Constraining Glacial Isostatic Adjustment with Horizontal GPS Velocities in Antarctica

Tim H.J. Hermans
CONSTRaining glacial ISostatic Adjustment with horizontal GPS Velocities in Antarctica

by

Tim H.J. Hermans

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Supervisor: Dr. ir. W. van der Wal
Thesis committee:
  Prof. dr. L.L.A. Vermeersen
  Dr. ir. W. van der Wal
  Dr. R.E.M. Riva
  Dr. S.A. Sherman

Delft University of Technology
Delft University of Technology
Delft University of Technology
Ohio State University
Abstract

Glacial isostatic adjustment is the viscoelastic response of the Earth to ice and ocean loads. In forward models of glacial isostatic adjustment, mantle viscosity is often assumed to be laterally homogeneous. However, a lateral transition in shear wave velocities suggests a sharp transition in viscosity between West and East Antarctica. Along this transition, horizontal GPS velocities of ANET/POLENET West of the Ross Sea Embayment point towards the ice load rather than away from it. It is unclear why, as the dependency of horizontal velocities on viscosity is not well understood. In this thesis, this dependency is clarified, and it is investigated with a 3D finite-element model if the horizontal GPS velocities can be used to constrain the viscosity transition.

It was found that horizontal velocities point away from the ice load for viscosities of $10^{20}$ Pa s and lower, whereas for $10^{21}$ Pa s and higher their direction is reversed. The results in this thesis show that the GPS measurements at the Ross Sea Embayment likely require a lateral viscosity transition. Preferred viscosities in the upper mantle are found to lay between $10^{18}$ and $10^{19}$ Pa s at the West Antarctic side of the transition, and between $10^{21}$ and $10^{22}$ Pa s at the East Antarctic side. The results demonstrate that horizontal GPS velocities can be used to constrain lateral variations in rheology. As more studies will start to use 3D Earth models, horizontal GPS velocities should be used as one of the primary constraints of glacial isostatic adjustment, since their direction can be reversed depending on mantle viscosity.
I would like to express my sincere gratitude to my daily supervisor Dr. Ir. Wouter van der Wal. Working with Wouter has been a real pleasure. Wouter has provided me with a lot of opportunities to present and discuss my work with other researchers in the field. Moreover, we could always talk about everything, both related and unrelated to my thesis. Thank you for going way beyond of what is expected from a supervisor.

This work would not have been the same without the valuable inputs of several people. My special thanks go to Dr. Stephanie A. Konfal of Ohio State University for allowing us to compare our predictions with the latest ANET/POLENET GPS solution, for her valuable other inputs to this work, and for answering my endless stream of questions. I would also like to express my gratitude to Prof. Dr. Michael G. Bevis of Ohio State University for transforming my GIA predictions to the GPS reference frame and to Prof Dr. Terry J. Wilson of Ohio State University and Dr. Pippa Whitehouse of Durham University for their comments and contributions to this thesis. Thank you also Pippa for being so helpful during my thesis and during the IAG conference in Iceland, and providing me with your scripts for GMT postprocessing. Prof. Dr. Patrick Wu of the University of Hong Kong is acknowledged for his support regarding the GIA model, Prof. Dr. Matt King of the University of Tasmania for pointing us to the possible effect of postseismic velocities in Victoria Land, and PhD candidate Ir. Bas Blank of Delft University of Technology for his advice in the beginning of my thesis. I am also indebted to the department and the European Space Agency, for making it possible to present my work at the 1st Circular Workshop on Glacial Isostatic Adjustment and Elastic Deformation of the International Association of Geodesy.

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I hope that the results of this master thesis may make a meaningful contribution to our knowledge of glacial isostatic adjustment.

Tim Hermans
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<tr>
<td>AIS</td>
<td>Antarctic Ice Sheet</td>
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<tr>
<td>ANET</td>
<td>Antarctic Network of POLENET</td>
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<td>AP</td>
<td>Antarctic Peninsula</td>
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<td>BP</td>
<td>Before Present</td>
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<td>CETOL</td>
<td>Creep Error Tolerance</td>
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<td>Core Mantle Boundary</td>
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<td>FEM</td>
<td>Finite Element Method</td>
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<td>GIA</td>
<td>Glacial Isostatic Adjustment</td>
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<td>GPS</td>
<td>Global Positioning System</td>
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<td>LGM</td>
<td>Last Glacial Maximum</td>
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<td>LM</td>
<td>Lower Mantle</td>
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<tr>
<td>POLENET</td>
<td>Polar Earth Observing Network</td>
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<tr>
<td>PREM</td>
<td>Preliminary Reference Earth Model</td>
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<td>RMS</td>
<td>Root Mean Square</td>
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<td>RSE</td>
<td>Ross Sea Embayment</td>
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<td>RSL</td>
<td>Relative Sea-Level</td>
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<td>TAM</td>
<td>Transantarctic Mountains</td>
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<td>TZ</td>
<td>Transition Zone</td>
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<td>UM</td>
<td>Upper Mantle</td>
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In this chapter, first Glacial Isostatic Adjustment will be introduced in Section 1.1. Next, the challenges of modelling and observing Glacial Isostatic Adjustment in Antarctica will be explained. In Section 1.2, the heterogeneous structure of the Earth beneath Antarctica, and its effect on horizontal velocities, will be discussed. The Antarctic Ice Sheet will be introduced in Section 1.3, and it will be explained that observations of Glacial Isostatic Adjustment in Antarctica are scarce. Next, a remarkable feature of the horizontal GPS observations at the Ross Sea Embayment in Antarctica will be presented in Section 1.4. In Section 1.5 opportunities to advance the scientific knowledge about Glacial Isostatic Adjustment will be pointed out, after which a set of research questions will be presented. The chapter will end with a brief outline of the structure of the report in Section 1.6.

1.1. **Glacial Isostatic Adjustment**

During a glacial period, variations in solar intensity lead to a decrease in temperature. As a result, ice sheets start to form on the surface of the Earth, and the sea-level drops. The ice sheets that accumulated in the most recent glacial cycle were at their largest at the Last Glacial Maximum (LGM). The subsequent ice melt started about 21,000 years ago (Peltier, 2004). The viscoelastic response of the solid Earth to the on-going spatiotemporal variation of ice and ocean loads on the surface since the LGM is called Glacial Isostatic Adjustment (GIA). Isostasy refers to the equilibrium of the Earth's crust and mantle. The crust of the Earth floats on top of the upper mantle (UM) at a height that depends on its thickness and weight (and thus density).

Due to the massive weights of the ice sheets that accumulated during the last glacial cycle, the surface of the Earth deformed as prescribed by isostatic equilibrium. The relief of the weight of the ice sheets after the LGM resulted in a delayed rebound. As this rebound is a process that takes place on large time scales, the resulting deformations of the Earth's surface are noticeable in geodetic measurements to date, even though most major ice sheets have completely disappeared. The relaxation time that governs the isostatic readjustment depends on the material properties of the Earth, and specifically on the viscosity of the mantle. The evolution of the loading on the Earth is not only a result of the decreasing mass of (former) ice sheets, but also of the redistribution of the meltwater that was distributed over the surface of the Earth. Historic records show the associated rise in relative sea-level (RSL), which is measured between the sea surface and a local moving datum at the continental surface.

The redistribution of surface mass can be observed with gravity rate measurements with e.g. the Gravity and Climate Recovery Experiment (GRACE) mission (Velicogna and Wahr, 2006). Past inferences for the Antarctic Ice Sheet (AIS) derived from GRACE measurements show an ice mass loss...
of \(-143 \pm 73 \text{ Gt/yr}\) between 2002 and 2009 (Velicogna, 2009). However, the readjustment of the solid Earth also changes the gravity field of the Earth, which contaminates the measurements. The contribution of GIA can be isolated with predictions from inversion of data sets with different sensitivities to GIA, e.g. (Riva et al., 2009, Wahr et al., 2000, Wu et al., 2010). Often a combination of satellite altimetry and gravity rate measurements is used. GIA can also be predicted with geophysical forward models (Peltier, 1976). The two main inputs of such models are the ice loading and unloading derived from a deglaciation history, and a parameterization of the structure of the Earth. In Figure 1.1 the predicted uplift rates are shown for a suite of different forward and inverse models of GIA in Antarctica (Martín-Español et al., 2016). It can be seen that there is a lot of variability in both the magnitude and the spatial pattern of the estimated uplift rates between the different models. When the predictions of forward models are compared with observations of GIA, valuable information about the deglaciation history and the properties of the Earth can be inferred. With a better understanding of the ice and the response of the Earth, GIA can be predicted more accurately. This leads to better estimates of the ice loss of the AIS and its contribution to global and regional sea-level rise. Furthermore, a better understanding of the Earth's structure and especially of its mantle viscosity is key to understanding mantle convection (King, 1995). Thus, inferring mantle viscosity is important for processes such as plate movement, subduction zones, earthquakes and the prediction of post-seismic velocities.

The viscoelastic problem of GIA in forward models (see Section 2.2) has traditionally been overcome by solving the associated elastic problem in the Laplace or Fourier domain (Peltier, 1974). This spectral method is not suited for modelling non-linearity and lateral variations in rheology, which is why most studies on GIA assume an Earth structure that varies radially only (1D). In contrast with the spectral method, Finite-Element Method (FEM) models, when coupled to the Laplace equation (Wu, 2004), are suited to include lateral variations in mantle rheology as well (3D). A coupled FEM-Laplace model, based on the approach of Wu (2004) and Wang and Wu (2006), was developed...
and verified at the Astrodynamics & Space Missions Department (van der Wal et al., 2015) and is used for this study. A few alternative modelling approaches exist. Latychev et al. (2005) developed a finite-volume model, but this can only incorporate eustatic changes in sea-level. Additionally, perturbation theory can be used (e.g. (Tromp and Mitrovica, 2000)), but convergence problems limit the order of magnitude of viscosity variations that can be modelled. Because of these limitations and the fact that the 3D coupled FEM-Laplace model is readily available, alternative models are not considered here.

1.2. Lateral Heterogeneity beneath Antarctica

There are significant uncertainties in the representation of the Earth’s interior underneath Antarctica in forward models of GIA. In order to limit model complexity, the structure and rheology of the solid Earth have traditionally been assumed to only vary radially (e.g. (Argus et al., 2014, Peltier, 2004, Whitehouse et al., 2012a). In reality, there is seismic evidence for a laterally heterogeneous structure in the lithosphere and the mantle beneath Antarctica.

1.2.1. Lateral Contrast in Shear Wave Velocities

Seismic studies indicate that there is a sharp lateral contrast in shear wave velocity anomalies between East and West Antarctica, especially West of the Ross Sea Embayment (RSE) (Danesi and Morelli, 2001, Ritzwoller et al., 2001). A shear wave velocity model of An et al. (2015) showing this transition at 100 km depth is displayed in Figure 1.2.

Figure 1.2: Shear wave velocities in Antarctica at a depth of 100 km (An et al., 2015).

Figure 1.2 shows shear wave velocities of up to 4.6 km/s underneath East Antarctica, and velocities of 4.2 km/s and lower in West Antarctica. Temperature can be derived from seismic tomography maps using physical scaling laws (Ivins and Sammis, 1995). Shear wave velocities are first converted to density perturbations, and the density perturbations are converted to temperature variation us-
ing a thermal expansion coefficient. Low shear wave velocities indicate relatively high temperatures and vice versa. However, when lateral heterogeneity in mantle petrology is not considered in the conversion of shear wave velocities to temperature, first-order uncertainties are introduced (Trampert and Van Der Hilst, 2005). Temperature can be related to absolute viscosity values through flow laws for olivine following the methodology of van der Wal (2009) detailed in Section 2.2.4. In this approach it is assumed that olivine is the main mantle material in the UM. The flow laws of olivine introduce additional parameters such as grain size and water content, of which the distribution in the UM is not well known.

Nield et al. (2014) showed that GPS uplift rates in the Antarctic Peninsula (AP) require a viscoelastic response of the Earth to recent changes in ice mass. An average UM viscosity was found of lower than $2 \cdot 10^{18}$ Pa s. Also for the Amundsen Sea Sector evidence for a low mantle viscosity was found (Barletta et al., 2017). Thus, there is evidence for low viscosity beneath West Antarctica, at least in these two regions. However, comprehensive GIA studies for the whole of Antarctica have found preferred viscosities in the UM of $0.5 \cdot 10^{21}$ Pa s (Argus et al., 2014) and $10^{21}$ Pa s (Whitehouse et al., 2012a). Therefore, the question arises if there is a transition from low mantle viscosity beneath West Antarctica to higher mantle viscosity elsewhere. Due to the uncertainties associated to the derivation of a 3D viscosity distribution from shear wave velocities, it is uncertain if the lateral contrast in shear wave velocities across the Transantarctic Mountains (TAM) corresponds with a transition from low mantle viscosity beneath West Antarctica to high mantle viscosity beneath East Antarctica, and what the magnitude of such a transition would be.

### 1.2.2. Effect on Horizontal Velocities

Only few studies investigated the effect of a 3D rheology on horizontal velocities in Antarctica. Kaufmann et al. (2005) investigated the influence of lateral variations in mantle viscosity with a 3D flat-Earth FEM model. A viscosity distribution was derived from the seismic tomography model S20a (Ekström and Dziewonski, 1998). The results of the study showed that the spatial pattern of the predicted uplift rates in Antarctica was not significantly affected by lateral heterogeneities. However, a strong effect on both the magnitude and direction of horizontal deformation rates was discovered. The resulting direction of the horizontal velocities correlated spatially with the viscosity decrease from East to West Antarctica. Wang and Wu (2006) also found a horizontal flow pattern from East to West Antarctica with a spherical Earth model (see Figure 1.3). However, Wang and Wu (2006) used the ICE-4G deglaciation history (Peltier, 1994), which is biased to a preferred 1D rheology. It was also shown that predicted horizontal flow patterns for Antarctica can be completely different when different 3D rheology models are used (King et al., 2010).
Figure 1.3: Predicted horizontal velocities in Antarctica for the ICE-4G deglaciation history with (a) a 1D Earth model, and (b) a 3D Earth model, adapted from (Wang and Wu, 2006).

The effect of lateral viscosity variations in the mantle on GIA-induced horizontal motion has been explored in several other regions as well. Wu (2005) found that lateral variations in rheology based on the S20a model affect horizontal motion for Laurentia/North America and weaken the divergent horizontal motion from the ice load found for a homogeneous reference Earth model. They also found that the extent to which that happens depends on the ice model and the distance of the ice load center with respect to the lateral variation, and that lateral variations in lithospheric thickness affect horizontal motion less than lateral variations in mantle viscosity. A significantly weakening effect on outward motion close to the ice load for North America was also found by (Latychev et al., 2005).

For Fennoscandia, Steffen et al. (2006) found with a suite of 3D Earth models that expected divergent motion for a 1D Earth model was overprinted when lateral variations were included. A horizontal motion from the Norwegian Coast towards the old Baltic Shield was found, which corresponds with a flow from low UM viscosity to high UM viscosity (in contrast with the predicted horizontal flow from East to West Antarctica found by Kaufmann et al. (2005) and Wang and Wu (2006)). The horizontal motion was found to be very sensitive to a reversed viscosity contrast in the transition zone (TZ), which is the layer between the UM and the lower mantle (LM). With a finite-volume approach, Whitehouse et al. (2006) also found for Fennoscandia that lateral variations in rheology result in an Eastward perturbation of the horizontal flow. However, they found that this is mainly due to lateral variations in the lithosphere and the UM. More recently, it was presented in a poster by Steffen et al. (2016) that preliminary results show that in order to explain distinct horizontal flow patterns in GNSS data for Fennoscandia, lateral rheology variations are required.

