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# Assessing Global Present-day Surface Mass Transport and Glacial Isostatic Adjustment from Inversion of Geodetic Observations

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Key points: Separation of present-day mass transport and glacier isostatic adjustment signatures
using multiple observation types.

26 Estimation of geocenter motion due to past and present-day surface mass changes.

27 Enhanced GIA signals detected in areas of low upper mantle viscosity and recent ice loss.28

29 Abstract: Long-term monitoring of global mass transport within the Earth system 30 improves our ability to mitigate natural hazards and better understand their relations to 31 climate change. Satellite gravity is widely used to monitor surface mass variations for its 32 unprecedented spatial and temporal coverage. However, the gravity data contain signals 33 from visco-elastic deformation in response to past ice sheet melting, preventing us from 34 extracting signals of present-day surface mass trend (PDMT) directly. Here we present a 35 global inversion scheme that separates PDMT and visco-elastic glacial isostatic adjustment (GIA) signatures by combining satellite gravimetry with satellite altimetry and ground 36 37 observations. Our inversion provides global dual data coverage that enables a robust 38 separation of PDMT and GIA spherical harmonic coefficients. It has the advantage to 39 provide estimates of the Earth's long wavelength deformation signatures and their 40 uncertainties. Our GIA result, along with its uncertainty estimates, can be used in future 41 GRACE processing to better assess the impact of GIA to surface mass change. Our GIA 42 estimates includes rapid GIA uplift in the Southeast Alaska and the Amundsen Sea 43 Embayment, due to the visco-elastic response to recent glacial unloading. We estimate the average surface mass change rate from 2002-2010 to be -203±3 GT·a<sup>-1</sup> in Greenland, -44 126±18 GT·a<sup>-1</sup> in Antarctica and -62±5 GT·a<sup>-1</sup> in Alaska. The GIA low degree spherical 45 46 harmonic coefficients are sensitive to rheological properties in Earth's deep interior. Our

47 low-degree GIA estimates include geocenter motion and  $J_2$  which provide unique

48 constraints to understand Earth's lower mantle and ice history.

49

50 Plain Language Summary:

51 Surface mass exchange between the Earth's 'spheres' – atmosphere, hydrosphere, cryosphere, 52 biosphere and pedosphere are enormous. Monitoring surface mass change helps to understand 53 climate change and mitigate hazardous effects such as extreme drought or flooding. 54 Measurements of surface mass change are perturbed by subsurface processes, such as mantle 55 flow underneath the Earth's crust. Often, a model is used to correct the subsurface signals from 56 observations, and likely to introduce un-modeled process errors into the surface mass estimates. 57 We use a mostly data-driven method to extract present-day surface mass change trends along 58 with their error estimates. The results give an enhanced view of the surface mass processes and 59 can help improving the accuracy of future surface mass estimations. Moreover, we get a more 60 accurate picture of the subsurface processes, that are mainly caused by the Earth's viscous 61 response to past ice sheet melting. We find that the Earth's crust bounces back more rapidly after 62 glacier melting events than we assumed before, especially in places where the mantle viscosity is 63 low.

64

65 1. Introduction

Global climate change modifies the mass redistribution process within the Earth system. Presentday surface mass transport, in turn, changes the geoid and global sea level. Surface mass
transport, mainly from fluid water moving between Earth's "spheres" impacts societies and its
relation to climate change needs to be better understood. Global monitoring of PDMT thus is
vital in itself and to understand climate change. Before the Gravity Recovery and Climate

71 Experiment (GRACE) satellites were launched, large-scale components of mass movement 72 within the Earth system were inferred from satellite laser ranging (SLR) observations [Cox and Chao, 2002; Dickey et al., 2002; Cheng et al., 2013]. The time-varying gravity field has been 73 74 monitored continuously with unprecedented accuracy since the GRACE era [Tapley et al., 2004; 75 Wouters et al., 2014]. However, GRACE is not only sensitive to PDMT, but also to subsurface 76 mass changes mainly caused by the Earth's viscous response to the removal of ice sheets, known 77 as glacial isostatic adjustment (GIA) [Peltier, 2004; Wouters et al., 2014]. Although efforts have 78 been made recently to improve our capability to monitor surface mass transport, the current 79 method is still subject to GIA model errors. The presence of the residual GIA signals, 80 specifically GIA signals due to the recent past ice loss, in most geodetic data and unquantified 81 GIA model errors make the direct observation of surface mass transport trend a difficult 82 endeavor.

83

84 Present-day surface mass redistribution can be estimated by measuring the instantaneous elastic 85 response of the Earth's crust to surface loading changes [Farrell, 1972; Wahr et al., 1998]. In 86 practice, the GIA process complicates interpretation of the linear trend observations. Physical 87 models have been developed to reconstruct present-day GIA signals [Peltier et al., 2015; Peltier 88 et al., 2018; Ivins et al., 2013], but such models are largely based on forward-fitting approaches, 89 partial-domain viscosity inversions, piecewise subsequent iterative modifications with new data, 90 and are thus subject to ambiguities and uncertainties in the past ice history, Earth rheology and 91 flow constitutive laws. Consequently, such development processes make quantification of model 92 uncertainties particularly hard [van der Wal et al., 2015; Caron et al., 2018], although for some 93 cases such as in North America, uncertainty estimates due to lateral viscosity have been

94	presented [Li et al., 2020]. Furthermore, ice loading histories in global models generally do not
95	include ice changes in the last centuries in regions of low viscosity such as Antarctica, Alaska,
96	Patagonia, Iceland [Jacob et al., 2012]. Thus, mass change estimates in those regions will be
97	biased when global models are used to correct for the GIA signal. On the other hand, GIA
98	models with parameters that are fit to regional GIA observations in areas of low viscosity such as
99	in West Antarctica [e.g. Nield et al. 2014] do not capture the low degree signal, while the far-
100	field GIA signal can contribute 40% of the total Antarctic signal [Caron and Ivins 2020].
101	
102	Earlier studies suggest that PDMT- and GIA-induced elastic and viscoelastic signatures have
103	distinct spatial-temporal patterns in different observation types [Wahr et al., 1995; Wu et al.,
104	2010]. In light of this, regional inversion models that combine multi-type observations have been
105	developed to separate PDMT and GIA [Martín-Español et al., 2016; Sasgen et al., 2017; Gunter
106	et al., 2014; Simon et al., 2017]. These regional inversions solve for GIA and PMDT
107	simultaneously and are successful in providing mass transport information on a local scale.
108	However, they cannot assess the impact of geocenter motion nor deliver global long-wavelength
109	spherical harmonic coefficients. On the other hand, global inversion of GIA and PDMT is
110	difficult, due to the paucity of dual observation types in both the oceans and the interior of the
111	Greenland and Antarctic ice sheets.
112	
113	Here, for the first time, we develop a high-resolution method to combine data from multiple
114	observation platforms to assure global coverage of dual observation types in estimating PDMT

and GIA simultaneously. Our results provide nearly data-driven estimates of PDMT, including

116 global non-steric sea level change, and valuable spherical harmonic coefficient constraints on ice

