GRACE constraints on Earth rheology of the Barents Sea and Fennoscandia

Response to Reviewers

We thank the reviewers for the comments provided, we think they helped improving the quality of the manuscript and clarify some relevant points. Below, we address the reviewers' comments (in blue). A new version of the manuscript with changes indicated follows. References to pages and lines (page,L line) refer to this version of the manuscript, lines and pages mentioned in the reviewer's comments correspond to the first version of the manuscript. Figures in this response are referenced as R.
Note that the individual response to each reviewer can also be found in the Interactive Discussion Forum.

Reviewer 1:

Major comments

Reviewer 1 express his concerns about some of the processing techniques used in this study. Below we address these comments:

My primary concern is about the substantial signal lost through the filtering and processing of the measurements and low resolution of the modeling. The limitation of maximum order number 60, which yields a minimum resolution of approximately 300 km, or one quarter of the linear extent of the Barents Sea. It is also cuts off a significant potion of the power of the authors' bandpass filter ranges. Thus the shape of the bandpass filter dominates the shape of the processed and modeled measurements.

The cut-off degree and filters used in GRACE data are equally applied to the simulated GIA signal (7, L22). This way the comparison of GRACE and simulated gravity rates is consistent

The effect of cutting the GIA signal at degree 60 is illustrated in Figure R1 and R2. Figure R1 shows the GIA signal obtained with the ICE-5G model cut at different degrees, Figure R2 gives the maximum gravity disturbance rate obtained in the Barents Sea for different cut-off degrees. Both plots evidence that the GIA signal does not have a high content of high degree harmonics and therefore little signal is lost.

The filters used to process the data are carefully chosen following the work of Root et al. 2015 (see supplementary data). A low pass filter is needed to filter out small wavelength noise. We vary the filter halfwidth between 200km and 300km. Below 200km noise becomes dominant and above 300km the positive signal located in the Barents Sea is very small (see Figure 3). The half-width of the high-pass filter is chosen using Figure 1 from Root et al. 2015 supplementary material. They use a synthetic GIA signal to show the effect of using a high-pass filter. From that figure it is concluded that a high pass filter between 500 and 700 km keeps the GIA signal while removing other long-wave signals.

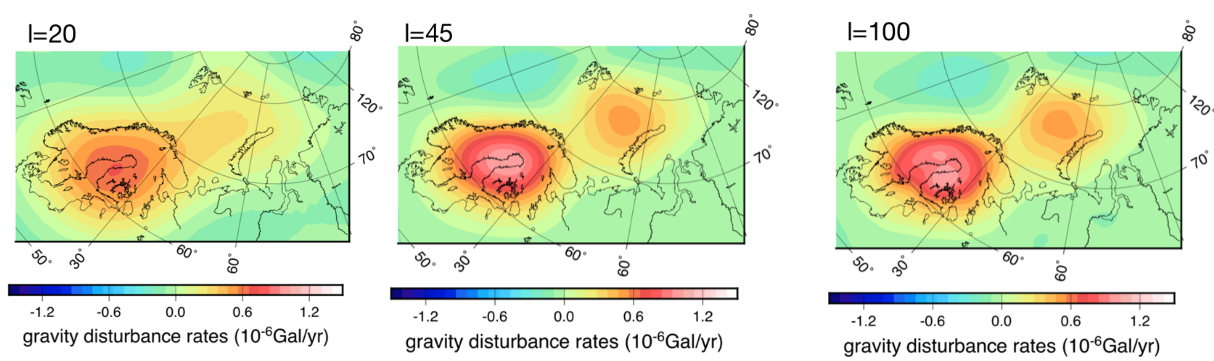

Figure R1: *Simulated GIA signal cut at different degrees*

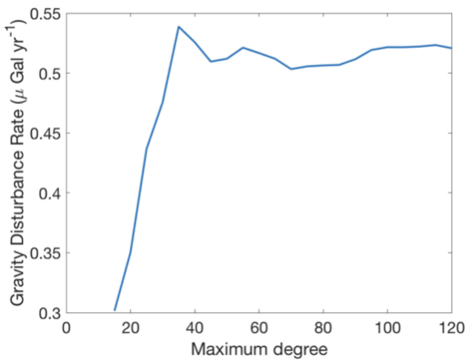

Figure R2: *Maximum gravity disturbance rate as function of cut-off degree*

This filtering occurs after a series of processing steps to extract the LGM signal. The GRACE measurements are processed one way to estimate the current mass loss off the archipelagos, another to estimate the ocean signal, and a final way to estimate the response to LGM deglaciation in the Barents. So, while I really appreciate the attempt to quantify all of the sources of error, the assumption that they are uncorrelated (page 5, line 12) requires further explanation. I would similarly like elaboration of the effect the GIA model chosen has on the estimate of mass loss (page 4, line 33).