It follows from the above studies that both the magnitude and the direction of predicted horizontal velocities are significantly affected when 3D variations in rheology are adopted. However, the predictions of how the horizontal velocities are affected by viscosity differ greatly for different studies and across different regions. It was also shown that the predicted horizontal flow pattern for, among other regions, Antarctica, depends strongly on the particular 3D rheology model that is used. Thus, the exact effect of viscosity and lateral variations of viscosity of the mantle on horizontal velocities remains unclear.
1.3. Antarctic Ice Sheet and Observations of GIA

Since the LGM ended, most of the ice sheets that covered the Earth melted completely. However, major ice sheets in Greenland and Antarctica persist. The Antarctic Ice Sheet (AIS) still covers almost the whole of the Antarctic surface. Major ice domes can be found at the Ross and Ronne Ice Shelves, the locations of which are shown in Figure 1.4 (taken from Whitehouse et al., 2012b). A deglaciation history for Antarctica is difficult to constrain, as only few historic records of RSL in Antarctica are available (see the blue circles in Figure 1.4). Because of the scarcity of constraints, Antarctic ice histories are often tuned in conjunction with Earth models, meaning that the fit of model predictions with observations is optimized by tuning the Earth and ice model simultaneously. As a result, the deglaciation history will depend on the assumptions made for the Earth model. Combining that ice history with any arbitrary Earth model that has different assumptions might result in erroneous predictions of GIA (King et al., 2010, Steffen et al., 2006). The three ice models that are currently used most often for Antarctica are W12(a) (Whitehouse et al., 2012a,b), ICE-6G_C (Argus et al., 2014) and IJ05_R2 (Ivins et al., 2013). W12a is an extension of W12 that includes ice thickness variability in the late Holocene in the AP (see Figure 1.4).

Figure 1.4: Map for Antarctica adapted from Whitehouse et al., 2012b), Figure 1. The projection of Antarctica shows the major regions of interest in this study. The largest ice domes can be found at the Ronne and Ross Ice Shelf. In the figure GPS sites are shown in red, RSL sites in blue, and ice extent sites in green.

Major discrepancies exist between W12, ICE-6G_C and IJ05_R2 regarding the total contribution of the ice that melted in Antarctica to eustatic sea-level rise, the timing of deglaciation and the exact location of major ice loading and unloading centres. The differences are a result of the use of different datasets and methodologies to derive the deglaciation histories, and form the most important source of uncertainty in GIA models. For a more elaborate discussion and comparison of these ice models the reader is referred to Argus et al., 2014 and the supplementary information of Ivins et al., 2013). Except for the extension of W12 (W12a), the aforementioned deglaciation histories do
not take into account changes of the AIS in the late Holocene. Especially in low viscosity regions in Antarctica like the AP, recent variability in the surface mass balance can have a significant effect on GIA-induced deformation rates (e.g. (Nield et al., 2014)).

In order to constrain GIA in Antarctica, primarily RSL and vertical GPS rates are used as observations (e.g. (Ivins et al., 2013, Whitehouse et al., 2012a). Just like the distribution of historic records of RSL, also the distribution of GPS sites is sparse due to the presence of the AIS. This is not only due to the harsh conditions that the continent imposes, but also because of the fact that GPS sites must be bedrock based in order to measure the deformation of the surface due to GIA directly. A global network of GPS data observing polar regions is the Polar Earth Observing Network (POLENET). GPS sites of the Antarctic Network (ANET) part of the POLENET and of collaborating networks can be seen in Figure 1.5. Additional GPS sites in East Antarctica can be seen as red circles in Figure 1.4. Horizontal GPS observations are often ignored because of their presumed sensitivity to lateral viscosity variations, which were explained to be neglected in most GIA models. Only Argus et al. (2014) published a comparison of their predictions to horizontal GPS observations in Antarctica, but used a 1D Earth model. Because of the scarcity of GIA observations in Antarctica, an important opportunity is seen to include 3D rheologies in GIA models and compare predictions with horizontal GPS observations as well. Horizontal GPS observations also contain less noise than vertical ones, but are more sensitive to LM viscosity (Mitrovica et al., 1994).
1.4. ANET GPS AT THE ROSS SEA EMBAYMENT

A remarkable feature of the horizontal GPS observations in Antarctica can be found West of the RSE. There, along the lateral contrast in shear wave velocities between West and East Antarctica, horizontal GPS velocities from ANET point towards the ice dome at the Ross Sea (Wilson et al., 2015). A recent GPS solution of ANET observations (Wilson et al., 2017-in prep) was provided by Stephanie Konfal of Ohio State University, and the resulting horizontal velocities surrounding the RSE are shown in Figure 1.6. A short description of this GPS solution follows, based on personal communication with Stephanie Konfal. The observations of the GPS solution have continuously been gathered for more than a decade. The solution was obtained through processing of the data with GPS analysis software called GAMIT/GLOBK (Herring et al., 2010). Horizontal velocities relative to estimated plate motion were obtained following the approach of Bevis et al. (2013). Plate motion was estimated and corrected for using 9 stable sites at the East Antarctic coast, in a manner that is independent on the treatment of vertical velocities. Note that based on the approach to obtain the GPS solution and the treatment of aspects like processing parameters, site equipment changes and many others, there can be variability between the final velocities in different GPS solutions for the same sites.

Figure 1.6: ANET/POLENET horizontal GPS observations surrounding the RSE. 1 sigma uncertainty ellipses are appended at the top of the velocity vectors. TAM = Transantarctic Mountains, WSB = Wilkes Subglacial Basin, Ross I.S. = Ross Ice Shelf.

As can be seen in the figure, GPS sites are mainly concentrated at the region West of the RSE. Along the Transantarctic Mountains (TAM) the velocities point towards the ice load rather than away from it. However, based on previous studies with 1D rheologies, a horizontal flow radially outward from the ice centre (divergent) is expected (James and Lambert, 1993, Mitrovica et al., 1994, Peltier, 1998). It is unclear whether the direction of the measurements is caused by a potential lateral viscosity transition West of the RSE, since the dependency of the direction of horizontal velocities on mantle viscosity is not well understood (as was discussed in Section 1.2).
1.5. **RESEARCH QUESTIONS**

In this study we focus on the effect of lateral rheology variations underneath Antarctica on horizontal velocities. It was explained that there is evidence for regions with low UM viscosity in West Antarctica. Additionally, seismic shear wave velocities show a lateral contrast between West and East Antarctica that might indicate a lateral transition in viscosity. As was discussed in the introduction, the direction of horizontal GPS velocities West of the RSE is towards the principle ice load rather than away from it, which is a feature that has not been explained yet. It is unclear if it can be explained by a lateral transition in viscosity, because the dependency of the direction of horizontal velocities on mantle viscosity is poorly understood. Therefore, the objectives in this study are twofold. First of all, we aim to clarify the dependency of horizontal velocities induced by GIA on mantle viscosity. Secondly, we investigate if we can use the horizontal GPS measurements of ANET/POLENET surrounding the RSE to constrain the presumed lateral transition in viscosity between West and East Antarctica. To guide the research in this thesis the following research questions are formulated, which will be answered in the remainder of this report:

I. To what extent can horizontal GPS observations at the Ross Sea Embayment constrain the lateral transition in mantle viscosity between West and East Antarctica in forward models of Glacial Isostatic Adjustment?

   (a) How does the direction of GIA-induced horizontal deformation rates depend on mantle viscosity?

   (b) What are the main differences and similarities between the predictions of horizontal deformation rates from GIA models with and without a lateral variation in mantle viscosity West of the Ross Sea Embayment?

   (c) What viscosity values are preferable at both sides of the viscosity transition between West and East Antarctica when comparing the predictions of horizontal deformation rates with GPS observations at the Ross Sea Embayment?

   (d) What is the effect of variation of lower mantle viscosity and lithospheric thickness on the fit of the predictions of the preferred Earth model with the horizontal GPS measurements at the Ross Sea Embayment?

1.6. **REPORT OUTLINE**

In the previous sections, the research objective was introduced. To answer the research questions that were posed, the main part of this report is written as a draft journal paper. Chapter 3 contains the article, in which the research done in this thesis is introduced once more, the results and findings are presented, and a short account is given of the implications of the results. Additionally, in the methods section the methodology used throughout this thesis is explained. The maximum number of words in the final version of the article (excluding methods) is fixed to 3000, and there is a maximum of 6 display items. In Chapter 4, the supplementary information to this article is presented. Theoretical concepts that are not explained in detail in the article or supplementary information, are included in this report in Chapter 2. Finally, in Chapter 5 several conclusions are drawn based on the results of this thesis. The research questions posed in Chapter 1 will be answered, and recommendations are given for further study. The report ends with a section in which the results of this thesis and their implications are expanded to a wider perspective.
As was mentioned, for this study a coupled FEM-Laplace model of GIA is used (van der Wal et al., 2015). In this chapter, the theory behind this model is discussed in Section 2.1. Next, the two main inputs to the model will be discussed. In Section 2.2 the structure of the Earth and how to model it will be explained. Next, in Section 2.3 the deglaciation history W12 that is used in this model will be briefly discussed, and it will be shown how to account for ocean loading. The theoretical concepts discussed in this chapter support the methodology of the research in this thesis that is explained in Chapter 3.

2.1. COUPLED FEM-LAPLACE GIA MODEL

FEM is a numerical technique that was traditionally a branch of solid mechanics but is now also used to solve problems of physics. Instead of analytically solving such complicated problems directly, with FEM, a structure is subdivided in a finite number of elements that are connected at their common nodes. The simpler algebraic equations that can be used to model these finite elements are then gathered in a system of (often partial differential) equations. Solving that system of equations under the prescription of a set of boundary conditions enables to approximate the solution to the original problem. For GIA, (visco)elastic equations are chosen to model the deformation of the solid Earth. A clear advantage of FEM models is that they are well suited to problems with a complicated geometry and large variations in properties (Wu, 2004), such as the lateral variations in mantle beneath Antarctica. In this study the FEM package Abaqus version 6.14 was used. To model the self-gravitation of the Earth, the FEM model is coupled to the Laplace equation (Poisson’s equation when accounting for internal buoyancy). In the sections below, the theory behind this will be clarified following the framework presented by Wu (2004).

2.1.1. EQUATION OF MOTION

For a body that is in equilibrium, the conservation of linear momentum dictates that the stress acting on the surface of an infinitesimal element of that body must be equal to the body forces acting on the element. For long time scales it is assumed that the intertial forces in the equation have disappeared (Sabadini and Vermeersen, 2004). Thus, when in engineering applications in Abaqus and other FEM programs no body forces are considered, the following equation of motion follows (Equation 2.1):

\[ \nabla \cdot \sigma_{ij} = 0 \]  \hspace{1cm} (2.1)

where \( \nabla \cdot \sigma_{ij} \) is the divergence of the stress tensor. For geophysical problems, Equation 2.1 is extended to the expression in Equation 2.2 (Wu, 2004, Zhong et al., 2003):
\[ \nabla \cdot \sigma_{ij} - \nabla (u \cdot \rho_0 g_0 \hat{r}) - \rho_1 g_0 \hat{r} - \rho_0 \nabla \phi_1 = 0 \] (2.2)

in which \( \rho, g \) and \( \phi \) are the density, gravitation and gravitational potential perturbation respectively, \( \hat{r} \) and \( \hat{u} \) are a unit vector in radial direction and the displacement vector, and the subscripts 0 and 1 refer to a hydrostatic or perturbed state. The second term on the left-hand side refers to the advection of hydrostatic pre-stress. Put differently, this is the restoring force of isostasy, owing to the fact that pressure has to increase when there is elastic displacement in the opposite direction of the pressure gradient in the Earth (Sabadini and Vermeersen, 2004). The third term on the left accounts for the effect of internal buoyancy, and the fourth term for the effect of self-gravitation. The perturbation of the gravitational potential is a result of the deformation of the Earth due to GIA and is described by Poisson’s equation:

\[ \nabla^2 \phi_1 = 4\pi G \rho_1 \] (2.3)

where \( G \) is the gravitational constant. The gravitational potential is also perturbed by the redistribution of the ocean due to ice melt, which will be discussed in Section 2.3.2. If the material is assumed to be incompressible and thus the stress term due to internal buoyancy is neglected and the so-called Eulerian density perturbation \( \rho_1 \) equals zero, Poisson’s equation reduces to the Laplace equation:

\[ \nabla^2 \phi_1 = 0 \] (2.4)

 Assuming no incompressibility, possible numerical instabilities may arise related to the representation of internal buoyancy due to the perturbed density when elements are deformed. To avoid this only material compressibility can be included by changing the Poisson’s ratio and thus the Young’s modulus. However, the separation of material compressibility cannot be done physically (Klemann et al., 2003), as in reality dilatating elements would introduce density perturbations. The possibly arising numerical errors when including compressibility impose a limitation on the model. When studying small wavelength effects (lower than 8000 km), the numerical instabilities in the model might be circumvented by adapting the elastic rigidity of an incompressible model to approximate the flexural rigidity of a compressible model (Tanaka et al., 2011).

2.1.2. SPHERICAL, SELF-GRAVITATING AND INCOMPRESSIBLE EARTH

In order to bring the form of Equation 2.2 back to that of Equation 2.1 normally used in commercial FEM packages, the stress tensor is rewritten to include the additional stress contributions. As a consequence, boundary conditions must be adapted as well, and the stress output of Abaqus has to be converted back before interpretation. As will become clear in the Methods section of Chapter 3, in this study a spherical, self-gravitating and incompressible Earth is assumed. Neglecting the internal buoyancy, the stress tensor can be redefined as in Equation 2.5:

\[ \nabla \cdot s_{ij} = \nabla \cdot \sigma_{ij} - \nabla (u \cdot \rho_0 g_0 \hat{r}) - \rho_0 \nabla \phi_1 \] (2.5)

The stress tensor \( \sigma_{ij} \) consists of two parts: hydrostatic pre-stress and the acquired non-hydrostatic stress (Sabadini and Vermeersen, 2004). The latter can be related to strain following the constitutive elastic and viscoelastic relations presented in Section 2.2.2. The pressure that is applied downward on the surface of the Earth is composed of the ice load (i.e. surface mass density times the gravitational acceleration) and the load due to the potential perturbation \( \rho_0 \phi_1 \). The term \( u \cdot \rho_0 g_0 \hat{r} \) is accounted for through a so-called Winkler foundation at the surface, i.e. an elastic spring with a spring constant that is equal to the difference in density times the hydrostatic gravity at the surface. At internal boundaries the potential load \( \nabla \phi_1 \) and Winkler foundations are also applied. For a detailed overview of the boundary conditions the reader is referred to (Wu, 2004), but for the purpose...
of clarifying the tests done in this study that will be explained later on, it suffices to understand how the different loads are applied. Having described the set of boundary conditions for this particular model, now the approach to solve the problem will be explained.

2.1.3. Iterative Solution
When assuming an incompressible, self-gravitating Earth, the perturbation of the gravitational potential in Equation 2.5 can be computed by solving the Laplace equation (Equation 2.4). This is done with the coefficients of the spherical harmonic expansion of the deformations obtained from the GIA model. An iterative procedure is applied to couple FEM to the Laplace equation. First, FEM is used to obtain the deformations for the case in which the Earth is not self-gravitating. Then, the deformations are expanded and their spherical harmonic coefficients are used to compute the perturbation of the gravitational potential. Next, with the new gravitational potential the boundary conditions at the surface are updated, effectively resulting in a new load to the FEM model. Finally, the previous steps are repeated for the self-gravitating case, until convergence criteria are met. Usually an acceptable level of convergence is found after 4 or 5 iterations (Wu, 2004), depending on the specific application and parameters choices that were made.

2.2. Modelling the Solid Earth
In order to compute the deformation of the Earth with the coupled FEM-Laplace model of GIA, a parameterization of the structure of the solid Earth is required as an input. In most Earth models, the Earth is assumed to be built up of a discrete number of spherical layers of elements having different material properties (e.g. density, elastic parameters and viscosity). In the following section, the radial structure of the Earth will be briefly discussed. Additionally, the theory behind the rheology of the Earth, i.e. the elastic and viscoelastic relations between stress and strain that govern the response of the Earth to loading in the FEM model, will be introduced. It will be explained how the density and elasticity of the different layers of the Earth can be obtained, and how a 3D viscosity distribution for the mantle can be derived from seismic measurements.