117 history and lower mantle viscosity. Compared to global dynamic GIA models, coefficient 118 ambiguities and uncertainties are reduced by our use of precise global geodetic data, and the 119 effect of recent ice unloading is included. Our estimates also include long-wavelength signal that 120 is missing in regional inversions (empirical GIA estimates). GRACE and GNSS have been used 121 to measure precisely the load induced deformation at both local and global scales [Wahr et al., 122 1998; Blewitt et al., 2001; Wu et al., 2010]. Although both datasets have the potential to reach 123 global coverage, for our inversion needs, gaps in both spatial and temporal domains exist. A few 124 examples are: the GRACE data system is not sensitive to degree-1 geocenter motion; GNSS 125 station distribution is sparse over the ocean and near glaciers; GNSS data gaps and 126 discontinuities exist in the time series; crustal deformation signals from tectonics also complicate 127 the interpretation. For our purpose of separating global PDMT and GIA, it is vital to use 128 complementary data for global dual coverage and reconcile information from overlapping 129 observations to reduce uncertainties. We therefore present a kinematic inversion that takes 130 advantage of the multiple globally distributed data sets to separate PDMT and GIA with robust 131 uncertainty estimation. In addition to GRACE observations, we included 786 geodetic stations 132 on land to help separating GIA and PDMT (Figure 1). Ocean Bottom Pressure (OBP) results 133 from JPL's Estimating the Circulation and Climate of the Ocean (ECCO) model 134 (https://ecco.jpl.nasa.gov/products/all/) were included to help separating GIA and PDMT in the 135 ocean [Fukumori et al., 1999]. Previous work of Wu et al. [2010] suffered from a lack of dual 136 observation types in the interior of the Greenland and Antarctic ice sheets, resulting in 137 considerable GIA uncertainties in these two regions. Here we supply new ICES at elevation 138 change rates over these ice sheets to fill the data gaps (Figure 2) in *Wu et al.*, [2010]. 139

140 In this paper, we present a simultaneous PDMT and GIA estimation using the method mainly 141 developed in *Wu et al.*, [2010], but with a much improved globally distributed observation 142 network of multiple satellite and ground data, including ICES at altimetry in particular, and 143 improved resolution and accuracy from a number of observation techniques. We first describe 144 the three data types and the OBP model as well as their processing strategies in detail in section 145 2. We conduct the inversion in the spectral domain from degree 1 to 60 for the GIA signatures 146 and up to degree 180 for the PDMT. The inverse methodology is presented with detail in section 147 3. Our inverted results for both GIA and PDMT in spherical harmonic and spatial domains are 148 presented in section 4, followed by discussion and conclusions in section 5.

149

150 2. Observation Types and Data Processing Methods

151 GIA-induced deformation and gravity changes have long relaxation times, thus are reasonable to 152 be approximated by linear rates over the time scale of our study, except for a few areas of low 153 viscosity where relaxation time can be in the order of decades such as parts of the Antarctica [e.g. 154 Nield et al., 2014]. On the other hand, PDMT has considerable annual and inter-annual 155 variability, and is accelerating in the Polar Regions and certain glaciated mountain regions 156 [Scambos et al., 2004; Jiang et al., 2010; Harig and Simons, 2015; Shepherd et al., 2018; 157 Shepherd et al., 2020]. To minimize the effect of nonlinear PDMT variations in different 158 datasets, and to maximize the overlapping time from multiple observations, we choose to pre-159 process all the data to derive linear change rates from the period of 2002-2010 which is the only 160 time window when all four datasets are available. Longer data sets exist for GRACE, OBP and 161 GPS, but not for ICESat, and the PDMT rate has changed considerably since 2010 [Shepherd et 162 al., 2018].

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164 We use CSR RL06 GRACE monthly solutions with spherical harmonics degree and order 60. 165 GRACE C<sub>20</sub> coefficient time series are replaced with estimates from SLR [*Cheng et al.*, 2013]. 166 The GRACE data system is insensitive to geocenter motion, which is driven by degree-one mass 167 changes of both PDMT and GIA. We do not correct the PDMT and GIA degree-one coefficients, 168 but estimate them in the subsequent global inversion. Calibrated full covariance matrices up to 169 degree and order 60 [John Ries, private communication] are used in deriving the trends and their 170 full covariance matrix. To reduce regional tectonic effects, we fit and remove the spherical 171 harmonic co-seismic deformation patterns of the 2004 Sumatra Earthquake and the 2007 172 Southern Sumatra Earthquake near Benkulu from GRACE coefficient rates [Han et al., 2013]. 173

174 We use monthly ECCO-JPL ocean bottom pressure (OBP) products [Fukumori et al., 1999] to derive linear trends in  $3^{\circ} \times 3^{\circ}$  grid cells. The assimilated OBP results provide oceanographically 175 176 induced surface load information, and do not account for tectonic and non-tectonic signals such 177 as earthquake and GIA induced deformation. In the inversion, the OBP trends are modeled as 178 mass changes within the water column between the sea surface and the time-variable geoid due 179 to dynamic ocean circulation. However, the OBP products are not actual measurements and 180 contain random and spatially correlated errors. Although Wu et al. [2010] included a large 181 random error component in their diagonal covariance matrix, spatial correlation was neglected. 182 Such an approach significantly simplified and underestimated the true errors in the OBP trends, 183 especially in the long-wavelength domain. To mitigate this problem, we add a spatially 184 correlated error component to the covariance matrix of the OBP trends to allow for likely long-185 wavelength errors. We add a 500-km Gaussian correlated error component in the construction of

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the covariance matrix for the OBP trends. The amplitudes of the random and correlated
components are chosen to be consistent with the grid point differences and low-degree spherical
harmonic differences of two OBP models (ECCO and the Ocean Model for Circulation and
Tides (OMCT) [*Dobslaw and Thomas*, 2007]). Such an empirical covariance matrix will serve to
strengthen the error statistics of the global inversion.

191 We choose space-geodetic stations in tectonically stable regions that have more than three years 192 of data to calculate the long-term crustal motion rate. To increase spatial resolution in areas of 193 rapid mass changes, we densify the station coverage by adding regional GPS networks in 194 Greenland, SE Alaska, and Antarctica. We only use vertical results for sites located in SE Alaska 195 to avoid possible bias in horizontal velocities from regional tectonics. We include additional 196 stations that are installed after 2010 but near GIA deformation centers (e.g. Hudson Bay, 197 Canada), and whose time series are only exhibiting linear motion. We process daily GPS data in 198 five-minute batches with JPL's GIPSY 6.4 software. We estimate wet tropospheric delay and 199 receiver clocks together with station coordinates. The Vienna Mapping Function [Boehm et al., 200 2006] is used to map zenith tropospheric delay to lower-elevation angles. We use the FES2004 201 ocean tide model [Lyard et al., 2006] to correct the surface displacements caused by ocean tides. 202 JPL's non-fiducial orbits are used for data reduction, and then we align all station coordinates to 203 the IGb08 reference frame [*Rebischung et al.*, 2012]. Finally, station velocities in the east, north, 204 and up directions are derived separately [Jiang et al., 2012], and are further projected to remove 205 the ITRF network translation information to avoid contamination from a possible origin drift of 206 the reference frame [Wu et al., 2010].