It is true that GRACE is used to (1) recover the LGM signal, (2) obtain the mass loss changes in the islands of the Arctic archipelago and (3) in a smaller degree in the ECCO model which makes it possible for the errors to be correlated.

We start by addressing the use of GRACE to estimate mass changes in the Arctic Archipelago. The "circularity" of the problem is explicitly mentioned in page 4, L32. We use an ensemble of ice sheet models and solid Earth rheologies to estimate the uncertainty in mass changes. Table 1 now gives the different combinations of ice-sheet and rheology models used to correct GRACE estimated mass changes and the obtained mass changes. Four different ice sheet models and three different rheologies are used

The ice sheet models correspond to two runs of the Glacial System Model for Northern Europe, the ICE-5G and the W12 models. As for the rheology we use the VM5a model and two models with an upper and lower mantle viscosities of $16*10^{20}$ and $512*10^{20}$ and $10*10^{20}$ Pas and $100*10^{20}$ Pas. We see that a weaker mantle leads to higher mass loss rates for the islands of the Artic Archipelago. We add this table as well as an explanation on the effect of the GIA model on the mass loss estimations in the tex.

GRACE data is not used in the creation of the OMCT ocean model but it is used in the ECCO ocean model. However, this is only 1 of the 40 data sets that are used to constrain the dynamic MITgcm ocean model and as shown in ECCO's documentation (https://ecco.jpl.nasa.gov/drive/files/Version4/Release3/doc/v4r3_estimation_synopsis.pdf, Figure 2), GRACE is one of the worst fitted observations. This fact is also evident in the results of Yu et al. 2018 who compare GRACE derived bottom pressure anomalies to the ECCO ocean model for the Argentine Gyre. We do not use the ECCO ocean model in our estimate itself, but only as a proxy of how much error we might expect from uncertainty in the ocean signal. Given the weak contribution of GRACE to the final output we think this is appropriate.

The correlation of GRACE's measurement error and that from ice loss estimates cannot be ruled out. The error in ice loss estimates have two components (1) GRACE's accuracy error and (2) error due to uncertainty in the GIA model. While the second is not correlated with GRACE's measurement error, it is true that the first can be. In light of this discussion we decide to add the following discussion after equation (1).

"The assumption that errors are uncorrelated requires further discussion. GRACE data is assimilated in the ECCO ocean model. However, GRACE is only one of the 40 data sets used in the inversion process and the final product does not fit GRACE data well (Yu et al. 2018). Therefore there will be only a weak correlation with the GRACE data used in our estimation. Correlation between land surface hydrology models and present-day ice melt is not expected, because hydrology models have little skill in predicting trends and do not model areas of permanent snow. Finally, ice loss changes errors (σ_{ice}) arise due to uncertainty in the GIA model and GRACE measurement error, we cannot rule out that the second error component might be correlated with σ_{GRACE}. "

In light of this concern, I would ask the authors to: 1) further quantify the effects of their processing technique for this area. In particular, by adding more discussion of the technique for idealized measurements in the context of the Barents; and

2) consider acknowledging the processed nature of these results by referring to them as “estimated gravity rates” rather than “observed gravity rates.” I feel this is particularly important when the authors substitute the phrase “observed gravity rate” for the estimated maximum gravity rate (e.g., page 6 - line 31).

The processing techniques were detailed in the supplementary material of Root et al. (2015), their effect for an idealised measurement in the Barents Sea is shown there.

Following the suggestions of the reviewer we:

(1) Refer to the Supplementary material of Root et al. 2015 for a detailed explanation of how our processing affects an idealised signal (4,L5).

(2) Include a Table showing the estimated mass loss changes obtained using different GIA models.

(3) Use the term estimated gravity rates instead of observed (8,L14).

The argument additionally suffers from another small, but troubling, circularity. The ensemble of ice models was chosen to represent two classes: empirical ice sheets developed using GIA observables and ice sheets developed from independent, process based models. However, all of the models are actually calibrated, in some way or another, to GIA observables with an implicit dependence on the assumed viscosity structure. For instance, the Tarasov samples are drawn from a distribution trained on GIA observables using the Peltier VM5a rheology. If the authors could comment on this bias and how that might account for the reference model being very near the best fit valley in all figures but the Siegert and Dowdeswell 2004 model, which is the only one to prefer an anomalously high viscosity, most likely because of it’s earlier ice-free time.