2.2.1. Radial Structure of the Earth
The Earth can be roughly divided into a distinct number of spherical shells with different properties and composition (see Figure 2.1). The outer layer of the Earth is the crust, which is a hard and rigid layer. Underneath oceanic basins, the crust is thin. Oceanic crust forms at mid-oceanic ridges, and is rich of iron and magnesium. Thus, oceanic crust is denser than continental crust. The continental parts of the crust are thicker and consist of felsic rocks like granite. The layer below the crust is the mantle. The transition in chemical composition between the crust and the mantle (Schubert et al., 2001) is called the Mohorovičić discontinuity (Moho). The depth of the Moho can be derived from reflection seismology. On average, the depth of the Moho was found to be 11 km underneath oceans and 35 km underneath continents (Dziewonski and Anderson, 1981). The lithosphere consists of the crust and the top part of the mantle. The part of the mantle that is not lithospheric is the asthenosphere. The transition of the lithosphere into the mantle is marked by a change in mechanical properties (Schubert et al., 2001). Throughout the time-span of a glacial cycle the mantle deforms viscoelastically, whereas the lithosphere deforms not viscously, but elastically. The definition of the lithosphere based on its mechanical properties is also used in GIA studies.

The upper part of the mantle up to a depth of 400 km consists mainly of the mineral olivine (Turcotte and Schubert, 2002). A recent global GIA study (ICE-6G_C (VM5a)) inferred an average viscosity of \(0.5 \cdot 10^{21}\) Pa s for the UM of the whole Earth (Peltier et al., 2015). A small number of xenolith findings in Antarctica show olivine grain sizes of sub-mm to mm size and the presence of hydrous material (see the Supplementary Information of van der Wal et al. (2015) for an overview),
indicating relatively low UM viscosity. However, these findings are spatially poorly distributed. At around 670 km, seismic velocity discontinuities indicate a transition between the upper and lower mantle. The seismic discontinuity that indicates this transition can be explained by mantle silicates transforming to perovskites, but compositional variations might also play a role (Schubert et al., 2001). Although viscosity decreases with increasing temperature, it increases with pressure. Thus, the LM is thought to have a viscosity of 1 to 2 orders of magnitude higher than the UM.

2.2.2. ELASTICITY AND VISCOELASTICITY

Over timescales associated with GIA, the mantle of the Earth deforms viscoelastically, and the lithosphere elastically. Viscoelasticity can be represented conceptually by various combinations of springs and dashpots. The most common constitutive model for viscoelasticity is the Maxwell model, in which a spring and a dashpot are connected in series. The impulse response of the Earth based on Maxwell viscoelasticity was first put forward by Peltier (1974) and is chosen as the representative model of the viscoelastic behaviour of the elements in the FEM model that is used for this study.

In the Maxwell model, the total strain equals the sum of the strain of the spring and the dashpot, and the stress in both elements must be the same. For elastically isotropic elements, the stress-strain relationship (Hooke's law) is given by (Malvern, 1969):

\[ \sigma_{ij} = \lambda \epsilon_{kk} \delta_{ij} + 2\mu \epsilon_{ij} \]  

(2.6)
2.2. M\textsc{ODELLING THE S\textsc{OLID EARTH}}

where $\sigma_{ij}$ is the stress tensor, $\epsilon_{ij}$ the strain tensor, $\delta_{ij}$ the Kronecker delta, and $\lambda$ and $\mu$ are the so-called Lamé parameters, which depend on the Young’s modulus $E$ and Poisson’s ratio $\nu$. For viscous materials, stress can be written as a function of strain rate as follows (Malvern, 1969):

$$\sigma_{ij} = -p\delta_{ij} + \dot{\lambda} \epsilon_{kk} \delta_{ij} + 2\eta \dot{\epsilon}_{ij}$$  (2.7)

in which $\dot{\epsilon}$ is the strain rate, $p$ is the thermodynamic pressure and $\dot{\lambda}$ and $\eta$ are independent parameters which describe the viscosity of the material (the bulk viscosity $\kappa = \dot{\lambda} + \frac{2}{3} \eta$). Combining the above relations for elastic and viscous behavior for the Maxwell model, a linear stress-strain rate relation for Maxwell viscoelasticity follows that is often used in GIA studies (Peltier, 1974):

$$\dot{\sigma}_{ij} + \frac{\mu}{\eta} (\sigma_{ij} - \frac{1}{3} \sigma_{kk} \delta_{ij}) = 2\mu \dot{\epsilon}_{ij} + \lambda \dot{\epsilon}_{kk} \delta_{ij}$$  (2.8)

in which again the Lamé parameters and the viscosity can be recognized.

The Maxwell model of viscoelasticity represents a \textit{linear} relation between stress and strain rate, implying that diffusion creep is the dominant creep mechanism in the mantle. Creep refers to the slow, plastic deformation of mantle rocks under the influence of stress. However, based on laboratory experiments (Goetze and Kohlstedt, 1973) it is thought that also dislocation creep plays a role in the deformation of mantle material, which is associated with \textit{non-linear} behavior. Therefore, the rheology of the Earth could also be partially non-linear. Such a power-law rheology can be formulated as follows (Ranalli, 1995):

$$\dot{\epsilon} = A\sigma^n$$  (2.9)

in which $A$ is a flow law parameter which depends among others on activation energy, pressure, temperature and material properties, and $n$ is the stress exponent (1 for linear rheology, higher for a power-law rheology). When a rheology is modelled to consist partially of linear and partially of non-linear behaviour (e.g. (Gasperini et al., 1992)), it is called a composite rheology.

2.2.3. DENSITY AND ELASTIC PARAMETERS

For the viscoelastic stress-strain rate relation introduced in Equation 2.8, the elastic parameters and density need to be estimated for each defined layer of the Earth model. The density and elastic parameters for each Earth layer are derived by volume-averaging of the values in the Preliminary Reference Earth Model (PREM) (Dziewonski and Anderson, 1981). PREM is a widely used model for the radial distribution of the elastic quality factor $Q$, elastic properties and density of the Earth, which was derived from astronomical geodetic data and observations of seismic velocities.

2.2.4. 3D VISCOSITY THROUGH CREEP FLOW LAWS

The viscosity of the mantle $\eta$ needs to be defined for each element of the FEM model. In a 1D Earth model, the viscosity is varied radially for a discrete number of layers. In reality viscosity also varies laterally. Temperature data in the upper mantle can be used to derive a both laterally and radially varying viscosity distribution from, when constitutive flow laws for the creep mechanisms in the mantle are used. Such an approach was presented by van der Wal (2009) and used in a number of studies on 3D rheology (Barnhoorn et al., 2011, van der Wal et al., 2013, 2015). This methodology will be briefly explained here.

As was mentioned earlier, two major creep mechanisms characterizing mantle rock rheology are diffusion creep and dislocation creep (Peltier, 1974). Creep for olivine, which was mentioned to be the main mantle material up to 400 km depth, is characterized by the following flow law (Hirth and Kohlstedt, 2003):
\begin{equation}
\dot{\epsilon} = A \sigma^n d^{-p} f H_2O^r \exp(\alpha \phi) \exp(-\frac{E + PV}{RT})
\end{equation}  \tag{2.10}

where $\dot{\epsilon}$ is the strain rate, $A$ a constant, $\sigma^n$ the stress to a certain power $n$, $d$ the grain size, $f H_2O$ the water fugacity, $r$ the fugacity exponent, $p$ the grain size exponent, $\alpha$ a constant, $\phi$ the melt fraction, $P$ the pressure, $E$ the activation energy, $R$ the gas constant, $V$ the activation volume, and $T$ the absolute temperature. The values of $n$ and $p$ can be derived from laboratory experiments. For diffusion creep, the flow law is linear and is characterized by $n = 1.0 \pm 0.1$, and $p = 3.0 \pm 0.5$. For dislocation creep, the stress exponent $n$ is around 3.5, and $p = 0$ (Hirth and Kohlstedt, 2003), resulting in a power-law rheology which is independent of grain size.

The strain in the elements of the FEM model can be computed as:

\begin{equation}
\epsilon = B_{\text{diff}} q t + B_{\text{disl}} q^n t
\end{equation}  \tag{2.11}

where the creep parameters $B_{\text{diff}}$ and $B_{\text{disl}}$ are derived from Equation 2.10, $q = \sqrt{\frac{3}{2} \sigma_{ij}' \sigma_{ij}'}$ is the Von Mises stress, and $\sigma_{ij}'$ is an element of the deviatoric stress tensor (the part of the stress that is not hydrostatic, see Section 2.1.2). The strain rates for dislocation and diffusion creep can be formulated as a function of stress as follows (van der Wal, 2009):

\begin{equation}
\dot{\epsilon}_{ij, \text{diff}} = \frac{3}{2} B_{\text{diff}} \sigma_{ij}' \quad \dot{\epsilon}_{ij, \text{disl}} = \frac{3}{2} B_{\text{disl}} q^{n-1} \sigma_{ij}'
\end{equation}  \tag{2.12}

Adding the two terms above to a combined strain rate, the following expression is obtained:

\begin{equation}
\dot{\epsilon}_{ij} = \frac{3}{2} B_{\text{diff}} \sigma_{ij}' + \frac{3}{2} B_{\text{disl}} q^{n-1} \sigma_{ij}'
\end{equation}  \tag{2.13}

When a 3D temperature distribution is used for the temperature $T$ in Equation 2.10, a 3D composite rheology can be modelled if estimates are available of the other parameters in the flow law. The resulting rheology can be represented by an effective viscosity. Introducing the effective viscosity as (Ranalli, 1995),

\begin{equation}
\eta_{\text{eff}} = \frac{\sigma_{ij}'}{2 \dot{\epsilon}_{ij}}
\end{equation}  \tag{2.14}

the effective viscosity for a composite rheology can be written as:

\begin{equation}
\eta_{\text{eff}} = \frac{1}{3 B_{\text{diff}} + 3 B_{\text{disl}} q^{n-1}}
\end{equation}  \tag{2.15}

The effective viscosity can be used to compare with other viscosity distributions. Using it to compute strain rates is equivalent to using Equation 2.12. The resulting 3D rheology is only valid up to 400 km depth, because only until there olivine is thought to be the main material governing the deformation of the mantle. Note that since the Von Mises stress appears in the flow law for olivine, the 3D viscosity obtained through the methodology of van der Wal (2009) depends on the deglaciation history that is used.
2.3. ICE AND OCEAN LOADING

The deglaciation history used in the GIA model determines the loading and unloading of the surface of the Earth, and is an important input to the model. The changes in RSL that accompany the loading and unloading of the surface due to changes in ice thickness can be taken into account in the model as well.

2.3.1. DEGLACIATION HISTORY

Constraints on the ice thickness can be obtained by means of glaciological, geomorphological and geological observations. Ice volume is primarily constrained by records of RSL (Peltier, 1998) and by measurements of oxygen isotopes (Shackleton, 1987). Glacial deposits such as moraines can be used to constrain the extent of ice sheets at different times. Additionally, in Antarctica the ice history is constrained by the configuration of present-day ice sheets. As mentioned earlier, the ice history can also be tuned based on predictions of GIA models. As was referred to in Section 1.3, the three latest deglaciation histories for Antarctica that are commonly used in the literature are W12(a) (Whitehouse et al., 2012a,b), IJ05_R2 (Ivins et al., 2013) and the Antarctic component of ICE-6G_C (Argus et al., 2014). Of these three models, only W12(a) is based on a physically-consistent numerical model. Moreover, W12(a) has not been tuned simultaneously with a 1D Earth rheology, in contrast with ICE-6G_C. Therefore, W12(a) was used for this study. The model by Whitehouse et al. (2012a) is developed using a shallow-ice 3D numerical code which is based on simple glacial physics. The extent and the thickness of the ice are described in a series of discrete time-steps since the LGM. The reconstruction is constrained by glaciological and geological data. Present-day bedrock topography is used to determine where ice is marine-grounded or floating. Different basal sliding laws are used for areas that are marine-grounded and areas that are not, as faster basal sliding is expected for ice that is underlain by marine sediments. For a more detailed account of how W12 was developed, the reader is referred to Whitehouse et al. (2012a). The ice height in the W12 model at 20 ka before present (BP) relative to the ice height at 2 ka BP, which is interpolated on our 2° by 2° grid, can be seen in Supplementary Figure 3 in Chapter 4.

2.3.2. SEA-LEVEL EQUATION

Conservation of mass dictates that the ice unloading since the LGM results in a corresponding increase in ocean volume. Water is distributed according to a specific equipotential surface that is referred to as the geoid. However, melting ice sheets and the rebound process change the shape of the geoid because they change the gravitational attraction. Thus, the change in relative sea level consists of an eustatic part because of the spatially-uniform redistribution of water when freed from or locked up in ice, and a non-uniform part because of perturbations to the geoid shape. Farrell and Clark (1976) formulated the so-called sea-level equation, which describes how water is distributed over the surface of the Earth:

$$\Delta S(\theta, \psi, t) = [\Delta G(\theta, \psi, t) - \Delta R(\theta, \psi, t)]C(\theta, \psi)$$

(2.16)

In Equation 2.16, the change in sea-level height $\Delta S(\theta, \psi, t)$ is equal to the difference between the change in geoid height $\Delta G(\theta, \psi, t)$ and solid surface height $\Delta R(\theta, \psi, t)$ as a function of position and time. The longitude $\theta$ and colatitude $\psi$ indicate the dependence on position. Since only changes in oceanic heights $S$ are considered, a time-invariant ocean function $C(\theta, \psi)$ is introduced that assumes a value of 1 in case bedrock topography is negative, and a value of 0 elsewhere (Farrell and Clark, 1976).

The change in the height of the surface $R$ is a result of the deformation of the Earth due to GIA. The change in geoid height $G$ consists of a spatially varying part and a spatially uniform part. The spatially varying part $\Delta G_{SV}(\theta, \psi, t)$ represents the change in geoid shape because of perturbations
to the gravitational potential. These include changes in direct gravitational attraction (e.g. by ice load removal) and indirectly through deformation of the solid Earth. The centrifugal potential of the Earth can be affected by the deformation of the Earth and changes in the surface mass load (rotational feedback), however this is not considered in this study. The spatially-uniform change of the geoid height $\Delta G_{UV}(\theta, \psi, t)$ is caused by an increase in the volume of ocean water because of the addition of meltwater. This relation is governed by the conservation of mass (the first term in Equation 2.18). Additionally, $\Delta G_{UV}$ includes a term that represents a uniform change in ocean height (the second term in Equation 2.18). This term is the difference between the spatially-varying capacity change of ocean basins and the spatially-varying addition of meltwater to this new basin configuration, calculated as an average over all ocean areas. Thus, Equation 2.16 can be expanded (Farrell and Clark, 1976):

\[
\Delta S(\theta, \psi, t) = \left[ \Delta G_{SV}(\theta, \psi, t) - \Delta R(\theta, \psi, t) + c(t) \right] C(\theta, \psi)
\] (2.17)

where $c(t)$ contains the eustatic sea-level change which is formulated as:

\[
c(t) = -\frac{\rho_I}{\rho_W} \frac{V_I(t)}{A_O} - \frac{\langle \Delta G_{SV}(\theta, \psi, t) - \Delta R(\theta, \psi, t) \rangle_O}{A_O}
\] (2.18)

In Equation 2.18, $\rho_I$ and $\rho_W$ are the ice and water densities, $V_I$ is the volume loss of the ice, $A_O$ is the ocean basin surface and $\langle \rangle_O$ indicates the integration over the ocean basins.