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208 NASA's ICES at mission measures long-term surface elevation changes in the Polar Regions 209 [Schutz et al., 2005; Zwally et al., 2002]. In this study, we use ICES at data in Greenland and 210 Antarctica from 2003.0-2010.0 to provide additional information on present-day ice mass 211 change. We process both ICES at RL633 and RL634 data in a consistent way, but report only 212 results from the RL 634 dataset. The major differences are that the RL 634 data have been 213 corrected for the Gaussian Centroid (G-C) error [Borsa et al., 2014], which will have impacts on 214 the ICES at inter-campaign bias (ICB). The ICB has been estimated to be in the range of 1-2 215 cm·a<sup>-1</sup> [Scambos and Shuman, 2016; Zwalley et al., 2015]. In our inversion, we do not correct for 216 ICB from the raw ICES at observations, but estimate the inter-campaign bias trend as an 217 additional parameter in the inversion. We assume that the ICB trend is constant in both 218 Greenland and Antarctica and estimate one auxiliary ICB trend parameter to mitigate possible 219 biases introduced by ICESat. We first perform data reduction based on a set of quality flags 220 indicating the health of the space segment, signal quality, saturation level, return surface 221 character, and cloud cover. Then we use along-track ground repeat track data from early 2003 to 222 the end of 2009 to calculate surface elevation changes accounting for topography, slope, and 223 elevation at each 1 km by 1 km grid cell. We convert the volumetric change to ice mass change 224 using estimates from regional climate models. In Greenland, the ICES at derived elevation rate is 225 first corrected for firn compaction rate [Sørensen et al., 2011] using results derived from a 226 regional climate model HIRHAM5 RCM [Lucas-Picher et al., 2012; Simonsen et al., 2013] within each grid cell. Then the corrected volumetric change is converted to a mass change rate 227 228 using density values derived from HIRHAM5. We use a simple glaciological criterion to 229 determine density in each grid box. We assume volume change is caused purely by ice dynamics 230 when the grid cell is located below the equilibrium line (at which the surface mass balance is

zero), and an ice density of 917 kg/m<sup>3</sup> is used for conversion. When the grid cell is located above 231 232 the equilibrium line, a composite density that represents the average density for all layers in the 233 vertical column is used during the conversion. In Antarctica, the volume to mass conversion is 234 based on the regional atmosphere and climate model RACMO2.3p2 [van Wessem et al., 2018]. 235 We first determine a firn compaction rate and its uncertainty for the period of ICES at operation 236 from the firn densification model (FDM). The FDM calculates time-varying firn depth and 237 density variations by accounting for the information on firn processes such as melt percolation 238 and refreezing, air content. We correct the ICES at elevation rate using the firn compaction 239 velocity. The elevation rate uncertainty includes both the standard deviation of ICES at and the 240 formal error of the firn correction. The residual elevation rate is considered the result of elevation 241 change due to ice dynamics only. In converting volumetric change to mass change for the ice 242 layer, we use ice density for the entire Antarctic ice sheet except where large positive residuals 243 are found. When the positive residual values are larger than the 95% confidence level, we 244 attribute the residual uplift to un-modeled snow accumulation, and an average firn density is 245 used. We then add the surface mass balance from the firn layer in the RACMO2.3p2 model to 246 derive the total mass change rate. The volume-to-mass conversion in Antarctica differs from 247 Greenland in that we assume the firn compaction corrected ICES at elevation rates are due to ice 248 dynamics only and mass balance in the firn layer is derived from the RACMO2.3p2 model 249 output. Lastly, we average ICES at results in both Greenland and Antarctica into 100 km by 100 250 km grid boxes, to be consistent with the spatial resolution of our global inversion (Figure 2). 251

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3. Global Kinematic Inversion Methodology

Our inversion separates PDMT and GIA signatures in the spectral domain using information from multiple datasets. GRACE measures geoid change rates due to both PDMT and GIA with near global coverage. Surface geodetic stations measure the sum of elastic and viscous crustal deformation rates in response to current and past mass redistribution. ICESat data are sensitive to current ice and firn volume changes and total bedrock uplift due to both elastic and viscous deformation in the Polar Regions. OBP products complement these data sets by providing partial PDMT constraints in the ocean basins.

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3.1 Measurement Models

263 The observation equations for ground geodetic velocities, GRACE geoid coefficient rates, and 264 OBP rates were described in the supplementary material of *Wu et al.* [2010]. These are listed and 265 briefly reviewed here for reference and ease of discussion, along with the new observation 266 equation for ICESat altimetry data. Throughout this paper, we will use a coordinate system with 267 its origin defined at the center-of-mass of the solid Earth (CE) with consistent load Love 268 numbers. In this system, the ground station velocities at  $(\theta, \varphi)$  relative to the center-of-mass of 269 the total Earth system (CM) can be written as:

$$270 \quad \dot{r} - \dot{r}_{cm} = \omega_p \times r + \frac{4\pi a^3}{M_E} \sum_{n=1}^{\infty} \sum_{m=0}^n \sum_{q=c,s} \frac{1}{2n+1} \times \left[ \left( h'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^{\nu,h} \right) Y_{nmq} \hat{e}_r + \left( l'_n \dot{M}_{nmq} + \dot{M}_{nmq}^$$

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$$\dot{M}_{nmq}^{\nu,l} \left( \partial_{\vartheta} Y_{nmq} \hat{e}_{\vartheta} + \frac{1}{\sin\vartheta} \partial_{\varphi} \partial_{\vartheta} Y_{nmq} \hat{e}_{\varphi} \right) \right] - \dot{r}_{cm} ,$$
 (1)

where the vectors  $\dot{r}$  and  $\dot{r}_{cm}$  are the station velocity and the velocity of CM in our coordinate system respectively.  $\omega_p$  is the angular velocity of the *p*th plate, p = 1, ..., 15 including major tectonic plates with at least 3 surface geodetic sites. *a* and  $M_E$  are the radius and mass of the Earth respectively.  $Y_{nmq}$  is the real normalized spherical harmonic function (with geodetic convention) of degree *n* and order *m*, with q=c, or *s* indicates the cosine or sine term. Note that

there is no sine term if m=0.  $h'_n$  and  $l'_n$  are vertical and horizontal elastic load Love numbers respectively.  $\dot{M}_{nmq}$  are the spherical harmonic coefficients of present-day surface mass density trend (a typical form of PDMT) in units of kg•m<sup>-2</sup>•a<sup>-1</sup>.  $\dot{M}_{nmq}^{v,h}$  and  $\dot{M}_{nmq}^{v,l}$  are spherical harmonic coefficients of GIA induced surface vertical and horizontal velocities respectively expressed in units of apparent surface mass density rate, and describe kinematic GIA signatures in ground geodetic data.  $\hat{e}_r$  and the like indicate unit spherical coordinate vectors.

