as a whole provides misfits consistently centred around zero (see
496 also Section S4 in the supplementary material). Tests performed with a narrower range of upper and
497 lower mantle viscosities predict modestly lower posterior uncertainties and create little change to the
498 predicted signal. Studies have also suggested that the effective elastic thickness of the lithosphere in
499 cratonic areas of North America may be ≥ 140 km (Zhao 2013, Tesauro et al. 2015). This value is
500 significantly larger than the effective elastic thickness of 90 km that is used in the suite of *a priori* Earth
501 models. Exploring the effect of variations in elastic lithospheric thickness and the use of more complex
502 mantle viscosity profiles in the construction of the model covariance matrix may be a worthwhile future
503 contribution to this work.

504

505 *5.4 Comparison with rates from late Holocene RSL data*

506 Finally, we evaluate the ability of the D3 GIA model to reproduce independent estimates of vertical
507 land motion. Using relative sea-level (RSL) data from Simon et al. (2016), late Holocene rates of
508 vertical motion are estimated at 8 locations in the northern part of the study area (Figure 7). The RSL
509 data consist of ages and elevations of radiocarbon-dated material, including shells, bones and plant
510 material. First, RSL change rates are calculated for each site by fitting a linear trend to RSL indicators
511 from the last 4 kyr. Vertical motion rates are subsequently determined by adding an estimate of the
512 contemporary geoid change rate. Over a time period of 4 kyr, relative sea level change can be
513 considered to be largely insensitive to shorter term forcing; the rates thus provide reasonable site-
514 specific proxies for contemporary vertical land motion associated with GIA. With the exception of
515 Igloolik, which is near the northern boundary of our study area, the fit is satisfactory between vertical
516 land motion rates predicted by D3 and the independent vertical rates estimated from the RSL data
517 (Figure 7). The D3 model strongly underpredicts vertical land motion at Igloolik, likely due to both
518 inadequate constraint in the input data and the site's relative proximity to large glaciers on Baffin
519 Island. A linear fit may somewhat overestimate the rate of vertical motion due to the non-linear nature
520 of GIA. This will be particularly true for time spans beyond 4 kyr and in load-central regions. However,

521 the selected data are not strongly non-linear from 4 kyr onwards, and tests with 1.5 and 2 kyr time
 522 windows yield a similarly good fit, although the geologically-derived rates have higher uncertainties
 523 due to the shorter time period. Therefore, with the exception of Igloodik, the results indicate that in
 524 general the D3 model provides a reliable prediction of the GIA signal, at least in the formerly glaciated
 525 part of the study area.



527 **Figure 7.** a) Late Holocene RSL data for 8 sites. 'RSL' gives the rate of change of relative sea level derived from
528 a linear fit (black line), 'v' gives the associated rate of vertical motion after addition of a term for the geoid rate of
529 change (see text). b) Comparison of model-predicted GIA vertical motion for model D3 (red circles) with the
530 independent estimates derived from the sea level data (grey boxes). c) Map of site locations.

531

532 **6. Conclusions**

533 We generate a data-driven prediction of the present-day GIA response in North America through the
534 simultaneous inversion of GPS-measured vertical land motion rates, GRACE-measured gravity
535 change rates, and *a priori* GIA model information. Our methodology expands on that of Hill et al.
536 (2010), and includes statistically appropriate weighting of the inputs via variance component
537 estimation. In models D1-D3, we compare the predicted results for the inversion of the vertical motion
538 data, the gravity data, and both datasets. There are two main differences between the scenarios: i) in
539 the RMS sense, the vertical motion data alone have the poorest ability to predict gravity change, and
540 vice versa, and ii) inversion of the gravity dataset, with its regular grid and smaller uncertainties
541 relative to the vertical motion data, serves to reduce the maximum posterior uncertainty of the
542 predicted model in the northern part of the study area where GPS coverage is less dense. All
543 scenarios show the robust prediction of two centres of uplift and mass gain to the west and east of
544 Hudson Bay. Predictions of the D3 model provide the best overall fit to both datasets, with respective
545 RMS misfits to the vertical and gravity change data of 1.25 mm/yr and 0.76 mm/yr EWLT.