The sea-level equation in Equation 2.16 holds for a non-rotating Earth. In reality, there is a rotational feedback. The changing loads and resulting GIA perturb the rotation vector of the Earth (Milne and Mitrovica, 1996). The perturbed centrifugal potential in turn deforms the solid Earth and the sea-level (Paulson et al., 2005). Additionally, the time-invariant ocean function $C$ that was introduced in Equation 2.16 is in reality changing over time, due to the migration of shorelines (Peltier, 1994). Within two successive times $t_i$ and $t_j$ the shoreline can move land inwards because of the addition of meltwater. When the bedrock topography is very steep, the effect of shoreline migration is less significant. Another extension of the ocean function can be derived for marine-grounded ice. When marine-grounded ice ablates, the effect of meltwater influx can significantly influence GIA, especially in areas with large ice shelves like the RSE. Marine-grounded ice can be incorporated by extending the ocean function following Mitrovica and Milne (2003). The ocean function is then formulated as $C^* = C * \beta$, where $\beta(\theta, \psi, t)$ assumes a value of 0 over grounded ice and a value of 1 in the absence of grounded ice. The sea-level equation is solved iteratively in the 3D GIA model used here. This part of the model was programmed by Hansheng Wang and was validated in (Wang et al., 2006).
3

ARTICLE
Constraining Glacial Isostatic Adjustment with Horizontal GPS Velocities in Antarctica

T.b.d.¹*

Abstract

Glacial isostatic adjustment is the viscoelastic response of the Earth to ice and ocean loads. In forward models of glacial isostatic adjustment, mantle viscosity is often assumed to be laterally homogeneous. However, a lateral transition in shear wave velocities suggests a sharp transition in viscosity between West and East Antarctica. Along this transition, horizontal GPS velocities West of the Ross Sea Embayment point towards the ice load rather than away from it. It is unclear why, as the dependency of horizontal velocities on viscosity is not well understood. Here, we clarify this dependency, and investigate with a 3D finite-element model if the horizontal GPS velocities can be used to constrain the viscosity transition. We found that horizontal velocities point away from the ice load for viscosities of \(10^{20}\) Pa s and lower, whereas for \(10^{21}\) Pa s and higher their direction is reversed. Our predictions show that the GPS measurements at the Ross Sea Embayment likely require a lateral viscosity transition. We found preferred viscosities in the upper mantle to lay between \(10^{18}\) and \(10^{19}\) Pa s in West Antarctica, and \(10^{21}\) and \(10^{22}\) Pa s in East Antarctica. Our results also demonstrate that horizontal GPS velocities can be used to constrain lateral variations in rheology. As more studies will start to use 3D Earth models, horizontal GPS velocities should be used as one of the primary constraints of glacial isostatic adjustment, since their direction can be reversed depending on mantle viscosity.

Introduction

Glacial Isostatic Adjustment (GIA) is the viscoelastic response of the solid Earth to ice and ocean loads on the surface since the Last Glacial Maximum (LGM). GIA can be predicted with geophysical forward models (Peltier, 1976). When comparing GIA predictions with observations, valuable information about deglaciation history and the Earth’s rheology can be inferred. In the majority of GIA models, mantle viscosity is assumed to vary radially only (1D) (e.g. Peltier, 2004, Argus et al., 2014) rather than radially and laterally (3D). However, recent GIA studies found low mantle viscosities beneath the Antarctic Peninsula (AP, \(< 2 \times 10^{18}\)) (Nield et al., 2014) and the Amundsen Sea Sector (Barletta et al, 2017), and seismic studies show a lateral contrast in shear wave velocities between West and East Antarctica (Ritzwoller et al., 2001, Danesi and Morelli, 2001). Shear wave velocities can be converted to temperature using physical scaling laws (Ivins & Sammis, 1995) and temperature can be related to viscosity, so a lateral contrast in shear wave velocities might indicate a transition from low viscosity beneath West Antarctica to high viscosity beneath East Antarctica. However, temperature cannot be derived from shear wave velocities with certainty without considering variations in chemical composition (Trampert & Van der Hilst, 2013).

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¹ T.b.d. *Corresponding Author, e-mail address: t.b.d.
For Antarctica, primarily historic records of relative sea-level (RSL) and vertical GPS velocities are used as observations in GIA studies (e.g. Whitehouse et al., 2012, Ivins et al., 2013). Since horizontal GPS velocities are sensitive to lateral viscosity variations (Kaufmann and Wu, 2005, Wang and Wu, 2006), these are often not included. Argus et al. (2014) compared their predictions to horizontal GPS velocities in Antarctica, but used a 1D Earth model. The few studies on the effect of a 3D rheology underneath Antarctica suggest that both the magnitude and the direction of horizontal velocities are significantly affected, and that a stiff mantle beneath East Antarctica induces a flow from East to West Antarctica (Kaufmann and Wu, 2005, Wang and Wu, 2006). However, these studies either did not include sphericity and self-gravitation (Kaufman and Wu, 2005), or used an ice model (ICE-4G (Peltier, 1994)) that was tuned simultaneously with a preferred 1D Earth model, which leads to inconsistent results. Moreover, it was shown that predicted horizontal flows are completely different when using different 3D rheology models (King et al., 2010), leaving the effect of viscosity unclear.

West of the Ross Sea Embayment (RSE), horizontal GPS observations from the Antarctic Network (ANET) part of the Polar Earth Observation Network (POLENET) point towards the ice dome at the RSE (Wilson et al., 2015), while a horizontal motion away from the ice is expected (James and Lambert, 1993, Mitrovica et al., 1994), with a magnitude depending on mantle viscosity (James and Morgan, 1990). It is unclear whether the inward direction of the horizontal GPS velocities is caused by the presumed lateral viscosity transition between West and East Antarctica, as the dependency of the direction of horizontal velocities on viscosity is not well understood. An alternative explanation could be additional ice loss at the Wilkes Subglacial Basin, but this cannot explain the GPS in all regions around the RSE (Konfal et al., 2015). Especially in Antarctica, where measurements are scarce, horizontal velocities could provide a unique opportunity to constrain lateral viscosity variations in 3D GIA models. In this study, we clarify the dependency of horizontal velocities on mantle viscosity and investigate to what extent the horizontal GPS observations West of the RSE can be used to constrain a lateral viscosity transition between West and East Antarctica.
Dependency of Horizontal Velocities on Mantle Viscosity

It is expected that during ice unloading, horizontal deformation rates on the Earth’s surface point away from the former ice centre, as the elastic lithosphere that was pulled towards the load centre moves outward. With an incompressible, spherical and self-gravitating 3D GIA model (Van der Wal et al., 2013, Van der Wal et al., 2015) (see Methods) we show that horizontal velocities do not always point away from former ice load, but that their direction strongly depends on mantle viscosity. We apply an axisymmetric ice cap (radius of 14°, see Methods) on the North Pole. The cap builds up linearly over 90 ka, melts during 10 ka, and has disappeared completely at 10 ka before present (BP). The homogeneous mantle viscosity is varied between $10^{19}$ and $10^{23}$ Pa s. The results in Figure 1 show that for viscosities of $10^{20}$ Pa s and lower, horizontal velocities point away from the former ice centre, while for viscosities of $10^{21}$ Pa s and higher velocities point inwards.

![Figure 1](dependency_of_horizontal_velocities_on_mantle_viscosity.png)

**Figure 1 | Dependency of horizontal velocities on mantle viscosities between $10^{19}$ and $10^{23}$ Pa s.** Predicted present-day horizontal deformation rates at the Earth’s surface induced by the loading and unloading of an axisymmetric ice cap. Positive velocities are directed outward from the former ice centre (located at the North Pole), and negative velocities are directed towards the ice centre.
Opposite Directions of Mantle and Lithosphere

Deformation rates through time at nodes in the lithosphere and upper mantle (UM) are shown in Figure 2 for a viscosity of $10^{20}$ Pa s and $10^{21}$ Pa s, i.e. the values between which the direction of our predictions reverses. See Methods for the computation of the deformation rates.

Figure 2 | Time-varying deformation rates at the nodes of layer boundaries in the lithosphere and upper mantle (UM). Predictions of the velocities at layer boundaries in the lithosphere (0 and 50 km depth) and UM (90, 200, 325 and 400 km depth) are shown for a homogeneous mantle viscosity of (a) $10^{20}$ Pa s and (b) $10^{21}$ Pa s, at three timesteps: 20, 15 and 5 ka BP. Note that the scale of the velocity arrows are different for each timestep.

For both viscosities the mantle and to a lesser extent the lithosphere move outwards during ice loading. During deglaciation, the mantle flows back towards the ice centre. The lithospheric velocities also point inward at 15 ka BP, suggesting that the inward mantle flow exerts a shear force on the lithosphere and pulls it inwards. This is confirmed with stress outputs of our FEM program. As time progresses, for a viscosity of $10^{20}$ Pa s (Figure 2a) the direction of the velocities at lithospheric nodes changes with respect to the inward mantle flow. At 5 ka BP, i.e. 5,000 years after the ice cap has completely melted away, the velocities at the surface point outward while the mantle still flows inward. For $10^{21}$ Pa s (Figure 2b) it can be seen that the direction of the deformation rates at the surface does not change with respect to the mantle flow, and is still inward at 5 ka BP (and later, see Figure 1).

The results in Figure 2 show that the sign of horizontal velocities depends on the interplay between the inward motion of the mantle flow and the outward motion of the lithosphere. Depending on the magnitude of the mantle viscosity, either the shear exerted on the lithosphere by the mantle flow at present-day is strong enough to cause inward horizontal velocities at the surface, or is not, in which case outward velocities are predicted. We show in Supplementary Figure 1 that for low enough mantle viscosities, the moment at which
horizontal velocities start to point outwards occurs earlier for lower viscosities. In Supplementary Note 1 & 2 we show that the sign change depending on viscosity, also holds when varying the shape, extent and amount of ice of the axisymmetric cap.

**Horizontal Velocities due to Ice Melt at the Ross Sea**

We now investigate whether the dependency of horizontal velocities on viscosity also holds for a more realistic deglaciation history. We apply the W12 deglaciation history (Whitehouse et al., 2012) on the same homogeneous Earth models as before. We only consider the deglaciation at the Ross Sea (see Supplementary Figure 3 for the isolated ice load). The resulting velocities (see Figures 3a-d) also point outwards for $10^{20}$ Pa s and lower, and inwards for $10^{21}$ Pa s and higher. However, for a viscosity of $10^{20}$ Pa s the relaxation time is such that the peak of the uplift rates and the largest outward horizontal velocities are concentrated in the South of the RSE, which is caused by the retreat of the Ross ice sheet from North to South (see Supplementary Figure 2 for the evolution of the ice sheet). For $10^{19}$ Pa s these effects can be recognized at earlier timesteps (see Supplementary Figure 4). We conclude that for specific viscosities the predictions also depend on major spatial variation in ice melt over time.

Next, we study the effect of a schematic viscosity transition (see Methods for an explanation of the transition) at constant longitudes of 155°E and 50°W. The ice dome at the RSE lies on the West Antarctic side of the transition. Predictions are computed for a transition of radially homogeneous viscosities of $10^{19}$, $10^{20}$ and $10^{21}$ Pa s in West Antarctica to $10^{23}$ Pa s in East Antarctica, and for a transition of $10^{19}$ to $10^{21}$ Pa s (see Figures 3e-h). In Figures 3e-g, horizontal velocities in West Antarctica show the same direction as for a laterally homogeneous mantle viscosity with the same magnitude (Figures 3a-c). In East Antarctica, the horizontal velocities point inwards and are much smaller, as expected for a high viscosity of $10^{23}$ Pa s.

In contrast with the homogeneous solutions, velocities at both sides of the viscosity transition point away from or towards the transition rather than away from or towards the principle load centre. This fundamentally different flow pattern is caused by the stiff mantle underneath East Antarctica which blocks the outward mantle flow towards East Antarctica during ice loading (see Supplementary Figure 5 for an example). This flow cannot be mimicked with 1D Earth models, since we have shown that for a homogeneous mantle the horizontal flow patterns East and West of the ice load do not differ in direction but are approximately symmetric around the ice load. When the magnitude of the viscosity transition is decreased (see Figures 3g-h), the mantle beneath East Antarctica accommodates more of the stress from the ice load, resulting in larger horizontal velocities in East Antarctica. Thus, for a viscosity transition from $10^{19}$ to $10^{21}$ Pa s, significant velocities in East Antarctica point towards the transition, and away from it in West Antarctica, agreeing with findings in previous studies (Kaufmann et al., 2005, Wang and Wu, 2006).
Figure 3 | Predictions of horizontal velocities for a realistic loading and unloading scenario derived from W12 (Whitehouse, 2012) at the RSE, applied to different Earth models. Predictions for an Earth Model with a homogeneous mantle viscosity of (a) $10^{19}$, (b) $10^{20}$, (c) $10^{21}$ and (d) $10^{22}$ Pa s, and for an Earth model with a schematic lateral transition from (e) $10^{19}$, (f) $10^{20}$ and (g) $10^{21}$ Pa s in West Antarctica to $10^{23}$ Pa s in East Antarctica, and (h) from $10^{19}$ to $10^{21}$ Pa s.
Constraining the Lateral Viscosity Transition with Horizontal GPS Measurements

GPS velocities West of the RSE point towards the ice, while East of the RSE they point away from the ice (see Figure 4c). Based on our earlier findings, it is highly likely that this can only be explained with a lateral viscosity transition between West and East Antarctica. We now attempt to constrain the viscosity at either side of the transition with the horizontal velocities surrounding the RSE of a recent ANET GPS solution (Wilson et al., 2017-in prep), see Methods for a description. We do not derive a fully 3D rheology through olivine flow laws from a temperature distribution in the UM and constrain grain size and water content, like Van der Wal et al. (2015), because we show in Supplementary Note 4 that using different temperature distributions can result in differences in predicted horizontal velocities of up to 0.88 mm/yr. Instead, we directly constrain the viscosity at either side of the schematic viscosity transition. The values we find can be used to constrain the effective viscosities of fully 3D rheologies in future studies. We compare the predictions of 24 different Earth models (see Supplementary Table 2) with the GPS. Viscosity is varied in two layers in the UM at either side of the transition to test for the absolute viscosity value, and the depth and magnitude of the transition. We do not vary LM viscosity and lithospheric thickness to avoid excessive computational time. UM viscosities of $10^{18}$ and $10^{23}$ Pa s are chosen as lower and upper limits respectively. Before comparison, GPS measurements are corrected for an estimated plate motion using 9 stable sites in East Antarctica. With the same sites a common rotation-like signal of the predictions is removed as well (see Methods). As a result, the corrected predictions at the RSE are sensitive to the viscosity at the East Antarctic coast. Finally, we acknowledge that horizontal postseismic deformations could be significant in Antarctica (King and Santamaria-Gómez, 2016a), but do not include them here since we cannot reliably model their effect.

The combined root mean square (rms) misfit of the East and North components of the horizontal velocity predictions with the GPS is computed for each Earth model (see Methods, and Supplementary Table 2 for the misfits). Earth models with the lowest rms, and for which also the inward or outward directions of the predictions with respect to the ice centre agree with that of the GPS observations in all regions surrounding the RSE, give preferred viscosities of $10^{18}$ to $10^{19}$ Pa s from 50 to 200 km and $10^{19}$ Pa s from 200 to 400 km for West Antarctica, and of $10^{22}$ Pa s from 50 to 200 km and $10^{21}$ to $10^{22}$ Pa s from 200 to 400 km for East Antarctica. This is lower in West Antarctica and higher in East Antarctica than the preferred homogeneous UM viscosity of $10^{21}$ Pa s of W12 and $0.5 * 10^{21}$ Pa s of ICE-6G_C (VM5a) (Argus et al., 2014). Since only GPS sites surrounding the RSE are used, these values are probably less valid further away.

The predictions of our best fitting model ($10^{19}$ Pa s from 50 to 400 km in West Antarctica, 1e22 from 50 to 200 km and $10^{21}$ Pa s from 200 to 400 km in East Antarctica) have a rms misfit of 0.718 mm/yr, and are shown in Figure 4a. Especially at the Central region (see Figure 4c), the predictions fit the direction and the small magnitude of the GPS well, which has not been resolved before. However, at most sites the predictions seem to systematically point too much to the North. It was found that this deviation can be resolved when the longitude of the transition is changed. The direction of the transition at 155°E was based on the mean direction of the lateral contrast in the temperature model of Pappa et al. (2017). Looking at the location of shear wave velocity contrasts in e.g. models of An et al. (2015) and
Heezel et al. (2016), it would be justified to vary the longitude of our schematic viscosity transition to up to 163°E. For the predictions of the best fitting Earth model with a transition at 163°E (Figure 4b), especially at the lower-Northern sites a better fit is achieved and the rms is reduced to 0.641 mm/yr. Varying the longitude of the schematic lateral transition primarily affects the predictions near the RSE and not these the East Antarctic coast, as the rotational portion that is removed from the predictions before comparison with GPS is similar for a longitude 155°E and 163°E (Figures 4d-e).