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284 Nominally, the velocity of CM relative to CE (our coordinate origin) has two components,  $\dot{r}_{cm} = \dot{r}_{cm}^e + \dot{r}_{cm}^v$  where the superscript e and v indicate the PDMT and GIA contributions 285 286 respectively. GIA does not cause current variations of water mass in the surface layer other than 287 its passive response to the changing gravity potential due to GIA. This passive response is included as part of PDMT in our study. Therefore,  $\dot{r}_{cm}^{\nu} = 0$ . Consequently, geocenter motion 288 289 between CM and the center-of-figure of the solid Earth surface (CF) can be expressed as:  $V_G = \dot{r}_{cm}^e - \dot{r}_{cf}^e - \dot{r}_{cf}^v = \frac{4\pi a^3}{\sqrt{3}M_F} \times \left[ \left( \dot{M}_{11c} - \frac{\dot{M}_{11c}^h + 2\dot{M}_{11c}^l}{3} \right) \hat{e}_x + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^l}{3} \right) \hat{e}_y + \left( \dot{M}_{11s} - \frac{\dot{M}_{11s}^h + 2\dot{M}_{11s}^h +$ 290  $\left(\dot{M}_{10c} - rac{\dot{M}_{10c}^{h} + 2\dot{M}_{10c}^{l}}{3}
ight)\hat{e}_{z}
ight]$  , 291 (2)where  $\dot{M}_{1mq}^h = h'_1 \dot{M}_{1mq} + \dot{M}_{1mq}^{\nu,h}$ , and  $\dot{M}_{1mq}^l = l'_1 \dot{M}_{1mq} + \dot{M}_{1mq}^{\nu,l}$ . The first term for each 292 293 coordinate component on the right-hand side of equation (2) is in fact the component for  $\dot{r}_{cm} = \dot{r}_{cm}^{e}$ , which should be substituted into equation (1). The terms containing elastic load Love 294 numbers (not including the negative sign) are the components for  $\dot{r}_{cf}^{e}$ .  $\dot{r}_{cm}^{e} - \dot{r}_{cf}^{e}$  is thus the 295 296 PDMT contribution to geocenter motion. The remaining terms indicated by superscript v are the visco-elastic components for  $\dot{r}_{cf}^{\nu}$ , and  $\dot{r}_{cm}^{\nu} - \dot{r}_{cf}^{\nu} = -\dot{r}_{cf}^{\nu}$  is the GIA contribution to geocenter 297

298 motion.

300 Similarly, GRACE measured gravitational geoid rate spherical harmonic coefficients can be301 described by:

302 
$$\dot{N}_{nmq} = \frac{4\pi a^3}{M_E(2n+1)} \Big( (1+k'_n) \dot{M}_{nmq} + \dot{M}^{\nu,k}_{nmq} \Big),$$
 (3)

303 where  $k'_n$  is the potential load Love number, and  $\dot{M}^{\nu,k}_{nmq}$  is the spherical harmonic coefficients for 304 the GIA induced geoid change rate expressed in units of apparent surface mass density rate. 305 Here, n = 2, 3, ... 60 in GRACE's level-2 data release RL-6.

306

307 The JPL ECCO OBP product is derived in a closed oceanic system with no water mass input or 308 output. ECCO products use the Boussinesq approximation where the density variations are 309 ignored. As is usually done, we apply a mass conservation correction to the OBP values by 310 removing a uniform layer of excess water mass resulted from the Boussinesq approximation. For 311 such corrected OBP values, it is adequate to model the dynamic OBP as the difference between 312 the total PDMT OBP and its hydrostatic equilibrium part. In other words, the oceanographic 313 OBP is the bottom pressure caused by the water column between sea surface and the equilibrium 314 (time-varying) gravitational geoid changes due to PDMT. The latter surface change depends only 315 on instantaneous global surface mass distribution and reflects the self-attraction and loading 316 effects on the geoid. Again, the hydrostatic part due to GIA with present-day water redistribution 317 is considered part of PDMT here. Thus, at the oceanic grid points:

318

$$\dot{P} = g \sum_{n=1}^{\infty} \sum_{m=0}^{n} \sum_{q=c,s} \dot{M}_{nmq} Y_{nmq} - g \dot{M}_{EQ}, \tag{4}$$

where g is normal gravity on the Earth's surface.  $\dot{M}_{EQ}$  is hydrostatic equilibrium surface mass density over the oceans and is a function of  $\dot{M}_{nmq}$  only.

14

322 The ICES at surface elevation rates are modeled as:

323 
$$\dot{h} = \dot{b} + \frac{1}{\rho} \left[ \sum_{n=1}^{\infty} \sum_{m=0}^{n} \sum_{q=c,s} \dot{M}_{nmq} Y_{nmq} - \dot{M}_{atm} \right] + (\dot{r} - \dot{r}_{cm}) \cdot \hat{e}_{r},$$
(5)

where  $\dot{b}$  is a bias drift parameter to accommodate a possible drift error due to inter-campaign biases.  $\rho$  is ice or modeled firn density depending on the ground track location.  $\dot{M}_{atm}$  is the atmospheric mass change rate per unit surface area computed by ECMWF Re-Analysis ERA-Interim model. Equation (1) can be substituted into the bedrock uplift rate term (the third term on the right-hand side) so that the observation is completely described by the bias drift, PDMT, and GIA parameters.

330 331

### 3.2 Parameterization and a priori information

332 We parameterize our inverse problem similar to that of Wu et al. [2010], except that PDMT 333 coefficients are estimated up to higher degree and order to accommodate ICESat's high-334 resolution data. It is well known that in the approach of spherical harmonic inversion that the 335 truncated higher-degree terms, although small, will alias into the low-degree harmonic estimates 336 along with data noises. The total uncertainty, including the aliasing effect and data noise, can be 337 assessed using a properly constructed data and a priori parameter covariance matrix in a 338 constrained least square method. Here, the truncation degree is chosen in a trial-and-error 339 manner. Benefiting from the dense ICES at vertical velocity estimates in the Polar Regions, we 340 extend the PDMT spherical harmonic degree to 180 to account for large spatial variability of the 341 ice mass change. In doing so, we are able to limit the aliasing errors from high-degree PDMT 342 parameters to the lower-degree PDMT and GIA parameters.

343

We use a similar degree and order 60 parameterization for the GIA spherical harmonics, as the GIA related deformation is generally longer-wavelength in nature. These include vertical and

15

horizontal coefficients from n=1 to 60. The geoid GIA coefficients are approximately

347 proportional to the corresponding vertical coefficients [*Wahr et al.*, 1995]:

348 
$$\dot{M}_{nmq}^{\nu,k} \approx \dot{M}_{nmq}^{\nu,h} \times \frac{2}{2n+1} \text{ for } n \ge 2.$$
 (6)

However, this relation contains significant errors at low-degrees and does not account for the rotational feedback effects of the GIA process. We therefore retain  $\dot{M}_{nmq}^{v,k}$  for  $2 \le n \le 7$  as additional unknown GIA parameters, and use the approximate relation above to substitute higher-degree  $\dot{M}_{nmq}^{v,k}$  in equation (3) with vertical coefficients  $\dot{M}_{nmq}^{v,h}$  multiplied by corresponding proportionality constants in equation (6). As shown in the observation equations, our parameter vector also include 15×3 plate rotation parameters and a constant altimeter bias drift.

355

356 Despite the use of multiple globally distributed data sets, our inverse problem remains a rank 357 deficient and underdetermined problem. For example, the relative velocities used to remove ITRF origin drift are only sensitive to  $\dot{M}_{1mq}^{\nu,h} - \dot{M}_{1mq}^{\nu,l}$  rather the individual *n*=1 GIA coefficients. 358 359 As another example, ICESat's dense ground tracks require a high spherical harmonic truncation. 360 But away from the ice sheets, high spatial resolution data coverage above GRACE's truncation 361 degree of 60 is generally not available. To overcome these problems, careful assessment and 362 incorporation of a priori information is required. Instead of using mathematically convenient 363 models to stabilize the inversion, we construct our *a priori* parameter information based on 364 physically plausible Earth models. Since the values of PDMT are barely known before, we set all 365 a priori values of PDMT parameters to 0. The a priori GIA parameter values are computed from 366 the average of global ICE-5G and Antarctic IJ05 ice models based on a simplified VM2 Earth 367 rheology profile. The degree-2 and order 1 GIA coefficients are from Paulson et al. [2007a] that 368 is also based on ICE-5G and VM2 models but includes a revised rotational feedback mechanism

*[Mitrovica et al., 2005; Peltier, 2015].* The a priori values of plate rotations and ICESat bias drift
are all set to 0.