Although the problem is certainly there, the S04, but also the UiT model are not fitted to GIA observations. In the main text we distinguish between two different types of ice models, (1) those that do not include ice sheet physics (ICE-5G, ICE-6G) and are entirely based on GIA observations and (2) those that incorporate ice sheet physics. However, as pointed out some of the ice models in the second subset do also include GIA observations. Tarasov’s models are calibrated using the fit to RSL curves and uplift rates obtained with the VM5a model, however the calibration accounts for spread in the decay times due to uncertainty in the viscosity model and is not tuned to a single viscosity model as much as ICE-xG models are (L. Tarasov personal communication). For the UiT model a simple hydrostatic model is used to account for ice-elevation feedbacks, but the model fit to GIA observables is assessed a-posteriori (Patton et al. 2016). We include this additional information extending our description of the ice models in Section 2.2.

Minor comments

page-line 4-33: “However, the GIA” It is not obvious that this should be so.

We tried to clarify this point by adding the individual mass loss estimates for each of GIA model in Table 1. Moreover, we clarify this point by showing that uncertainty the GIA model used to recover the mass changes is of the same order of magnitude as GRACE’s formal error (5,L7).

8-19: citing the nchiˆ2 might make this point clearer. It is hard to tell that S04 is significantly worse than, say UiT, from figure 3.

We state that the model performs worse than the T1, T2 and T3 (which is clear from figure 3). Later on this is evidenced with the \chi^2. We also add some new discussion on the fit of the S04 model for different Earth models in Section 3.2 (9,L32).

10-32 might include “explicitly” in “not explicitly tied to a viscosity model”

We follow the suggestion

Figure 5 and Figure 6 - Could you note with a symbol the reference model and the best fit model in each of these plots?

For each lithospheric thickness we indicate the best fitting model with a red line and the reference model with a red dot.

page-line

1-2: in-> to “insight to the” Done

1-4: Split sentence Done

1-5 remove “a” in “a GIA models” Done

1-6 “is not negligible” and “should be taken into account” are redundant Done

1-16 Inconsistent use of “gravity disturbance rate” and “gravity rate” We use gravity disturbance rates until it is stated that the term gravity rates will be used instead in 4-L9.

3-7 missing word in “while best fitting models uplift rate measurements” Done

3-22 missing “and” in “GIA, and (post-) seismic” Done

5-4 GAB undefined GAB is not an acronym. We add a reference to Flechtner et al. 2015 where the GAB files are defined.

5-9 “respectively” has no antecedent. Consider “both the OMCT and ECCO ocean models” We follow the suggestion

5-15 “while when” is difficult to parse We rephrase the sentence.

5-22 missing word in “This still allows” We rephrase accordingly.

5-23 correct citation parenthesis Done

5-25 missing “the” in “that of the unknown” Done

5-26 remove nested parentheses Done

5-31 missing “the” in “the Earth’s rheology” Done

8-12 I believe deglaciation starts earlier in T2 than in T1, unless I am much mistaken. That is true, we correct the misspelling.

8-31,8-32,9-30 “lower upper mantle viscosity” is pretty cumbersome to read. Consider something like “less viscous upper mantle” We follow the suggestion.

9-9 repeated word “which that” Error corrected

9-12 large->high “high upper mantle viscosity” We follow the suggestion.

9-16 typo“form” Error corrected

Figure 3 and Figure 4 - inconsistent x-axis label We modify the label to ensure consistency.

Reviewer 2:

Major comments

1. My main concern is the lack of discussion of uncertainties in the resulting viscosities. The conclusion that Fennoscandian upper mantle viscosity is a factor of 2 higher than that in the Barents Sea is given with very little discussion on uncertainties:

(a) Most studies would state resulting viscosities and elastic thickness as an interval determined one way or another from the statistics of the inversion process. A differently normalized chiˆ2 range or a variance reduction, for example. On page 8, line 27, a 2 sigma interval is mentioned but not further referred to. The very different chiˆ2 distributions for Barents Sea and Fennoscandia in Figures 5 and 6 make it difficult to asses which parts of the model space is appropriate to compare to one another. In addition, the color scale in the Figures does not enhance the well fitting regions very well, I suggest a scale with a better visible range.

We modify Figure 5 and 6 to give idea of the range of models that perform well. We define a confidence interval following Press et al. 1992 (Chapter 15) (8,L4). We indicate the values within the 95% confidence interval in grey. Additionally, to ease comparison, we indicate in red the best fitting model for each lithospheric thickness which eases the discussion in Section 3.2.

b) At least for the T1-T3 and S04 ice histories, even though the well fitting viscosity range starts at lower viscosities for the Barents Sea than for Fennoscandia, there is significant overlap at higher viscosity in Figure 5. This is less pronounced for ICE-XG and UiT in Figure 6, but is there at thicker elastic thickness. A more well defined range of which models are considered good fits would ease the comparison.