546

547 An important assumption implicit to the methodology is that the input data arise only from the GIA
548 process. Realization of this assumption is encouraged by correction for hydrological changes in both
549 datasets, as well as by the exclusion of particular data points where non-GIA contributions are
550 expected to be large. However, remaining vertical motion and gravity change misfits in D3 may be
551 partially explained by significant non-GIA signals still present in the data, including unmodelled
552 hydrological changes in the central Prairies west of Lake Winnipeg (Lambert et al. 2013, Wang et al.
553 2013, 2015), and possibly unmodelled regional vertical motion changes due to active mining
554 (Samsonov et al. 2014). Further effort to clearly identify such features and determine their appropriate
555 treatment (i.e., their incorporation into an *a priori* correction, or simply their removal from the input

556 dataset) is needed. Aside from these regions of misfit, the overall agreement between model
557 predictions and observations is very good (Figure 6); the D3 model predicts rates of vertical motion
558 and gravity change that are within the 2σ uncertainty of the observations throughout much of the study
559 area. When compared to the predictions of the independent ICE-6G forward model, D3 performs quite
560 well; D3's prediction of vertical motion is slightly better (RMS = 1.25 mm/yr) but still comparable to that
561 of ICE-6G (RMS = 1.52 mm/yr), and the prediction of gravity change is improved (RMS = 0.76 versus
562 RMS = 0.96 mm/yr EWLTT) (Table 2).

563

564 At present, the inability to quantify the long-term GIA signal accurately is a significant obstacle to
565 achieving a comprehensive understanding of how shorter time-scale processes (e.g., ice mass loss,
566 continental hydrology variations, etc.) contribute to present-day rates of crustal deformation, gravity
567 change, and sea-level variation, and how these contributions may change over time. GIA uncertainty
568 estimates generated through the variation of ice sheet and Earth rheology parameters offer little
569 additional constraint, as these estimates are often of the same order of magnitude as the predicted
570 signal. While a data-driven model of the kind presented here has limited ability to constrain ice/Earth
571 model combinations, an advantage to the approach in this study is the ability to estimate the formal
572 uncertainty of the present-day GIA response at magnitudes that are significantly smaller than the a
573 *priori* uncertainty estimates yielded by forward modelling. The results, therefore, provide future work
574 with a useful basis from which to constrain better non-GIA contributions to present-day rates of
575 change.

576

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787 **Tables**

Data Incorporated	σ^2 Squared Value			Ratios	
	σ_1^2 (Vertical)	σ_2^2 (Gravity)	σ_μ^2 (Prior)	σ_1^2/σ_2^2	$\sigma_1^2/\sigma_\mu^2, \sigma_2^2/\sigma_\mu^2$
D1: Vertical only	1.1	-	7.1	-	0.15, -
D2: Gravity only	-	9.3	0.5	-	-, 18.6
D3: Vertical+Gravity	1.4	10.9	1.1	0.13	1.3, 9.9

788

789 **Table 1.** Results of the variance component analysis. σ_1^2 and σ_2^2 are the variance factors applied to the
 790 vertical motion data (dataset 1) and gravity change data (dataset 2), respectively, and σ_μ^2 is the
 791 variance factor applied to the prior information. The ratios describe how each input is weighted relative
 792 to the other(s).

793

Model Type	RMS Values	
	Vertical Motion (mm/yr)	Gravity Change (mm/yr EWLT)
Evaluation of Inputs		
Null	3.36	2.65
Prior	2.64	1.02
Data-Driven Inversion		
D1: Vertical only	1.06	1.24
D2: Gravity only	2.06	0.57
D3: Vertical+Gravity	1.25	0.76
Forward GIA Model Comparison		
ICE-6G	1.52	0.82

794

795 **Table 2.** RMS values of model predictions relative to observations for vertical crustal motion and
796 gravity change. Null model gives the RMS value of the data set relative to no model prediction, and
797 prior model gives the RMS of the *a priori* average model relative to the observations. Data-driven
798 model results shown for three input data scenarios: vertical land motion data ('Vertical only', D1),
799 gravity change observations ('Gravity only', D2) and both datasets ('Vertical+Gravity', D3). Vertical
800 motion RMS values calculated for $n_{gps} = 340$ sites; gravity change RMS values calculated for $n_{grace} =$
801 468 sites. Units for vertical and gravity change are mm/yr and mm/yr of EWLT change, respectively.