371

372 We construct a loose but plausible high-resolution *a priori* PDMT covariance matrix that 373 stipulates much higher PDMT variability over ice and land areas than over the oceans to 374 accommodate the previously observed large variability in ice mass balance and terrestrial water 375 storage. The covariance matrix also includes both random and correlated components in space. 376 With realistic high-resolution geographic boundaries, this *a priori* information will strengthen 377 isolation of signal sources and improve spatial resolution. Although the a priori GIA model is 378 fairly close to state-of-the-art, unfortunately, its uncertainties are largely unknown. To account 379 for plausible uncertainties in our kinematic GIA parameters, we construct an *a priori* covariance 380 matrix by perturbing the GIA model dynamically with large ice history and mantle viscosity 381 variations. We use a simplified layered viscoelastic Earth model based on VM2 with an elastic 382 lithosphere, two viscoelastic mantle layers, and an inviscid fluid core for our calculations and 383 propagate the covariance matrices to the kinematic parameter domain. The horizontal extent of 384 the deglaciation is fixed liberally over possible areas either confirmed or suspected in the 385 literature including Alaska. We use random ice thickness uncertainties that are roughly 75% of 386 the ICE-3G model values [Tushingham and Peltier, 1991]. Spatially correlated thickness errors 387 are also assumed with an uncertainty of 750 m and a correlation length of 600 km. For Greenland 388 and Antarctica, we include additional correlated thickness errors with an uncertainty of 1 km and 389 a correlation length of 250 km to reflect the less constrained deglaciation history there. We assume that the a priori upper and lower mantle viscosity values have uncertainties of  $4 \times 10^{20}$ 390 Pa s and  $7.8 \times 10^{21}$  Pa s. The covariance matrices of these errors are propagated to our 391

392 parameter domain and summed up with an additional small diagonal matrix so that the resulting 393 a priori covariance matrix is full ranked and roughly reflects the differences of various ice 394 history and Earth rheology models. It also includes cross-correlations between low-degree GIA 395 geoid coefficients and vertical coefficients and other intrinsic information on GIA dynamics. 396 While the ice thickness history and Earth rheology constraints are pretty loose, the dynamically 397 constructed full a priori GIA covariance matrix provides valuable tight constraints for the 398 remainder linear combinations of the GIA spherical harmonic model coefficients based on 399 physical laws that govern the GIA process, and global horizontal extent of historical 400 deglaciation.

401

To obtain a meaningful solution of the parameters and their realistic uncertainties, we adopted a
least squares estimation method with reduced a priori information. The observation equation is in
matrix form:

$$405 \quad L = HX + \Delta \tag{7}$$

where L is the  $l \times 1$  observation vector which includes the projected surface geodetic vertical and horizontal velocity, gridded OBP rate and height change rate of the ICESat, plus the GRACE geoid coefficients. Note that the displacement equations in equation (7) are the projected version of equation (1-5). *X* is the  $k \times 1$  parameter vector. Equation (7) is further re-organized to include the a priori parameter vector  $X_0$  by subtracting  $HX_0$  from both sides of equation (7):

$$411 \quad \Delta_L = H \Delta_x + \Delta \tag{8}$$

412 where  $\Delta_L = L - HX_0$  and  $\Delta_x = X - X_0$ . This observation equation can be solved using the 413 classical least squares method with a priori information. Here we adopt the method used in *Wu et* 414 *al.*, [2010] to improve the solution by using a customized singular value decomposition (SVD)

415 method to compare the data and parameter covariance matrices. The data and a priori parameter 416 covariance matrices  $C_{\Delta}$  and  $C_x$  are first square root decomposed into:

$$417 \quad C_{\Delta} = R_{\Delta}^{-1} R_{\Delta}^{-T} , \qquad (9)$$

418 
$$C_x = R_x^{-1} R_x^{-T}$$
, (10)

419 where  $R_{\Delta}$  and  $R_x$  are upper triangular square root information matrices. We then multiply

420 equation (8) by 
$$R_{\Delta}$$

$$421 \quad R_{\Delta}\Delta_{L} = R_{\Delta}H\Delta_{x} + R_{\Delta}\Delta \,. \tag{11}$$

422 we re-write equation (11) in terms of the normalized vectors:

$$423 \quad y = Bz + \delta \quad , \tag{12}$$

424 where the normalized data, noise and parameter vectors are defined as

425 
$$y = R_{\Delta}\Delta_L, \delta = R_{\Delta}\Delta, z = R_x\Delta_x,$$

426 and 
$$B = R_{\Delta} H R_X^{-1}$$
.

427 We obtain the solution to equation (12) using singular value decomposition of the new

$$429 \quad B = UAV \,, \tag{13}$$

- 430 where U and V are  $l \times l$  and  $k \times k$  orthogonal matrices, and  $\Lambda$  is the  $l \times k$  rectangular diagonal
- 431 matrix, where the first  $k \times k$  diagonal matrix containing the singular values  $\lambda_i$  in descending
- 432 order. We define the first  $k \times k$  diagonal matrix of  $\Lambda$  as

433 
$$\Lambda_k = \begin{pmatrix} \Lambda_1 & 0\\ 0 & \Lambda_2 \end{pmatrix}$$
(14)

- A cutoff threshold  $\lambda_c$  is set to separate all singular values into group  $\Lambda_1$  and  $\Lambda_2$ , where all values
- 435 in  $\Lambda_1$  are greater than  $\lambda_c$ , and  $\Lambda_2$  includes all singular values less or equal to  $\lambda_c$ .
- 436 We obtain the final solution as:

437 
$$\hat{X} = X_0 + R_x^{-1} V^T \Gamma U^T R_\Delta (L - H X_0)$$
 (15)

438 where the regulation matrix  $\Gamma$  defined as

439 
$$\Gamma = \begin{pmatrix} \Lambda_1^{-1} & 0 & 0 \\ 0 & (\Lambda_2^2 + I)^{-1} \Lambda_2 & 0 \end{pmatrix}$$
 (16)

The key elements in the inversion are the data and a priori parameter covariance matrices and the cutoff threshold  $\lambda_c$ . The cutoff threshold reflects the relative strength of data and a priori information to the solution. It is apparent that, when the singular values are large, the solution does not depend on the *a priori* model. This avoids unnecessary contamination from possible large model errors. However, the *a priori* model helps to stabilize the inversion when the singular values are small. The choice of  $\lambda_c$  is done in an ad hoc manner and we use the value of 1 in our study.

447

448 To validate our inversion algorithm and demonstrate the improved resolution in the Polar 449 Regions from added ICES at data, we perform a simulation study using synthetic data with and 450 without ICES at data in Greenland (Figure 3). We generate synthetic surface 451 geodetic/ICESat/OBP/GRACE observations from reference GIA (Figure 3a) and PDMT 452 scenarios as well as MORVEL plate model [DeMets et al., 2010]. We also generate data noise 453 according to the data covariance matrices from real observations. To convert PDMT into ICESat elevation change, we use a uniform firn density of  $600 \text{ kg/m}^3$ . For simplicity, we include only 454 455 Greenland ICES at data in the simulation. Adding more ICES at data in Antarctica will not change 456 our conclusion. We compare inversion results with and without ICES at data to show the 457 achieved improvement. In the presence of noise, the inversion without ICES at recovers the truth 458 to within uncertainty (Figure 3b). Nevertheless, when we include the ICES at data, the truth is 459 recovered with better accuracy (Figure 3c).