Figure 4 | Horizontal GPS measurements of ANET/POLENET and predictions of horizontal velocities for the best fitting Earth model with a viscosity of $10^{19}$ Pa s from 50 to 400 km depth in West Antarctica and of $10^{22}$ from 50 to 200 km and $10^{21}$ Pa s from 200 to 400 km in East Antarctica. (a) GPS velocities and predictions at the RSE for the best fitting Earth model with a transition at a constant longitude of 155°E. (b) GPS and predictions for the best fitting Earth model with a transition at the RSE at a constant longitude of 163°E. (c) ANET/POLENET GPS measurements surrounding the RSE with 1σ uncertainty (see Supplementary Note 8) ellipses. (d) Euler velocity field with rotational residual of GIA predictions in (a). (e) Euler velocity field with rotational residual of GIA predictions in (b).
For most of the models the predictions at the most Northern sites VL01, VL12 and VL30 are too small. We show in Supplementary Note 7 that for two representative Earth models additional ice loss at the WSB does not lead to significantly larger velocities at these sites. Another explanation might be found in the methodology used to derive the GPS solution, as the horizontal solutions for the VL01, VL12 and VL30 sites are much smaller in the solution of Zanutta et al. (2017) than in the one we use. Differences in final relative velocities of a solution arise from different approaches to data processing (e.g. different software, length of data) and plate motion estimation, and should be further investigated. Although East of the RSE both our predictions and the GPS point away from the ice, the predictions point too much to the North. Close to the South Pole, none of our Earth models fit the measurements well. Uncertainties in the amount of ice loss, the timing of ice loss and the location of the maximum ice loss in the deglaciation history probably have a significant contribution to the misfit. A stronger mass loss at the North of the RSE might result in the prediction of larger velocities at the most Northern sites and a better fit with GPS. However, this also depends on mantle viscosity. It should be noted that our schematic viscosity transition is a simplification of reality. It is unlikely that the transition in viscosity has a constant longitude as was assumed here. Additionally, if more lateral variations in viscosity are captured in our model a better fit might be obtained.

To investigate the effect of LM viscosity and lithospheric thickness, we vary these parameters for our best fitting Earth model. LM viscosity is varied from 400 to 670 km and 670 km and deeper using the preferred values of (1) W12: $10^{21}$ and $10^{22}$ Pa s (Whitehouse et al., 2012), (2) of ICE-6G_C: $0.5 \times 10^{21}$ and $3 \times 10^{21}$ Pa s (Argus et al., 2014), and (3) the untested combination: $10^{21}$ and $3 \times 10^{21}$ Pa s, respectively. The largest deviation in rms we find with respect to the original misfit is an improvement of 0.031 mm/yr for combination 3 (see Supplementary Note 6 for all misfits). Since this difference is small, varying LM viscosity within the preferred boundaries reported in the literature will not significantly affect the preferred range of UM viscosities we find. Additionally, we compute predictions for our best fitting Earth model with a constant lithosphere thickness of 70 km and a transition from 50 to 70 km (see Methods for how this is modelled), similar to the preferred lithospheric thickness of 71 km of W12 for East Antarctica (Whitehouse et al., 2012). Based on the results for a thickness of 70 km which were not yet transformed to the GPS frame, it can be seen that a higher lithospheric thickness leads to lower velocities in both West and East Antarctica. However, since also the estimation of a common rotational signal of the predictions is affected, the general effect of varying lithospheric thickness is difficult to interpret. The transformed predictions for a lithosphere of 70 km are bigger than for 50 km, and the resulting misfit is improved to 0.670 mm/yr. For a transition from 50 to 70 km, the predicted velocities at the RSE are smaller and are rotated slightly to the North. The resulting misfit is 0.746 mm/yr.
**Implications**

Depending on mantle viscosity, the direction of horizontal velocities can be reversed. When W12 is applied in combination with a schematic lateral transition in viscosity between West and East Antarctica, a horizontal flow pattern results which cannot be mimicked with any 1D rheology, and which agrees with observations surrounding the RSE. Thus, the ANET/POLENT horizontal GPS measurements require a lateral transition in viscosity. With this transition we find preferred average viscosities to lay between $10^{18}$ and $10^{19}$ Pa s in West Antarctica, and $10^{21}$ and $10^{22}$ Pa s in East Antarctica. At the central and lower Northern regions, the predictions of our best fitting Earth model agree well with the GPS. However, further away predictions are too small or point too much to the North. We argue that these deviations arise mainly due to uncertainties in the W12 deglaciation history, and the fact that we applied only a single transition in viscosity. When more certain UM temperature distributions become available for Antarctica, the simplifications in our Earth models can be addressed when a fully 3D rheology is adopted. Then, also a preferred lithospheric thickness can be found simultaneously with mantle viscosity.

Our results confirm that the contrast in shear wave velocities between West and East Antarctica translates to a viscosity transition in the UM of at least two orders of magnitude. The low viscosity we find for West Antarctica is in line with results for other regions in West Antarctica such as the AP (Nield et al, 2014), and results in lower uplift rates over the Ross Sea than predicted with the preferred models of W12 and ICE-6G_C. This agrees with the mean uplift rates of 3 inverse models of GIA (Martín-Español et al., 2016). The evidence for a significant lateral transition in viscosity between West and East Antarctica implies that 1D GIA models are unrealistic when used for the whole of Antarctica, as they overestimate the viscosity in West Antarctica. Thus, estimates of the total ice mass loss and the evolution of the AIS following from GRACE mass balance estimates that are corrected for GIA with 1D models are indeed biased, as was suggested by Van der Wal et al. (2015). A predominantly low viscous mantle beneath West Antarctica also implies a higher sensitivity to late-Holocene surface mass variability. Additionally, as viscosity drives plate tectonics, lateral viscosity variations could influence stresses in the lithosphere. In order to compute accurate predictions of postseismic velocities in Antarctica (e.g. King and Santamaría-Gómez, 2016), viscoelastic models should include the lateral transition in viscosity between West and East Antarctica that we confirmed. When more GIA studies start adopting fully 3D rheologies that capture lateral viscosity variations, the dependency of horizontal velocities on viscosity demonstrated here can be used to tightly constrain 3D GIA models.
Methods

Coupled FE-Laplace GIA Model. The GIA model uses the commercial finite-element software Abaqus version 6.14, following the iterative coupled FE-Laplace approach of Wu (2004). In the main article a 3D model (Van der Wal, 2013, Van der Wal, 2015) is used, and for the tests in Note 1, 2 and 3 of the Supplementary Information an axisymmetric model is used (Wu and Van der Wal, 2003). In both models advection of pre-stress and self-gravitation are taken into account through iterative computation but geocenter motion is not accounted for. The model is incompressible. Horizontal motion induced by GIA is thought to be sensitive to compressibility, but when compressibility is taken into account following the approach of Wu (2004), possible numerical instabilities may arise related to the representation of internal buoyancy due to the perturbed density when elements are deformed.

Only material compressibility can be included, i.e. changing the Poisson’s ratio and Young’s modulus. However, material compressibility cannot be separated physically without the associated density perturbations (Klemann et al., 2003). Nevertheless, the effect of material compressibility was tested on our best fitting Earth model. Only the magnitude of the predictions of horizontal velocities was found to be significantly affected, and the resulting rms with the observations is 0.708 mm/yr. The difference of that misfit with the original rms of 0.718 mm/yr for the best fitting, incompressible Earth model, is very small. For an Earth model with a much higher viscosity the effect of compressibility on the rms is similar small. Thus, we conclude that including material compressibility will not change the outcomes of this study. When studying small wavelength effects (lower than 8000 km), the numerical instabilities in the 3D GIA model when including full compressibility might be circumvented by adapting the elastic rigidity of an incompressible model to approximate the flexural rigidity of a compressible model (Tanaka et al., 2011). Future studies should investigate the effect of full compressibility on predictions of horizontal rates in the RSE.

For easier interpretation of the predictions of horizontal velocities, we choose not to consider the effect of ocean loading, and prescribe a bedrock topography of 0 m everywhere. Additionally, with our schematic viscosity transition ocean loading cannot be incorporated consistently, as the viscosity profile used to approximate the mantle underneath Antarctica is not realistic further away. We show in Supplementary Note 5 that horizontal deformation rates at the GPS sites in the RSE due to ocean loading are on average 0.03 mm/yr for a relatively low and 0.055 mm/yr for a relatively high 3D mantle viscosity. For both cases, by omitting ocean loading a bias is introduced that is similar to or lower than the measurement uncertainties at most GPS sites. Thus, neglecting ocean loading will not significantly impact our results.

3D GIA Model. The GIA model we use for the tests in the main article is a computationally intensive 3D model, for which the computation of one iteration typically takes around a day with eight processor cores. We show in Supplementary Note 3 that for a representative Earth model, the maximum difference in horizontal velocities between the results after two and three iterations is 3.93%. This is sufficiently small to perform only two iterations for each test, which allows to study a wider variety of parameters. The mesh of the 3D Earth model exists of 178,560 elements. The inner part of the spherical mesh, i.e. the fluid, inviscid core of the Earth, is omitted as it does not play a role in GIA. In the lowest four layers (LM) elements are spaced on a $4^\circ \times 4^\circ$ grid. In the layers above and on the surface, the resolution of the
elements is $2^\circ \times 2^\circ$. Eight-node brick elements are used except for at the poles, where 6-node wedge elements are defined to create a spherical Earth.

For the 3D GIA model, we use a baseline Earth model with 14 layers. For eight distinctive layers elastic parameters are obtained from volume-averaging of the Preliminary Reference Earth Model (Dziewonski & Anderson, 1981) and are the same as in Van der Wal et al. (2013). See Supplementary Table 1 for a detailed overview of how the layers are defined. Density contrasts are present at layer boundaries at the surface, at 50, 200, 275 and 325 km depth, and at seismic discontinuities at 400, 670, 1171 and 2891 km depth. The density jumps at these boundaries are accounted for in the boundary conditions for the advection of pre-stress by applying Winkler foundations with a spring constant equal to the difference in density times the unperturbed gravitational acceleration at the corresponding depth (Wu, 2004). The lithospheric layer is modelled to deform elastically only, whereas the other layers are viscoelastic, with a viscosity distribution that is varied for each specific test.

**Ice Models.** In addition to the properties of the Earth model, the loading and unloading of the ice that is represented in a deglaciation history is the second key ingredient to the predictions of deformation rates due to GIA. To study the fundamental dependency of horizontal velocities on mantle viscosity we use an axisymmetric ice cap on the North Pole. To reduce the Gibbs effect, the ice is given an elliptical shape that is parameterized as in equation (1):

$$h_{ice}(\phi, t) = H(t) \frac{\cos(\phi) - \cos(\alpha)}{1 - \cos(\alpha)}$$

where $\phi$ is the colatitude, $\alpha$ the colatitude of the ice margin, and $H$ the height of the ice at the centre of the ice cap. The peak height is taken to be 1500 m, and the ice margin is located at $14^\circ$ colatitude. The ice cap is modelled to grow linearly during 90 ka before it reaches its peak height at 20 ka BP, and then melts linearly during the next 10 ka. So, at present the ice has been gone for 10 ka. The ice load at each timestep is applied as a distributed load at the top of the corresponding surface elements. For our 3D tests we use the W12 deglaciation history (Whitehouse et al., 2012). An important motivation to use W12 is that this deglaciation history has not been tuned to fit observations assuming a laterally homogeneous Earth and is based on a physically-consistent numerical model, in contrast with other deglaciation histories that are often used for Antarctica like IJ05_R2 (Ivins et al., 2013) and ICE-6G_C (Argus et al., 2014). In W12, the ice grows during 90 ka until 20 ka BP, after which it decreases linearly. The ice sheet is reconstructed using palaeo-temperature records, surface temperatures and related accumulation rates and historic records of RSL, and tuned to reproduce the geometry of the present-day ice sheets at Antarctica. Different flow laws for basal sliding are used to model faster flow at concave ice profiles like in the Ross Sea (Whitehouse et al., 2012). In W12, late-Holocene ice changes in Antarctica are not included. An elastic correction of predicted velocities for ongoing ice mass changes is not applied here, as the mean elastic trend in the RSE due to recent mass variability is thought to be relatively small (Martín-Esparzol et al., 2016).

The W12 model ice load is interpolated onto the $2^\circ \times 2^\circ$ grid of our 3D GIA model (see Supplementary Figure 3). The changes in ice thickness are given at intervals of 1 ka. The zeroth and first degrees of the ice load are removed to avoid a shift in the centre of mass of the Earth, because a translation of the Earth is not constrained in Abaqus. Ice loads are included up to spherical harmonic degree 90. Only considering loads up to degree 60 does
not significantly affect our predictions, so results from this study can directly be compared with inverse models using GRACE observations that are considered accurate only up to degree 60. In the tests on the dependency of horizontal velocities at the RSE on homogeneous viscosity and on a simple lateral viscosity transition, only the ice at the Siple Coast in the W12 model was used (see Supplementary Figure 3). In the tests with a schematic viscosity transition between West and East Antarctica, all the ice in Antarctica is considered. The tests discussed in Supplementary Note 7 contain an additional ice wedge at the Wilkes Coast that starts to grow linearly at 26 ka BP and decreases linearly starting from 20 ka BP (Konfal et al., 2015).

**Lateral Viscosity Transition.** The lateral contrast in viscosity between East and West Antarctica is modelled as a schematic viscosity transition with constant longitudes of 155°E on one half of the Earth and 50°W on the other (see Supplementary Figure 2 for an illustration of the transition). This contrast is based on a temperature distribution in the UM for Antarctica from LitMod3D (Pappa et al., 2017), and is a feature that is also recognized in seismic models for Antarctica (e.g. An et al, 2015). The assumption that the viscosity transition has a constant longitude approximates reality well close to the RSE, but is less realistic near the South pole and Ronne ice shelf. For tests used to study the qualitative effect of such a transition on horizontal velocities, we assume a radially homogeneous viscosity on both sides of the transition. For the comparison of predictions with GPS, we vary the viscosity between 50 and 200 km and 200 and 400 km. Below that, we use a viscosity of $0.5 \times 10^{21}$ Pa·s in the TZ, and $10^{22}$ Pa·s in the LM, which are the extreme values of the preferred 1D Earth rheology for ICE-6G_C (Argus et al., 2014) and W12 (Whitehouse et al., 2012). The set of 24 different Earth models based on this schematic transition can be seen in Supplementary Table 2. Because our schematic viscosity profile is not realistic in other continents, it is difficult to consistently include the effect of ocean loading and global deglaciation in our model, as the viscosity globally will affect the propagation of sea-level and global deglaciation in predictions of horizontal velocities in Antarctica. Consequently, we neglect the effect of ocean loading and global deglaciation. It was found by Wang and Wu (2006) that the the effect of the ice sheets in the Northern hemisphere have only a small effect on horizontal motion in Antarctica. Moreover, when low viscosity zones are applied to represent lateral variations in Earth structure at plate boundaries it is expected that long-wave stress transmission will be limited (Latychev et al., 2005, Whitehouse et al., 2006). Additionally, we show in Supplementary Note 5 that neglecting ocean loading leads to biases in the predictions that are below measurement uncertainty for almost all GPS sites.

A transition in lithospheric thickness is also applied in our best fitting Earth model in terms of a transition in viscosity of a layer. A lithospheric element is given a quasi-infinite viscosity such that it deforms only elastically. A fully consistent transition in lithospheric thickness would also include a contrast in the extent of the depth of lithospheric elastic parameters and density. However, this significantly complicates applying proper Winkler foundations in the code of the GIA model in order to account for density jumps at the layer boundaries. Therefore, the transition in lithospheric thickness is only modelled as a transition in viscosity.