20

460

461 ICESat derived mass change rates depend on the average density profile used for the firn layer.
462 As presented in the above section, the density profile is determined from ICESat residual values
463 after the firn compaction correction. The ICESat residual velocity is a function of GIA and
464 PDMT induced vertical velocities. Thus our method requires iteration to update the firn density.
465 We first compute the surface uplift from our initial GIA and PDMT values, then we iterate our
466 inversion to update the firn density profile. Our inversion commonly converges within 1-2
467 iterations, indicating small GIA and PDMT vertical rates compared to ICESat elevation rates.

469

#### 4. Inversion Results

470 Table 1 summarizes the key parameters in our inversion. Our PDMT results show strong mass loss in Greenland and West Antarctica, up to  $10 \text{ cm} \cdot a^{-1}$  water equivalent mass change (Figure 4). 471 The average total mass loss rate from 2002-2010 is  $126.4\pm18.4$  GT  $\cdot a^{-1}$  for Antarctica and 472  $203.3\pm3.1$  GT·a<sup>-1</sup> for Greenland. Our Greenland mass loss estimate is significantly larger than 473 474 that of Wu et al. [2010], who accidentally used a smaller Greenland land mask function. Their revised estimate with the new land-sea mask yields  $160\pm16$  GT·a<sup>-1</sup> mass loss [*Wu et al.*, 2011]. 475 476 Despite the large inter-annual variability in mass loss rate as shown in *Shepherd et al.* [2020], our Greenland estimate agrees well with the latest IMBIE Greenland result. Based on the IMBIE 477 datasets, the average Greenland ice sheet mass change rate is 199  $GT \cdot a^{-1}$  from 2002.0-2010.0. 478 479 We include a handful of GPS sites in Greenland that were installed after 2007, these sites are 480 recording faster melting rate between 2007-2010 that can lead to slightly increased overall 481 melting rate in Greenland. Our new estimate in Greenland is smaller than recent GRACE based 482 results because our high-resolution inversion is able to separate mass losses in Greenland proper 483 from those over its peripheral areas. Mass redistribution due to glacier melting has been found in

484 the Canadian Arctic Archipelago (CAA), Iceland, Svalbard, the Arctic Ocean, and SE Alaska, 485 the Antarctic Peninsula, Patagonia, and high mountain glacier regions in Asia. We report an average of 5.6 $\pm$ 16.7 GT·a<sup>-1</sup> mass change rate in East Antarctica over the period of 2002-2010, 486 487 statistically indistinguishable from zero. Our results reconcile with other results in East 488 Antarctica averaged over the past two decades [Shepherd et al., 2018]. However, we recall the 489 large inter-annual variability in surface mass balance in Antarctica, and therefore, comparing 490 results of different periods may lead to incorrect conclusions. Although the coseismic 491 deformation patterns for the two largest earthquakes during the study period are removed from 492 the geoid change rate, a residual pattern can be seen in this region possibly due to slightly 493 different post-seismic deformation patterns. But such aliasing effects are very much localized 494 and will not impact our global results. We find a general pattern of mass increase in the US high 495 plains and the Prairies in Canada, likely caused by increased precipitation. The patterns are 496 significant, because the uncertainties for PDMT are generally an order of magnitude smaller than 497 the PDMT values themselves. The largest uncertainty is located in the interior of Antarctica, due 498 to lack of land-based GPS data.

499

Our method provides a robust estimation of global GIA and its uncertainty. Although our GIA
estimate is different from other global GIA models as it also includes the effect from recent ice
mass changes, it recovers the large-scale global GIA pattern in recent models, while differs
considerably in some regions (Figure 5). The most striking difference is the uplift found in West
Antarctica, near the Amundsen Sea Embayment (ASE). The uplift center near the ASE is
suggested to be due to the fast mantle viscous response to recent ice melting [*Nield et al., 2014*].
Recent seismic tomography studies reveal drastically different S wave velocity contrast for East

507 and West Antarctica, with the latter inferred to have an anomalously high upper mantle 508 temperature and a thin lithosphere [An et al., 2015]. This combination of high mantle 509 temperature and thin lithosphere likely leads to the fast visco-elastic response to the melting in the recent past, which induces regional uplift rates of more than 20 mm ·a<sup>-1</sup> near ASE. Few GIA 510 511 models consider lateral heterogeneity in the Earth structure [van der Wal et al., 2015] and recent 512 ice melting [*Nield et al.*, 2014], making a direct comparison of GIA models with our results 513 difficult in areas similar to ASE. Our inversion shows enhanced viscous response in areas of low 514 effective viscosity, thin asthenosphere and rapid melting in the recent past, such as Southeast 515 Alaska, the west coast of Canada, Iceland, Svalbard, and the Antarctica Peninsula. Global GIA 516 models built on data from other regions, and data from before the ice change in the last centuries 517 (e.g. RSL data) generally do not show enhanced viscous response of GIA, and can introduce 518 errors if used to correct observations for present-day ice mass change estimates.

519

520 We separate large-scale PDMT and GIA-induced deformation and geoid changes by conducting 521 an inversion in the spectral domain with a global distribution of multiple observation types 522 (Table 1). Our results show a GIA-induced  $J_2$  similar to the one found by Wu et al. [2010]. It is 523 also very close to the SLR total  $J_2$  observation before the early 1990s [Roy and Peltier, 2011]. 524 PDMT  $J_2$  shows a large positive value owing to the increased Earth oblateness from polar ice 525 melting since the 1990s [Roy and Peltier, 2011]. Our determined PDMT geocenter motion is small, with -0.16 $\pm$ 0.08 mm·a<sup>-1</sup> along the Z-axis, contributing about 15 GT·a<sup>-1</sup> to mass change in 526 527 Antarctica. The values of the separate components differ from those in *Wu et al* [2010], most 528 likely due to the increased a priori uncertainties in long-wavelength OBP rates and thus their 529 reduced weights in the inversion. The GIA-induced geocenter motion estimate remains very

similar to that of *Wu et al.* [2010], which is significantly larger in magnitude than -0.4 - -0.5mm·a<sup>-1</sup> indicated by traditional GIA models and much larger than that induced by PDMT.

533 5. Discussions and conclusions

534 Although advances have been made in GIA modeling during the past decades, e.g. Peltier et al. 535 [2018] and quantification of GIA model uncertainty [Caron et al., 2018; Li et al., 2020], large 536 discrepancies remain in different GIA models and model-based GIA corrections to PDMT 537 [Shepherd et al., 2018]. Further, errors in GIA models are difficult to quantify because most are 538 built based on a forward-fitting approach or partial domain inverse that suffers from non-539 uniqueness in the obtained earth and ice history parameters (e.g. Paulson et al, 2007b) and 540 limited coverage of GIA observations which can bias the predictions in data-sparse regions. 541 Subsequent iterative modifications have also been carried out to adjust the load and rheology 542 parameters to forward-fit additional data. Also, most GIA models use a spherically symmetric 543 Earth structure, which ignores the impact of lateral homogeneity. More advanced numerical 544 models with realistic rheology are computationally expensive, and require more detailed 545 observations of Earth composition. Our inversion benefits from relatively dense globally 546 distributed observation networks. By exploiting intrinsic relations among gravity and 547 deformation due to PDMT and GIA, we separate their signatures and provide realistic 548 uncertainty estimates. A spherically symmetric Earth structure is used and only a few depthdependent rheological parameters are perturbed in the construction of the a priori GIA 549 550 covariance matrix. However, very heterogeneous ice history and large mantle viscosity 551 uncertainties are allowed. Given the ambiguities between ice history and Earth rheology in 552 interpreting modern geodetic data, possible lateral heterogeneity in Earth rheology should not

have a significant impact on our kinematic inversion, because geodetic GIA signatures due to
lower upper mantle viscosity may be compensated by indiscernible signatures due to stronger
melting and vice versa.

556

557 Rapid visco-elastic response to deglaciation in the recent past in areas of low upper mantle 558 viscosity and thin crust is also suggested by our results. Most affected zones are near subduction 559 zones (e.g. Western Canada and SE Alaska; Patagonia) or close to known weak mantle zones 560 (e.g. Iceland, ASE). Inferred upper mantle viscosity from seismological and geodetic data in those regions are near 10<sup>18</sup> Pa·s [Larsen et al., 2005; James et al., 2009], two orders of 561 562 magnitude lower than the global average value. This brings the load-induced solid earth response time down to  $10^{1}$ - $10^{2}$  years. Our result validates those regional models, and on top of that 563 564 provides estimates of low-degree spherical harmonic coefficients that can improve the accuracy 565 of our regional estimates and hence the accuracy of mass balance estimates obtained from joint 566 inversion. We compare our global present-day GIA geoid rate with three other dynamically 567 constructed GIA models (Figure 6). Even though our a priori GIA model differs from more 568 recent GIA models, the long-wavelength pattern of these models [Peltier et al. 2018; Caron et al. 569 2018; A et al., 2013] is recovered well, which could indicate that effects of lateral heterogeneity 570 that is allowed in our inversion are relatively small. Our GIA  $J_2$  is very close to that of ICE-6G 571 and pre-1992 SLR observations [Roy and Peltier, 2011], although our inverted GIA geoid 572 includes the effect of enhanced GIA. This indicates a small effect of 3D Earth structure to global 573 scale deformation or this effect is balanced by a slightly different values of mantle viscosity. In 574 addition, we find our result showing enhanced GIA viscoelastic responses in areas with possible 575 recent past ice mass changes. In the Polar Region, our GIA geoid shows enhanced positive geoid

576 rates in SE Alaska and Western Canada, Iceland, Svalbard, ASE, and in the coastal areas of East 577 Antarctica (Figure 7). We also find a negative geoid rate in southeast Greenland that is less clear 578 in other models. This negative anomaly does not affect the accuracy of PDMT estimates in 579 Greenland as our result agrees well with the latest IMBIE estimates in Greenland. We compare 580 our GIA results and ICE-6G in Greenland and find the absolute difference between them is less 581 than 0.2 mm·a<sup>-1</sup>. Such a difference is possible from variations in 3D Earth structures [e.g. *Milne* 582 et al., 2018]. We speculate this signal could be real geophysical signal e.g. influence of the 583 lateral heterogeneous Earth structure [*Milne et al.*, 2018], past ice mass fluctuations, or a 584 combination of both. Because GIA induced bedrock uplift serves to reduce the water depth near 585 the marine based ice sheet, it acts to slow down the rate of ice loss [Gomez et al., 2010]. In 586 regions of low mantle viscosity (e.g. ASE), the stabilizing effect will be stronger. Our result can 587 provide better estimates of current vertical rate in those regions, and contribute to better 588 modeling of the coupled effect of tectonic and ice dynamics in marine based ice sheets. 589 Furthermore, using an ice history that includes late Holocene ice mass change and accounting for 590 Earth's lateral structural heterogeneity will improve estimates of present-day GIA signatures on 591 both local and global scales for the forward models.

592

593 To date, most data-driven GIA models developed are used to understand regional GIA or PDMT
594 signals [*Gunter et al.*, 2014; *Martín-Español et al.*, 2016; *Sasgen et al.*, 2017; *Simon et al.*,

595 2017]. By including multiple global datasets, our simultaneous global inversion separates long-

596 wavelength PDMT and GIA signatures and avoids the related biases in the regional approaches.

597 On the other hand, because our method is conducted in the spectral domain, the small scale

results may perturbed by the truncation error. We improve the accuracy of our method by

including a high-resolution land-sea mask, incorporating a physically plausible a priori information, and a modified SVD solver to solve the least square problem. The low-degree GIA spherical harmonic coefficients offer insights into deep Earth rheology and global change history. Our results confirm that GIA-induced geocenter velocity is larger than the standard model predictions [*Wu et al.*, 2010], but GIA-induced  $J_2$  is consistent with the model value adopted from earlier SLR observations [*Peltier*, 2004] and in stark contrast with the – (6.0-6.5)×10<sup>-11</sup> a<sup>-1</sup> value based on the SLR/AR5 scenario [*Nakada et al.*, 2015].

607 Although it is not a dynamic inversion for earth rheology and ice history, our unique approach 608 provides a robust separation of GIA signatures from those of PDMT. We provide a snapshot 609 view of GIA and PDMT at present-day with realistic estimations of uncertainties. The derived 610 GIA rates, along with their uncertainties, can be used to reveal weakness in the data sets and 611 chart for future improvements. Our method has the potential to include more observations, such 612 as paleo relative sea level (RSL) data to conduct a global inversion with PDMT and dynamic 613 GIA parameters. Current ice history models are largely built by forward-fitting the RSL records 614 without a quantitative uncertainty assessment. This unfortunate situation led to the very 615 conservative a priori GIA information which has to discard a large amount of information 616 contained in the global RSL records. A dynamic global inversion incorporating RSL data will 617 not only remedy this problem, but also gain insights into deep earth rheology, reconcile the 618 difference in mantle viscosity from geodetic and seismic data, and further constrain ice history, 619 in an unambiguous way.

620

The causes and the validity of the ICB are still under debate [Borsa et al., 2019]. Given the 621 622 relatively small ice sheet surface extent, the impact of the ICB on Greenland mass balance 623 estimation is relatively small. However, owing to the large extent of the Antarctic ice sheet, 1  $\text{cm} \cdot \text{a}^{-1}$  of ice elevation change in Antarctica roughly translates to 100 GT \cdot \text{a}^{-1} mass change. This 624 625 implies that the accuracy of ICES at ICB estimates is critical when evaluating Antarctic ice mass 626 balance. Conventionally, the ICBs are validated/calibrated with independent in situ 627 measurements over time [Zwalley et al., 2015; Borsa, et al., 2019; Schröder, et al., 2017]. In 628 this study, we take advantage of the globally distributed datasets to estimate the ICB trend 629 directly from data. We use RL634 GLA12 data products (Antarctica and Greenland Ice Sheet 630 Altimetry Data). It is reasonable to assume a constant ICB trend for ICES at measurement over 631 the ice sheets in Greenland and Antarctica. We parameterize the ICB trend into our inverse 632 formulization. Although our parameterization method does not determine ICBs on individual 633 laser campaigns, the trend estimation has a small uncertainty, probably reflects the improved 634 statistics from stacking a large amount of data. Our ICB trend agrees well with the latest result of Schröder, et al., [2017], who estimated ICB trend (L2a-L2d) to be 6.8±4.1 mm·a<sup>-1</sup>., although 635 636 we notice a large increase of ICB trend in their reported result if the last two laser campaigns are 637 included. The amount of data from the last two ICESat laser campaigns are relatively small, and 638 will have very limited overall impact on our estimated trend. For the same period, Borsa et al., [2019] determines the ICB trend to be  $2\pm4 \text{ mm}\cdot\text{a}^{-1}$ , suggesting a statistically insignificant ICB 639 640 trend in their study area (The salar de Uyuni, Bolivia). One source of the spatially varying ICB 641 trends is the switch of ITRF reference frames during the course of the ICES at mission. The 642 ICES at elevations were geolocated to the ITRF2000 (on and before Nov 5, 2006) and ITRF2005 643 (on and after Nov 6, 2006) separately [Urban et al., 2011]. This complication in reference frame