**Computing Horizontal Velocities and Comparison with GPS.** In the model of Van der Wal et al. (2015) adopted in this study, present-day deformation rates are computed as the difference between the deformation at present and at 50 years in the future, divided by 50 years. Therefore, the predictions are centered around 25 years in the future. This was done to avoid the elastic contribution to deformation rates due to the late Holocene changes in ice
thickness at the Antarctic Peninsula in the W12a deglaciation history. Although in this study W12 is used, the timesteps were not changed in the computation of our predictions. However, the predictions of horizontal velocities centered around present or around 25 years in the future differ less than 0.01 mm/yr, which is a negligible difference. Deformation rates at all other timesteps between 20 ka BP and present are computed as the average deformation of that timestep and the timestep before, divided by 1,000 years. Deformation rates at 20ka BP are computed as the average deformation over 90,000 years of the linear ice build-up.

The measurements of the GPS solution (Wilson et al., 2017-in prep) that we compare to our GIA predictions have been derived through the Antarctic Network (ANET) part of the Polar Earth Observing Network (POLENET) and were provided by Stephanie Konfal of Ohio State University. The observations of the GPS solution have continuously been gathered for more than a decade. The solution was obtained through processing of the data with GPS analysis software called GAMIT/GLOBK (Herring et al., 2010). Horizontal velocities relative to estimated plate motion were obtained following the approach of Bevis et al. (2013). Plate motion was estimated and corrected for using 9 stable sites at the East Antarctic coast, in a manner that is independent on the treatment of vertical velocities. Essentially, with this plate motion is removed. However, also a common rotational mode of GIA is removed from the observations with this approach. Thus, for a consistent comparison, also the rotation-like portion of the GIA predictions is removed using the best fit Euler velocity field for the same 9 GPS sites in East Antarctica. This was done by Michael Bevis of Ohio State University. This approach was also suggested in a previous study dedicated to how to use horizontal GPS measurements to validate GIA models in Antarctica (King et al., 2016b).

Although consistent, this correction of rotation has the downside of neglecting a meaningful part of the predictions. Also, in our tests we found that the rotational residual depends strongly on the viscosity that is chosen underneath the East Antarctic GPS sites that are used to estimate the Euler flow field. In most shear wave velocity models of Antarctica there is evidence that the stiff cratonic root underneath East Antarctica extents from the TAM to further than the sites at the East Antarctic coast, but it is not certain if the viscosity assumed close to the RSE is also valid there. Nevertheless, the rotational residuals will most likely be more realistic than when the same methodology is applied to a 1D Earth model as in (Argus et al., 2014), as the assumption that the viscosity underneath the East Antarctic coast is the same as underneath West Antarctica is likely incorrect. The fit of the predictions of horizontal velocities, of which a common rotational signal is removed, with the GPS measurements surrounding the RSE relative to estimated plate rotation, is assessed qualitatively using figures and quantitatively using the combined rms misfit of the longitudinal and latitudinal components of the predicted horizontal velocities, so not only the magnitude of the velocity vector.
References


**Methods-only**


**Acknowledgements**

*To be completed.*

**Author Contributions**

*S.A. Konfal provided the g08c GPS solution of ANET/POLENET.*

*M.G. Bevis transformed the GIA predictions to the g08c GPS frame.*

*To be completed.*

**Additional Information**

The authors declare no competing financial interests. Supplementary information accompanies this paper in the following chapter.
4

Supplementary Information
Supplementary Information

Constraining Glacial Isostatic Adjustment with Horizontal GPS Velocities in Antarctica

T.b.d.

This document provides supplementary information to the draft journal article “Constraining Glacial Isostatic Adjustment with Horizontal GPS Velocities in Antarctica”. This supplementary information contains notes about tests that were not discussed in detail in the article, and additional figures and tables that are referred to in the main text.

Supplementary Notes

1. Benchmark of 3D and Axisymmetric GIA Models
2. Effect of Axisymmetric Ice Shape
3. Effect of Convergence and Creep Tolerance Parameter
4. Variability in Results of 3D Viscosity based on Olivine Flow Laws
5. Effect of Ocean Loading
6. Effect of Lower Mantle Viscosity
7. Effect of an Ice Wedge at the Wilkes Subglacial Basin
8. ANET/POLENET GPS Solution

Supplementary Figures

1. Nodal Velocities at Layer Boundaries
2. Schematic Viscosity Transition
3. W12 Deglaciation History
4. Horizontal Velocities through Time for Ice at the RSE on a 1D Earth
5. Horizontal Velocities through Time for Ice at the RSE on a 3D Earth

Supplementary Tables

1. Baseline Layering of 3D GIA Model
2. Viscosity of Earth Models and Rms Misfits of Predictions with GPS

Supplementary references are provided at the end of this document.
Supplementary Notes

1. Benchmark of 3D and Axisymmetric GIA Models

In the main article we use a computationally intensive 3D GIA model (Van der Wal, 2013, Van der Wal, 2015). In the supplementary information we also use a less intensive axisymmetric GIA model (Wu en van der Wal, 2003), for tests that only require axisymmetric settings. The mesh of the axisymmetric model is a half-circle, which is rotated around the axis of symmetry. The mesh consists of 4693 nodes with a grid spacing of 0.5°, making up for a total of 4,320 four-node axisymmetric elements. Thus, the axisymmetric model has a higher grid resolution than the 3D GIA model. The computation of four iterations with the axisymmetric model takes around half an hour, enabling to carry out more tests than with the 3D model.

We compare the two models to see if (a) the conclusions drawn from one model can be extrapolated to the other, given the use of appropriate settings, and (b) to uncover the effect of any exclusive model artefacts on the results. For the benchmark we test two simple cases in which an ice cap is applied on the North Pole of the Earth with a peak height of 1500 m and a radius of 14° colatitude (see Figure S1.1 for a schematic representation). Two mantle viscosities of $10^{20}$ and $10^{21}$ Pa are tested, and the Earth model is radially and laterally homogeneous.

![Figure S1.1](image)

**Figure S1.1** | Schematic overview of the axisymmetric ice cap on a homogeneous Earth, in which $\eta$ denotes the homogeneous mantle viscosity (adapted from (Wu and Van der Wal, 2003), Figure 3)

The defined layering and material parameters of the models are shown in Table S1.1. In the axisymmetric GIA model only a limited number of density jumps is modelled through Winkler foundations. Note that in the 3D model the gravitational acceleration in the Winkler foundations is recomputed based on the actual displacement of the nodes at the boundary for every iteration, whereas for the axisymmetric model the gravitational acceleration at the layer boundaries is constant over the iterations, but this is likely to lead to only a very small deviation. The 3D model has more layers, but a lower grid resolution ($2° \times 2°$).
Table S1.1 | Benchmark settings for material parameters in Earth layers. Elasticity, density, gravitational acceleration and compressibility settings are shown for the lithosphere, UM, TZ, LM1 and LM2, used in both GIA models.

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<th>g (m/s²)</th>
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Figure S1.2 | Benchmark of predicted horizontal velocities centred around 500 years BP for 3D and axisymmetric GIA model. (a) Predictions for a homogeneous mantle viscosity of $10^{20}$ Pa s. (b) Predictions for a homogeneous mantle viscosity of $10^{21}$ Pa s.

The resulting horizontal velocities from both models after 1 iteration are shown in Figure S1.2. Because the present-day velocities of the 3D GIA model are centred around 25 years in the future, in this benchmark the velocities from both models are compared centred around 500 years BP. For 1e20 Pa s, the horizontal velocities agree well. The maximum difference between the two models is 0.04 mm/yr, which is sufficiently close for our study. It is most likely that this is due to the difference in resolution of the two models. For $10^{21}$ Pa s, a remarkable feature is observed between 14° and 18° colatitude, corresponding spatially with the presence of the forebulge. The axisymmetric GIA model predicts a dip in the inward velocities that is not recognized in the results of the 3D GIA model, leading to differences between the two models of over 0.1 mm/yr. It is believed that the dip is caused by the subsidence of the forebulge, which is not captured by the 3D GIA model accurately because of its lower resolution. For $10^{20}$ Pa s this dip cannot be seen, since the forebulge has already relaxed. We conclude that for more accurate computations of horizontal velocities especially close to the ice margin it is advisable to use GIA models with a better resolution.
2. Effect of Axisymmetric Ice Shape and Height

In the main article we found that for mantle viscosities of \(10^{20}\) Pa s and lower GIA-induced horizontal velocities point away from the former ice centre, and for \(10^{21}\) Pa s and higher they point towards the former ice centre. Here we show that the presented findings also hold when the ice shape, height and radius are varied.

We repeat the test in the main text with the axisymmetric GIA model with a homogeneous viscosity of \(10^{20}\) Pa s and of \(10^{21}\) Pa s. The layering, elasticity and density are as in Supplementary Note 1. We define 5 variations of the ice cap loading and unloading scenario. For the first, we define a constant ice height at all colatitudes, resulting in an ice disc with a volume equal to that of an ice cap with a peak height of 1500 m. For the second and the third scenario we change the peak height to 750 m and 3000 m respectively. For scenario four and five, we keep the peak height of 1500 m, but change the ice radius to 11° and 17° colatitude respectively.

Figure S2.1 | Predicted present-day horizontal deformation rates as a function of ice shape, height and margin. (a) Results for a homogeneous mantle viscosity of \(10^{20}\) Pa s. (b) Results for a homogeneous mantle viscosity of \(10^{21}\) Pa s.

The predictions following from each deglaciation scenario can be seen in Figure S2.1. It is observed that the results from an ice cap and an ice disc mainly differ close to the ice margin. The outward velocities seen for a mantle viscosity of \(10^{20}\) Pa s (Figure S2.1a, top) and the inward velocities for a viscosity of \(10^{21}\) Pa s (Figure S2.1b, top), have a higher magnitude for an ice disc than for a cap. For a viscosity of \(10^{21}\) Pa s, a more significant dip in
the peak of the horizontal velocities can be seen around 14° colatitude for the ice disc scenario. This is thought to be an effect due to the deformation of the lithosphere just before and at the forebulge. The ice disc has a higher ice height at its margin, and therefore the effects of a forebulge would be more significant. The relaxation time for a viscosity of $10^{20}$ Pa s is smaller, so there no dip in the peak of the present-day horizontal velocities can be observed. For both an ice cap and a disc, the direction of the velocities with respect to the ice as a function of viscosity centre remains unchanged, suggesting that regardless of these ice shapes the dependency of horizontal velocities on mantle viscosity shows the same trend.

It is also showed in Figure S2.1 that for a lower and higher peak ice height, the predicted velocities have a lower and higher magnitude respectively, but the sign of the velocities is the same. We therefore conclude that for an ice cap the direction of the horizontal velocities is not sensitive to the amount of ice. When varying the ice margin (Figure S2.1a and b, bottom), we observe for both mantle viscosities that the colatitude of the maximum of the horizontal velocities, which is located just outside the ice margin, is shifted accordingly. The magnitude of the predicted velocities is also affected slightly, because a change in the ice margin results in a change of total pressure exerted on the surface. Again, the sign of the horizontal velocities is unaffected by the changes in the ice.

A final note is made on the Gibbs effect that could result from a discontinuity in the load at the ice margin. In the axisymmetric GIA model, the ice load is tapered in the frequency domain above a certain maximum degree. The purpose of this taper is to reduce the Gibbs effect when an ice disc is defined as deglaciation history. In the 3D GIA model this taper is not defined. With a test it was found that varying the maximum degree above which this taper is applied has a marginal effect on the results, and even less so for when an ice cap instead of an ice disc is applied. Moreover, the benchmark study in Supplementary Note 1 shows that regardless of the load taper, the results of both GIA models agree sufficiently well.

3. Effect of Convergence and Creep Tolerance Parameter

The predictions by the 3D GIA model are computed with only two iterations because of the substantial computational time every iteration takes. The significantly smaller computation time of the axisymmetric GIA model warrants the use of four iterations for that model, after which typically an acceptable convergence is achieved (Wu, 2004). For the 3D GIA model, we show with an arbitrary viscosity profile taken from Supplementary Table 2 how the results after two and after three iterations differ. The particular 3D Earth model has viscosity values of $10^{19}$ and $10^{21}$ at 50-200 and 200-400 km depth at the West Antarctic side of the lateral viscosity transition, and $10^{23}$ and $10^{22}$ at the East Antarctic side.

We find that between one and two iterations, the maximum absolute difference in the predictions of horizontal velocities is a significant 12.23%. Between two and three iterations, the predictions only vary maximally by 3.93%. This corresponds with an absolute difference of 0.04 mm/yr. When looking at the solution for GPS observations of POLENET/ANET, the uncertainties of horizontal measurements at most sites are higher than this number. Therefore, we conclude that for the purpose of our study two iterations is a sufficiently good compromise between computation time and precision.

For the time-dependent viscoelastic behaviour of the mantle elements, in Abaqus a creep error tolerance parameter CETOL is used that governs the maximum difference in creep strain increments, controlling the accuracy of the creep integration. Generally, the creep
strain increment should be much smaller than its elastic counterpart. It was explained in (Van der Wal, 2013) that a more stringent creep error tolerance does not significantly change the uplift rates and sea level resulting from the GIA model. We perform a similar test for the predictions of horizontal velocities using the same Earth model as before. At all timesteps, the CETOL parameter is taken one order of magnitude smaller (i.e. divided by ten). After 1 iteration, we find a maximum difference of 4.63%. This is a similar difference as between two and three iterations of the standard GIA model. Although the resulting predictions are more accurate with a smaller CETOL, the computation time was more than twice as big as for the standard settings. As in our study it is important to perform many tests to study a large parameter space, we choose not to use a more stringent creep error tolerance. This will give a slight inaccuracy in our results.

4. Variability in Results of GIA Models with 3D Mantle Viscosity

Instead of using a fully 3D viscosity distribution derived from temperature models, a schematic viscosity transition between West and East Antarctica was used for the tests described in the main article. A realistic 3D distribution of effective mantle viscosity can be obtained with experimental flow laws for olivine (Hirth & Kohlsted, 2003) and a 3D temperature distribution (Van der Wal, 2013). However, these flow laws introduces several new parameters such as grain size and water content, of which the 3D distributions are unknown. Moreover, significant uncertainties are associated to 3D temperature distributions. Temperature models can be obtained from seismic tomography maps using physical scaling laws (Ivins & Sammis, 1995), but cannot be quantitatively derived from seismological data sets without first-order uncertainties when heterogeneities in the composition of major bulk elements are not considered (Trampert & Van Der Hilst, 2005). As a result, there is large variability in the predictions resulting from the 3D viscosity distributions for Antarctica derived from these temperature models. Here, we compute the difference between the predictions of horizontal velocities following from two Earth models with a 3D viscosity distribution based on two different temperature models.

We derive an effective 3D mantle viscosity distribution through olivine flow laws from two temperature models with the approach of Van der Wal (2013). The first temperature model we use is based on the global seismic model of Shaeffer & Lebedev (2013), and refined locally in Antarctica with the model of Heeszel et al. (2016). The second model is based on a local 3D temperature model for Antarctica (Pappa et al, 2017). This is a model that is primarily based on observations of gravity gradients, isostatic equilibrium and seismological models. The temperature models are referred to as SLHE and Kiel respectively. In the flow laws it is assumed that until 400 km depth, olivine is the main mantle material. Water content and grain size are varied as the main parameters in the flow laws, and, because their 3D distribution is unknown, are assumed to be radially and laterally homogeneous. We select a discrete combination of water content and grain size that gave the best fit to GPS in previous studies (Van der Wal et al., 2013, Van der Wal et al., 2015), with a completely dry rheology and a grain size of 4 mm. Layer definitions and material properties are identical to those used by Van der Wal (2013), and are shown in Table S4.1 as well. The viscosity below 400 km is $0.5 \times 10^{21}$ Pa s in the TZ, and $10^{22}$ Pa s in the LM, identical to the tests in the main article.
Figure S4.1 | Predicted present-day horizontal deformation rates resulting from two Earth models with a 3D mantle viscosity obtained through olivine flow laws and a 3D temperature distribution. (a) Results with the SLHE temperature model. (b) Results with the Kiel temperature model.

The results of the GIA models are shown in Figure S4.1 for the isolated loading and unloading of ice at the Siple Coast. For the same grain size and water content, the predictions of horizontal velocities from the two models have large differences. Evaluated at the coordinates of the GPS sites around the RSE used in the main article, the untransformed predictions show a rms difference of up to 0.88 mm/yr. This is larger than most detected horizontal velocities. We conclude that constraining grain size and water content with GPS is highly dependent on the temperature model of choice. To avoid this, in our study we use horizontal GPS observations to directly constrain mantle viscosity in the RSE. The viscosity values we find can then be used to constrain the effective viscosities of fully 3D rheologies in future studies.

Table S4.1 | Layer definitions and material properties for the two Earth models with a 3D mantle viscosity obtained through olivine flow laws and a 3D temperature distribution.

<table>
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<tr>
<th>Layer</th>
<th>$r$ (km)</th>
<th>$E$ (GPa)</th>
<th>$\rho$ (kg/m$^3$)</th>
<th>$v$ (-)</th>
<th>$g_0$ (m/s$^2$)</th>
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5. Effect of Ocean Loading

To estimate the biases that the omission of ocean loading introduces, we compute the contribution of ocean loading for a fully 3D viscosity distribution and a global deglaciation. To compute RSL, the sea-level equation of Farrell & Clark (1976) is included in the coupled FEM-Laplace model following Wu (2004) and solved iteratively. This part of the model was programmed by Hansheng Wang and validated in Wang et al. (2006). Changes to the centrifugal potential due to changes in sea level are not considered. The effect of meltwater influx, which is especially important in areas with large ice shelves like at the Ross Sea, is incorporated using the ocean function detailed by Mitrovica and Milne (2003). For the bedrock topography we interpolate the ALBMAP product (Le Brocq et al., 2010) on our 2°x2° grid.

We derive two global viscosity distributions from the SLHE temperature model with a dry and a wet rheology and a grain size of 4 mm (relatively high and low viscosities respectively, see also Supplementary Note 4). With these Earth models we compute predictions of horizontal velocities with and without ocean loading for a global deglaciation history. Outside Antarctica we use the ICE-6G (Peltier, 2015) ice model. For both viscosity distributions, the predicted horizontal velocities resulting from the model with ocean loading are subtracted from the model without ocean loading, and corrected for a rotational signal (see Methods). We assume that the residual velocities represent the contribution of ocean loading. The resulting absolute values of these velocities for a dry rheology with a grain size of 4 mm at the GPS sites surrounding the RSE is on average 0.055 mm/yr, which is similar to the measurement uncertainty at most sites. For a wet rheology, the contribution of ocean loading results in an even smaller average velocity of 0.03 mm/yr. The deformation rates due to ocean loading do differ significantly in direction between both viscosity profiles. Therefore, we opt not to derive any eustatic signal from the above tests and add that to the predictions of our models in the main article, as it is likely that then a bigger error is made due to the dependency of the velocities resulting from ocean loading on viscosity than when not including ocean loading at all. We conclude based on two representative Earth models that by not including ocean loading a bias is introduced which is below or similar to the measurement uncertainty at the majority of GPS sites. When a fully 3D viscosity distribution is used, the effect of ocean loading on horizontal velocities can be included in a consistent manner in future studies.

6: Effect of Lower Mantle Viscosity

We explore the effects of LM viscosity and lithospheric thickness on the predictions of our best fitting Earth model. LM viscosity is varied from 400 to 670 km and 670 km and deeper using the preferred values of (1) W12: $10^{21}$ and $10^{22}$ Pa s (Whitehouse et al., 2012), (2) of ICE-6G_C: $0.5 \times 10^{21}$ and $3 \times 10^{21}$ Pa s (Argus et al., 2014), and (3) the untested combination: $10^{21}$ and $3 \times 10^{21}$ Pa s, respectively. The resulting misfits for these variations are 0.726 mm/yr for option (1), 0.694 mm/yr for option (2) and 0.687 mm/yr for option (3). Compared to the original misfit of 0.718 mm/yr, these results show a preference for a higher viscosity from 400 to 670 km and a lower viscosity below 670 km.
7: Effect of an Ice Wedge at the Wilkes Subglacial Basin

In an attempt to explain the inward GPS velocities along the TAM with a 1D rheology, an alternative deglaciation history was investigated by Konfal et al. (2015). In addition to the ice history defined in W12, an ice wedge was added at the Wilkes Subglacial Basin based on glaciological constraints. Both with and without this additional deglaciation scenario in the Wilkes Subglacial Basin, the direction of the most Northern ANET GPS velocities could be resolved for some Earth models, but not those further South.

Here, we investigate what the effect of the additional ice loading and unloading at the Wilkes Subglacial Basin is on the horizontal velocities when the Earth model includes a lateral transition in viscosity between East and West Antarctica as in the set of experiments presented in the main article. We apply the W12 deglaciation history both with and without the additional ice loss in the Wilkes Subglacial Basin to two Earth models with a lateral transition in viscosity. The additional ice history is interpolated on our 2° x 2° grid using the spherical law of cosines. The first model has a relatively low viscosity of $10^{18}$ and $10^{19}$ Pa s in West Antarctica from 50 to 200 and 200 to 400 km respectively, and $10^{21}$ and $10^{21}$ Pa s in East Antarctica, and will be referred to as profile A. The second has a relatively high viscosity of $10^{19}$ and $10^{21}$ Pa s in West Antarctica, and $10^{23}$ and $10^{22}$ Pa s in East Antarctica, and will be referred to as profile B.

![Figure S8.1](image-url) **Figure S8.1** | Horizontal ANET GPS solution and model predictions at the RSE with and without an additional wedge loading and unloading scenario at the Wilkes Subglacial Basin. (a) Model predictions for viscosity profile A without ice wedge. (b) Model predictions for viscosity profile A with ice wedge. (c) Model predictions for viscosity profile B without ice wedge. (d) Model predictions for viscosity profile B with ice wedge.
The results in Figure S8.1 show that for both Earth models, the additional loading and unloading in the Wilkes Subglacial Basin only slightly affects the predictions of horizontal velocities. A small change in magnitude is observed between Figures S8.1a and S8.1b, and between S8.1c and S8.1d. Concerning uplift rates, especially for Earth model A, in which the viscosity at the right side of the transition is relatively low, a difference can be observed at the Wilkes Subglacial Basin. For Earth model B, the uplift rates above the RSE seem to be smaller, but not much.

8: ANET/POLENET GPS Solution

The ANET/POLENET GPS solution (Wilson et al., 2017-in prep) is provided to us by Stephanie Konfal and the Byrd Polar and Climate Research Center of Ohio State University. The site names, latitude and longitude, the East and North components, and the uncertainties of the horizontal GPS observations of 24 sites surrounding the RSE are shown in Table S8.1. These correspond with the velocities in Figure 4c in the main article.

<table>
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<th>Site Name</th>
<th>Latitude (deg)</th>
<th>Longitude (deg)</th>
<th>East (mm/yr)</th>
<th>North (mm/yr)</th>
<th>Se East (mm/yr)</th>
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Table S8.1 | Horizontal GPS measurements of ANET/POLENET at 24 sites surrounding the RSE

The sites in East Antarctica that are used to remove plate rotation from the GPS and to correct the rotational portion of the GIA predictions are VESL, SYOG, MAWI, DAVI, CASI, ABOA, BURI, LWN0 and WHN0. The values at these sites are shown in Table S8.2.
<table>
<thead>
<tr>
<th>Site Name</th>
<th>Latitude (deg)</th>
<th>Longitude (deg)</th>
<th>East (mm/yr)</th>
<th>North (mm/yr)</th>
<th>Se East (mm/yr)</th>
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Table S8.2 | Horizontal GPS measurements of ANET/POLENET at 9 reference sites along the East Antarctic Coast.

We find for the predictions of all our Earth models that there is a consistent misfit with the GPS measurements at the FALL GPS site. It is probable that this is caused by a mismodeling of the ice loading and unloading at this location or a local lateral variation in viscosity. Because of the systematic misfit with the measurements at FALL, this site is excluded from the analyses in the main article.
Supplementary Figure 1: Nodal Velocities at Layer Boundaries

Predictions of the deformation rates at nodes of layer boundaries in the lithosphere (0 and 50 km depth) and UM (90, 200, 325 and 400 km depth) are shown for a homogeneous mantle viscosity of (a) $10^{19}$ and (b) $10^{20}$ Pa s, at 12 and 8 ka BP. The plots at 12 and 8 ka BP show that for a viscosity of $10^{19}$ Pa s, the lithospheric velocities have a stronger outward direction relative to the mantle flow than for $10^{20}$ Pa s. A mantle with a viscosity of $10^{19}$ Pa s relaxes faster than a mantle with a viscosity of $10^{20}$ Pa s. Therefore, for a viscosity of $10^{19}$ Pa s the outward motion of the lithosphere becomes dominant for the direction of the horizontal velocities at the surface slightly earlier than for a viscosity of $10^{20}$ Pa s. Additionally, it can be seen that the nodes at layer boundaries in the mantle have larger deformation rates for a lower viscosity.
The schematic lateral viscosity transition between East and West Antarctica is selected based on a mantle temperature model by (Pappa et al., 2017). On the temperature plots at the right hand side, the transition with constant longitudes of 155°E and 50°W is superposed on temperature maps at 100 and 150 km depth. The left hand side of the figure illustrates that viscosity values at the West and East Antarctic side of the transition are varied in two layer of the UM.

The simplified transition approximates the temperature contrast in the temperature model in the figure reasonably well close to the RSE, but is less realistic close to the South Pole and Ronne Ice Shelf. Therefore, only the GPS sites surrounding the RSE are considered.
Supplementary Figure 3: W12 Deglaciation History
The W12 deglaciation history is shown over time at multiple timesteps between 20 and 0 ka BP. Additionally, the isolated ice dome at the Ross Sea that was used to investigate the dependency of horizontal velocities on mantle viscosity, is shown at 20 ka BP.
Supplementary Figure 4: Horizontal Velocities Through Time for Ice at the RSE on a 1D Earth

In this figure the deformation rates due to the ice dome at the Ross Sea on a homogeneous Earth with a viscosity of $10^{19}$ Pa s are shown at multiple time steps (the deformation rates at present are the same as in Figure 3a in the main article). The figure shows that for $10^{19}$ Pa s, until 8 ka BP the inward mantle flow dominates the horizontal velocities at the surface. After that, the direction of the velocities changes. The vertical and horizontal deformation rate patterns at 4 ka BP at the Southern part of the former ice dome are similar to those at present-day for $10^{20}$ Pa s, as was discussed in the main article.
Supplementary Figure 5: Horizontal Velocities Through Time for Ice at the RSE on a 3D Earth

In this figure the deformation rates due to the ice dome at the Ross Sea on a laterally heterogeneous Earth are shown at multiple time steps (the deformation rates at present are the same as in Figure 3e in the main article). A lateral transition in viscosity runs from 155°E to 50°W, from $10^{19}$ to $10^{23}$ Pa s. At both sides of the viscosity transition the viscosity is radially homogeneous. Compared to Supplementary Figure 4, a clear difference is observed in the direction of horizontal velocities. At 19 ka BP, for a homogeneous Earth the mantle clearly flows towards the ice dome throughout the whole of Antarctica. For the heterogeneous case only in West Antarctica horizontal velocities are significant, and point towards the viscosity transition. The stiff mantle underneath East Antarctica blocks the horizontal mantle flow during the ice loading, as was mentioned in the main article. As soon as the ice unloading starts, only in West Antarctica the mantle flows back inwards.
Supplementary Table 1: Baseline Layering 3D GIA Model
Baseline layer definitions and material properties for the Earth models of the 3D GIA model. These settings are used throughout the article unless stated otherwise. Density and Young’s modulus are obtained through volume-averaging of the values in PREM for the corresponding depths of the layer boundaries.

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<th>(v) (-)</th>
<th>(g_0) (m/s(^2))</th>
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**Supplementary Table 2: Viscosity of Earth Models and Rms Misfits of Predictions with GPS**

Viscosity variations of the schematic lateral transition between West and East Antarctica between 50 and 200 and 200 and 400 km depths at each side of the transition are shown. The resulting rms misfit of predictions with GPS observations can be seen in column 4, and the misfit excluding FALL in column 5. Rms misfits below 0.8 mm are shown in bold. For the corresponding Earth models the direction at the different regions surrounding the RSE are inspected. Of these Earth models, for four models the direction of the predictions with respect to the ice load agrees with that of the GPS at all regions. The Earth model with the lowest rms misfit that also fulfils the requirement that the direction of the predictions agree with the GPS observations, is Earth model 9.

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<th>WA Viscosity 50-200, 200-400 km (log η)</th>
<th>EA Viscosity 50-200, 200-400 km (log η)</th>
<th>RMS Misfit (mm/yr)</th>
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Supplementary References


CONCLUSIONS AND RECOMMENDATIONS

In this chapter, the most important conclusions are given of the work that was done in this thesis. In Section 5.1 the research questions that were posed in Chapter 1 will be answered based on the results presented in the article. In Section 5.2, it will be discussed how to look at the answer to the main research question in a critical way. Recommendations are formulated for future work. Together, these sections form the main conclusion to this research. Finally, the implications of the results will be explained in Section 5.3.

5.1. CONCLUSIONS

The research questions posed in the introduction are repeated for convenience. The sub-questions are answered directly underneath the questions, after which the main question is answered below.

1. To what extent can horizontal GPS observations at the Ross Sea Embayment constrain the lateral transition in mantle viscosity between West and East Antarctica in forward models of Glacial Isostatic Adjustment?

   (a) How does the direction of GIA-induced horizontal deformation rates depend on mantle viscosity?

   It was found that the direction of present-day horizontal velocities on the surface of the Earth with respect to the ice load center can be reversed, depending on the magnitude of the viscosity in the UM. For viscosities of $10^{20}$ Pa s and lower, horizontal velocities point away from the (former) ice centre, while for viscosities of $10^{21}$ Pa s and higher velocities point inwards. These results are also obtained with a more realistic deglaciation history at the RSE. The velocities at nodes at layer boundaries in the lithosphere and UM show that the direction dependency is caused by the opposite motions of the inward mantle flow and outward movement of the lithosphere. Depending on the magnitude of the mantle viscosity, either the shear exerted on the lithosphere by the mantle flow is strong enough to cause inward horizontal velocities at the surface, or is not strong enough, in which case outward velocities are predicted. The finding that the direction of horizontal velocities can be reversed by changing mantle viscosity was not reported in the literature before, and has important implications on the use of horizontal GPS observations as a constraint of GIA models.
(b) What are the main differences and similarities between the predictions of horizontal deformation rates from GIA models with and without a lateral variation in mantle viscosity West of the Ross Sea Embayment?

The W12 deglaciation history was isolated at the RSE and applied on the surface of a series of 1D Earth models, and on Earth models with a schematic lateral viscosity transition. For the homogeneous Earth models the same dependency of the direction of the horizontal velocities on viscosity was found as for an axisymmetric ice cap. When a lateral transition in viscosity between West and East Antarctica was applied, it was found that the direction of the horizontal velocities predicted at either side of the transition can have different directions. With respect to the viscosity transition, the direction of the horizontal velocities in West Antarctica and in East Antarctica had the same dependency on viscosity as the horizontal velocities in the whole of Antarctica for a homogeneous Earth model. For example, for an Earth model with a homogeneous viscosity of $10^{19}$ Pa s, horizontal velocities point away from the ice, and for a homogeneous viscosity of $10^{21}$ horizontal velocities point towards the ice. Thus, for a heterogeneous Earth model with a viscosity transition from $10^{19}$ Pa s in West Antarctica to $10^{23}$ Pa s in East Antarctica, horizontal velocities point away from the viscosity transition in West Antarctica and towards the viscosity transition in East Antarctica.

An important feature in the behaviour of horizontal velocities for Earth models with a lateral transition, is that the horizontal velocities point away from or towards the viscosity transition rather than the principle ice load. This can be understood as follows: the stiff mantle underneath East Antarctica blocks the outward mantle flow from underneath the ice towards East Antarctica during the ice loading. This flow cannot be mimicked with 1D Earth models, since it was shown that for a homogeneous mantle the horizontal flow pattern East and West of the ice load do not differ in direction but are approximately symmetric around the load centre. When the difference in magnitude between the viscosities East and West of the transition is decreased, the mantle underneath East Antarctica will accommodate more of the stress due to the ice loading, leading to relatively larger deformation rates at the surface at present-day. Concluding, the horizontal flow pattern resulting from Earth models with a lateral viscosity transition is fundamentally different from that from homogeneous Earth models.