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introduces a -5.8±0.3 mm translation and a -1.8±0.3 mm  $\cdot a^{-1}$  translation rate in the Z axis start on 644 645 the epoch of 2000.0. This origin drift in the Z-axis will have limited impact on height change in 646 the tropical area, and mostly will manifest itself as horizontal deformation. But most of it will 647 translates into height change in Antarctica. Since none of the studies correct for this frame 648 change, it will introduce a cumulative subsidence of 2-3 cm for ICESat derived products 649 (campaign L2g-L2d) in Antarctica and 1-2 cm uplift in Greenland. Although we did not consider 650 the effect of frame translation in ICES at height rate as its overall effect will be corrected by the 651 additional ICB trend parameter. It is advised that future studies utilizing the ICES at data should 652 correct the bias introduced by switching the ITRF frames.

653

654 In this paper, we present results from a mostly data-driven global inversion method that provides 655 estimates of PDMT and kinematic GIA signatures with their uncertainties. The PDMT and GIA 656 results are only weakly dependent on the a priori GIA model, and can be used to validate the 657 dynamically constructed forward model, or to provide GIA corrections to the GRACE-FO 658 mission. Our results show enhanced GIA signals in regions of low mantle viscosity and recent 659 deglaciation that are consistent with regional studies, but which were not present in global GIA 660 model results. We provide new estimates of low-degree GIA spherical harmonic coefficients that are related to deep earth rheology. Our results confirm the earlier GIA  $J_2$  estimation from SLR 661 662 data. The estimated ICESat inter-campaign bias trend is similar to the most recent study using in-663 situ data in Antarctica.

664

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		Velocity (mm $\cdot a^{-1}$ )
	GIA Geocenter Motion (CE w.r.t. CF):	
	X	-0.11±0.01
	Y	0.13±0.02
	Z	-0.72±0.06
	PDMT Geocenter Motion (CM w.r.t. CF):	
	Х	-0.19±0.07
	Y	0.07±0.08
	Z	-0.16±0.08
	ICESat elevation bias drift	6.03±1.70
	Non-steric global sea level	1.37±0.13
	†PDMT induced $\dot{J}_2$	$(6.0\pm0.5)\cdot10^{-11}$
	†GIA induced $\dot{J}_2$	(-3.9±0.3)·10 <sup>-11</sup>
+		Mass change rate $(GT \cdot a^{-1})$
	Alaska	-62.4±4.8
	Antarctica: Total	-126.4±18.4
	West Antarctica	-111.8±4.0
	East Antarctica	5.6±16.7
	Antarctica Peninsula	-20.3±2.2
	Greenland	-203.3±3.1
	Canadian AA	-49.4±5.6
	Arctic Ocean	-1.5±8.7

Table 1. Geocenter motion and regional mass transportation from our joint global inversion.

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897 † Unit for J_2 is a^{-1}
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Figure 1. Distribution of global datasets. We use surface geodetic station velocities (black dots,
GNSS/VLBI/SLR), ocean bottom pressure from the ECCO model (black triangles), ICESat (red
dots) elevation change rate over the Greenland and Antarctic ice sheets, and GRACE spherical
harmonic coefficients (not shown here). All data are sampled from 2002-2010.

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906 Figure 2. left. Enlarged map of Antarctica ICESat elevation change rate and uncertainty. We 907 have excluded ICES at data over floating ice shelves. Each colored dot represents a 100 by 100 908 km grid box on the ice sheet. The average vertical rates within that grid box are color-coded. The 909 rate uncertainty defined as 2-sigma confidence values of the residual elevation rates after 910 subtracting the mean rate of the 100 km by 100 km grid. The ICES at elevation rate uncertainty 911 reflects not only the noise character of ICES at data itself, but also changes in ice surface 912 topography, and heterogeneity in ice dynamics within the grid. We find that data uncertainties on 913 the periphery of the ice sheet are 2-3 times larger than in the interior, reflecting a smoother 914 topography and less change in ice dynamic behaviors in the interior of the ice sheet. Note the 915 difference in color scale for the Antarctica and Greenland plot. Right. Enlarged map of 916 Greenland ICES at elevation change rate and uncertainty.

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919 Figure 3. Global simulation results. We have combined an arbitrary global GIA geoid rate truth 920 with PDMT truth. Both GIA and PDMT truth are truncated to spherical harmonic degree and 921 order 60. We then produce synthetic datasets for all surface geodetic sites, ICESat (Greenland 922 only), OBP and GRACE measurements. We show inversion results with and without ICESat 923 data to highlight the improvements of including ICES at data in Greenland. a). input GIA geoid 924 rate truth. b). inversion results without including Greenland ICESat data. c). by including ICESat 925 data in Greenland, we are able to recover GIA signal in Greenland better. The result shows 926 negligible difference in Greenland GIA with the truth data.

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Figure 4. Global distribution of present-day mass transport (PDMT) from 2002-2010 from the global inversion, expressed in equivalent water height (EWH) in  $cm \cdot a^{-1}$ . A 350 km Gaussian filter is applied to the PDMT spherical harmonic coefficients to smooth the result. The lower panel shows the uncertainty associated with the PDMT estimates, expressed in EWH. Note that most areas have uncertainties less than 0.5 cm  $\cdot a^{-1}$ .

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Figure 5. Estimated GIA geoid rate and its uncertainty from the global inversion. Upper figure
shows GIA induced geoid change rate, with positive anomaly in red and negative anomaly in
blue. We also calculated uncertainty of the GIA geoid, which is roughly one order of magnitude
smaller than the GIA geoid rate. The two regions of larger uncertainty are found in Far North
Russia and East Antarctica, presumably related to the lack of land geodetic stations to separate
GIA from PDMT.

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944	Figure 6.	Compari	son of glo	bal GIA	geoid rate	estimates.	We com	bare our i	inverted	GIA :	geoid
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- rate with three model results, C18 [*Caron et al.*, 2018], AW13 [*A et al.*, 2013], and ICE6G-D
- 946 [*Peltier et al.*, 2018]. All three geoid rates have been smoothed with a 300km Gaussian filter and
- 947 are downloaded from JPL's PODAAC data center (Last accessed, 2020-04-25).

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- 950 Figure 7. Comparison of GIA geoid rate in the polar regions. The left panel are plots of the
- 951 ICE6G-D [*Peltier et al.*, 2018] geoid rate in the North and South Pole. The right are two plots of
- 952 our GIA inversion results in the same region.

Accepted