(c) What viscosity values are preferable at both sides of the viscosity transition between West and East Antarctica when comparing the predictions of horizontal deformation rates with GPS observations at the Ross Sea Embayment?

In the introduction in Chapter 1 it was shown that horizontal GPS measurements of ANET point inward to the ice load West of the RSE, and outward from the ice East of the RSE. As the predictions of horizontal velocities for a 1D Earth model are approximately symmetric in direction around the ice load, it is likely that the horizontal ANET GPS velocities require a lateral transition in viscosity West of the RSE. In order to obtain preferred viscosity values at either site of this transition, predictions for 24 different Earth models were computed varying viscosity in two layers of the UM. The four Earth models with the lowest rms misfits and for which also the direction of the horizontal velocities with respect to the ice load agrees with that of the GPS at all regions surrounding the RSE, constrain a viscosity range of $10^{18}$ to $10^{19}$ Pa s from 50 to 200 km and $10^{19}$ Pa s from 200 to 400 km for West Antarctica, and of $10^{22}$ Pa s from 50 to 200 km and $10^{21}$ to $10^{22}$ Pa s from 200 to 400 km for East Antarctica. The best fitting Earth model has a viscosity of $10^{19}$ Pa s from 50 to 400 km in West Antarctica, and of $10^{22}$ Pa s from 50 to 200 km and $10^{21}$ Pa s from 200 to 400 km in East Antarctica. The misfit of the best fitting Earth model is 0.718 mm/yr.
What is the effect of variation of lower mantle viscosity and lithospheric thickness on the fit of the predictions of the preferred Earth model with the horizontal GPS measurements at the Ross Sea Embayment?

The fit of the best Earth model was assessed for four different combinations of LM viscosity from 400 to 670 km and from 670 km and lower. The standard values used in this study were $5 \cdot 10^{20}$ Pa s and $10^{22}$ Pa s respectively. The LM viscosity was varied using the preferred values of (1) W12: $10^{21}$ and $10^{22}$ Pa s, (2) of ICE-6G_C: $0.5 \cdot 10^{21}$ and $3 \cdot 10^{21}$ Pa s, and (3) the untested combination: $10^{21}$ and $3 \cdot 10^{21}$ Pa s, respectively. The combination of LM values for which the resulting rms misfit differs the most compared to that of the best fitting Earth model with the standard LM viscosity values, is combination 3. The misfit of that model is 0.687 mm/yr, corresponding to a difference of 0.031 mm/yr with the initial model. This difference is relatively small and would not make a significant impact on the range of preferred UM viscosities that was reported above. However, if LM viscosity is varied outside of the boundaries formed by the preferred values of W12 and ICE-6G_C, LM viscosity might change the results.

Instead of the original lithospheric thickness of 50 km, also a lithospheric thickness of 70 km and a transition of 50 km in West Antarctica to 70 km in East Antarctica were tested. Based on the results for a lithospheric thickness of 70 km which were not yet transformed to the GPS frame, it can be seen that a higher lithospheric thickness leads to lower velocities in both West and East Antarctica. This is most likely due to a different response of the lithosphere to the shear stress that is exerted by the mantle flow. A thicker lithosphere will be more resistant to the shear stress (Mitrovica et al., 1994). However, since also the estimation of a common rotational signal of the predictions is affected, the general effect of varying lithospheric thickness is difficult to interpret. The transformed predictions for a lithosphere of 70 km are bigger than for 50 km, and the resulting misfit was found to improve to 0.670 mm/yr. For a transition from 50 to 70 km, the predicted velocities at the RSE are smaller and are rotated slightly to the North. The resulting misfit was decreased to 0.746 mm/yr.

I. To what extent can horizontal GPS observations at the Ross Sea Embayment constrain the lateral transition in mantle viscosity between West and East Antarctica in forward models of Glacial Isostatic Adjustment?

It was found that depending on mantle viscosity, the direction of horizontal velocities with respect to the ice load can be reversed. Predictions of horizontal velocities point outward for viscosities of $10^{20}$ Pa s and lower, and inward for $10^{21}$ Pa s and higher. With a schematic lateral transition in viscosity West of the RSE, the horizontal flow around the ice is affected in a way which cannot be mimicked with any 1D rheology. The inward direction of the ANET/POLENET GPS velocities West and outward East of the RSE, reflect this flow. Thus, it is concluded that the ANET horizontal GPS observations surrounding the RSE require a lateral transition in viscosity between West and East Antarctica, and that the observations can be used to constrain the lateral transition at least near the RSE.

Comparing the GPS measurements at the RSE with predictions of a suite of Earth models with a schematic lateral transition in viscosity, a range of preferred UM viscosities was found of $10^{18}$ and $10^{19}$ Pa s in West Antarctica and $10^{21}$ to $10^{22}$ Pa s in East Antarctica. Four of our best fitting Earth model, the inward or outward directions with respect to the ice load of the predictions agree with those of the horizontal GPS velocities surrounding the RSE. We find a rms misfit of 0.718 mm/yr, which can be improved up to 0.641 mm/yr when the direction of the schematic
viscosity transition is changed to a constant longitude that agrees with the shear wave velocity contrast in models of An et al. (2015) and Heeszel et al. (2016). A good fit is achieved especially in the central and lower-Northern regions, which has not been accomplished with 1D Earth models before. However, further away, the predictions are too small or point too much to the North. This is most likely due to uncertainties in deglaciation history, and due to the simplifications in our Earth and GIA model.

It is concluded that due to the dependency of horizontal velocities on mantle viscosity that was found, the horizontal GPS observations at the RSE can be used to constrain the lateral transition in mantle viscosity between West and East Antarctica West of the RSE. The ANET observations form a strong evidence that the lateral contrast in shear wave velocities indeed translates to a transition in viscosity. This has important implications for GIA studies and other fields, which will be discussed in Section 5.3. However, only the W12 deglaciation history for Antarctica was used in this study. Also, a constant lithospheric thickness and incompressibility were assumed, and postseismic velocities were neglected. These aspects should be further investigated, and recommendations to adress them will be formulated in the following section.

5.2. RECOMMENDATIONS

The most important aspects of this work that deserve further attention will be highlighted. Recommendations are formulated for future work on these topics.

1. The community should work towards a more accurate model of a 3D temperature distribution in the UM.
Discrepancies between 3D viscosity distributions derived from different temperature models can propagate to differences in predicted horizontal velocities of up to 0.88 mm/yr. The conversion of temperature from shear wave velocity models knows first-order uncertainties if variations in chemical composition are not considered. Additionally, the conversion of temperature to viscosity introduces additional variables with large uncertainty. Instead of deriving a 3D viscosity distribution from temperature through flow laws for olivine, a schematic lateral viscosity transition was adopted in this study with two different viscosity profiles in the UM at either side of the Earth, in order to constrain viscosity directly. The simplifications in the methodology that this required result in a limited quality of the the resulting fit of the predictions with GPS surrounding the RSE. Additionally, the preferred viscosities found for West and East Antarctica will lose meaning further away from the RSE.

It would be beneficial to use a fully 3D rheology derived from a 3D UM temperature model, when more reliable temperature models become available. The uncertainty in the derivation of temperature from shear-wave velocities could be decreased by including petrological models and constraints from gravity measurements, or by using surface heat flow models if more observations become available in Antarctica. With a fully 3D rheology then also a realistic lithospheric thickness can be can be tuned simultaneously with UM viscosity. The preferred UM viscosities in this report were found assuming a constant lithospheric thickness of 50 km. As lithospheric thickness was shown to significantly affect horizontal velocities, the average viscosities found at either side of the transition might change for different lithospheric thicknesses. Also, the common rotation-like signal in the GIA predictions that should be corrected for before comparison with GPS, can be better estimated when the viscosity underneath the reference sites in East Antarctica is more accurately represented.

2. If possible, the computational time of 3D GIA Earth models should be reduced.
The large computational time of the 3D coupled FEM-Laplace model is a severe constraint
on this study, as it requires a constant trade-off between the quantity and quality of computations. For example, a large computational time limits the search in a wide parameter space, but computing less iterations or using a larger integration error tolerance will have adverse effects on the accuracy of the predictions. If computational effort could be reduced, viscosity values could be varied within less than one order of magnitude difference, more deglaciation histories could be added, and variations of lithospheric thickness and LM viscosity could be included. Computational time can be decreased by allocating more processor cores. Also, the computational effort could be decreased by using less distinctive layers and/or less elements especially in lower layers, which should be carefully traded off with the accuracy of the predictions. A variable grid size could be used to have a high resolution in specific areas of interest, and a lower resolution elsewhere.

3. The effect of the resolution of the 3D GIA model on horizontal velocities should be studied.
   As the number of elements at the surface of the 3D GIA model that was used in this thesis was hardcoded, it was not attempted to change the resolution in this work. However, with a benchmark of the 3D and the axisymmetric GIA model, it was found that especially close to the ice margin and the forebulge, a model with a too coarse grid spacing at the surface cannot capture small wavelength effects. Additionally, a small difference between the results of the 3D and axisymmetric models was found at every colatitude, which is most likely caused by the difference in resolution. It is recommended that the effect of an improved resolution on the prediction of horizontal velocities following from 3D GIA models is studied in order to investigate whether smaller elements are required for an accurate computation.

4. Future studies that use predictions of horizontal velocities from forward GIA models should include compressibility.
   It was shown in a number of studies (A et al., 2013, Mitrovica et al., 1994, Tanaka et al., 2011) that compressibility in GIA models has a significant effect on predictions of horizontal velocities. In this study an incompressible Earth was assumed, because density perturbations resulting from compressibility can lead to numerical errors in the model (Wu (2004) and personal communication with Patrick Wu). It is believed that primarily the magnitude of the horizontal velocities is affected rather than their direction. To still include compressibility while avoiding the associated numerical instabilities, only material compressibility can be assumed (i.e. only changing the Poisson’s ratio and Young’s modulus) (Wu, 2004), but this is physically incorrect without including density perturbations. Nevertheless, when material compressibility was included for the best fitting Earth model, it was indeed found that only the magnitude of the velocities is affected. However, the change in rms misfit with the GPS observations is only small. It is recommended for future studies to investigate the effect of density perturbations on horizontal velocity predictions in the RSE as well, and include compressibility in a consistent manner when horizontal velocities are compared to GPS measurements. This is limited in models following the approach of Wu (2004) (correspondence with author). A possible circumvention is to adapt the elastic rigidity of an incompressible model so that it approximates the flexural rigidity of a compressible model, when studying effects of wavelengths shorter than 8000 km (Tanaka et al., 2011).
5. **The effect of uncertainties of deglaciation histories in Antarctica on the results should be investigated.**

The uncertainties in deglaciation histories of Antarctica remain one of the biggest error sources of forward GIA models. Only the W12 (Whitehouse et al., 2012a) deglaciation history was used to find preferred viscosity values in this study, as the ice in ICE-6G_C (Argus et al., 2014) is tuned to fit observations in combination with a 1D Earth model, and IJ05_R2 (Ivins et al., 2013) is based on a series of circular ice discs which are inconsistent with glacial physics. Future work should assess the robustness of the results in this thesis given the uncertainties in deglaciation histories for Antarctica. The total mass loss of the AIS in the ICE-6G_C model is much higher than that in W12 and IJ05_R2. Additionally, ICE-6G_C includes rapid ice loss in early timesteps compared to W12. The location of maximum ice loss over the Ross Sea in ICE-6G_C is located more to the East of the RSE compared to that in W12 and IJ05_R2. With respect to the West coast of the Ross Sea however, the spatial extent of the ice sheet is similar for all models. There is reasonable coverage of geophysical evidence of past grounding lines of the ice sheet at the Ross Sea (Whitehouse et al., 2012a).

The differences in ice loading and unloading will affect preferred viscosities that are found for the RSE. Thus, it is recommended to investigate how sensitive the outcomes of this thesis are to different ice models. First of all, even though it is likely inconsistent to apply an ice model which is tied to a 1D Earth model, to a 3D Earth model, the ICE-6G_C and IJ05_R2 ice models can be applied on a series of representative 3D Earth models in order to get a first idea of how the predictions change. Additionally, it can be studied how the predictions differ when modifications are made to the W12 deglaciation history. The total ice melt in the W12 model can be increased with a constant factor in order to simulate the effect of higher mass loss of the AIS that is predicted by the ICE-6G_C model. Also the timing of that mass loss can be modified. Furthermore, the ice loss can be increased locally in the East of the RSE in order to simulate the effect of the maximum ice loss which is located more towards the East of the RSE and thus further away from the lateral transition in viscosity. For all of the three deglaciation histories that were mentioned here the extent of the ice sheet with respect to the West of the RSE and the lateral transition in viscosity is similar. It is therefore expected that regardless of the deglaciation history that is used, the relatively stiff mantle underneath East Antarctica will have a pronounced effect on the horizontal velocities West of the RSE, as it blocks the outward mantle flow during the ice build-up.
5.3. IMPLICATIONS

Recent GIA studies found low mantle viscosities beneath the AP \((\text{AP}, < 2 \cdot 10^{18})\) (Nield et al., 2014) and the Amundsen Sea Sector (Barletta et al., 2017). Seismic studies (e.g. (An et al., 2015, Heeszel et al., 2016)) show a lateral contrast in shear wave velocities between West and East Antarctica, which might indicate a transition from low viscosity beneath West Antarctica to higher viscosity in East Antarctica. Our results confirm that the contrast in shear wave velocities between West and East Antarctica indeed translates to a viscosity transition in the UM of at least two orders of magnitude. This shows that regardless of variations in chemical composition, the contrast in shear wave velocities is at least partially caused by a contrast in temperature. Our conclusion contradicts the finding of Argus et al. (2014) that there is no need for lateral variation in mantle viscosity to fit the horizontal data in Antarctica well. However, as was mentioned before, their method of comparing predictions with GPS is circular (emphasized also by King et al. (2016)), as it is assumed in their approach of correcting for plate motion with sites in East Antarctica that the viscosity underneath the East Antarctic is the same as in West Antarctica, which is most likely incorrect.

The low viscosity that was found for West Antarctica is in line with results for other regions in West Antarctica such as the AP (Nield et al., 2014) and results in lower uplift rates over the Ross Sea than predicted with the preferred models of W12 and ICE-6G_C. This agrees with the mean uplift rates of 3 inverse models of GIA (Martín-Español et al., 2016). The evidence for a significant lateral transition in viscosity between West and East Antarctica implies that 1D GIA models are likely unrealistic when used for the whole of Antarctica, as they overestimate the viscosity in West Antarctica. Thus, estimates of the total ice mass loss and the evolution of the AIS following from GRACE mass balance estimates that are corrected for GIA with 1D models are indeed biased, as was suggested by van der Wal et al. (2015). A predominantly low viscous mantle beneath West Antarctica also implies a higher sensitivity to late-Holocene surface mass variability. Additionally, as viscosity drives plate tectonics, lateral viscosity variations could influence stresses in the lithosphere. In order to compute accurate predictions of postseismic velocities, viscoelastic models of postseismic deformation in Antarctica (e.g. (King and Santamaría-Gómez, 2016)) should include the lateral transition in viscosity between West and East Antarctica that was confirmed here.

When more certain 3D mantle rheologies underneath Antarctica become available, comprehensive studies should be done that constrain GIA with a suite of observations that also includes horizontal GPS velocities. Due to the dependency of the magnitude and especially the direction of horizontal velocities on viscosity that was found in this study, it is likely that not only at the RSE but also in other regions with potentially large lateral transitions in viscosity (e.g. North America (Godey et al., 2004, Yuan et al., 2011)), horizontal GPS data can be used to constrain GIA when 3D GIA models are used. More accurate predictions of the horizontal deformations due to GIA can also help to isolate horizontal velocities due to ongoing postseismic deformations which were not considered in this study. In the long term, an improved understanding of lateral heterogeneities in Earth structure is thought to be crucial to improvements in the uncertainties of deglaciation histories, as then deglaciation histories could be developed simultaneously with a preferred 3D Earth model.


BIBLIOGRAPHY