We agree there is an overlap between both regions. We chose to compare the best fitting upper mantle viscosity obtained for each lithospheric thickness in both regions (indicated in red in Figures 5 and 6) to illustrate the difference in upper mantle viscosity and elaborate more on the discussion (11,L11):

“We can infer lateral rheology changes by comparing the optimal Earth rheological parameters obtained for both regions. For each ice deglaciation chronology, we compare the two confidence intervals as well as the best fitting upper-mantle viscosity obtained for each lithospheric thickness. We observe that for the UiT, ICE-5G and ICE-6G model both the confidence interval as well as the best fitting models have a systematically higher upper upper mantle viscosity in the Barents Sea as compared with Fennoscandia. This is also the case for the T1,T2 and T3 models when the best fitting models are compared, although there is an overlap of models of high upper mantle viscosity and thick lithospheres with a good fit in both regions. This systematic difference is likely evidence of lateral variation in Earth rheology. ”

c) The lower bounds on viscosity is very similar for all ice models in the two regions. That is a little odd. Is there some bias somewhere? Such as they having similar Earth models during construction?

The lower bounds obtained for Fennoscandia and the Barents Sea are quite similar. This can be understood by studying the sensitivity of gravity disturbance rates to mass changes during different epochs (Figure R3). As upper mantle viscosity decreases the results are less sensitive to mass changes in the past where differences between ice models are more acute (i.e., Figure 1), making it more difficult to distinguish between ice sheet models. This agree with previous studies that have used gravity data to constrain solid Earth parameters. For instance, Steffen et al. 2010 finds similar lower bounds for upper mantle viscosity using the ICE-5G and the RSES models.

To illustrate this point we add the Figure R3 and add an explain along these lines in section 3.1 (10,L10). “The lower bound obtained with the other ice models is similar as models with a low viscosity have little sensitivity to mass changes during the early deglaciation phase, where differences between ice models are more manifest (Figure 2)”

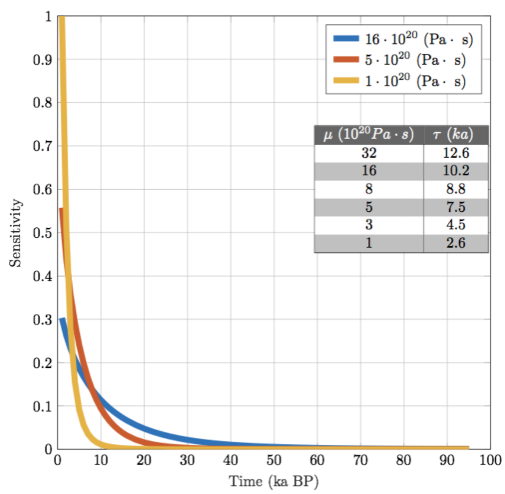

Figure R3: *Gravity disturbance rates sensitivity to mass changes at a given epoch for different upper mantle viscosities. Relaxation time for the different upper mantle viscosities is also given.*

d) Elastic thickness is discussed very briefly in the manuscript. The clear correlations in Figures 5 and 6 between viscosity and elastic thickness should be discussed further. This is different to the results in e.g. Steffen et al. (2010), Root et al. (2015a,2015b). How much are the results for a thicker elastic layer affected by the GRACE filtering process? In addition, there are surely estimates from seismology of the (seismic) thickness of the lithosphere in the Barents and in Fennoscandia. These could perhaps also be used for comparison purposes, even though the measure a slightly different property.

The effect of lithospheric thickness is more evident than in other studies of GIA of the region. We compute the chi2 using the maximum gravity disturbance rates instead of averaging over a region as the shape of the signal in the Barents Sea is difficult to distinguish, for consistency we use the same strategy for Fennoscandia. A thinner lithosphere results in higher gravity rates and can counterbalance a low value of upper mantle viscosity. However, they also lead to narrow region of uplift which does not fit the signal well in Scandinavia. The effect of lithospheric thickness is also evident in Root et al. 2015a. An explanation along these lines is given in (10,L5)

We do not want to draw conclusions on the absolute values for the lithosphere thickness from our misfit plots, because we do not use the spatial pattern of the gravity rate (added in Section 3.1 10,L11) and this is why we compare best fitting upper mantle viscosity for each lithospheric thickness.

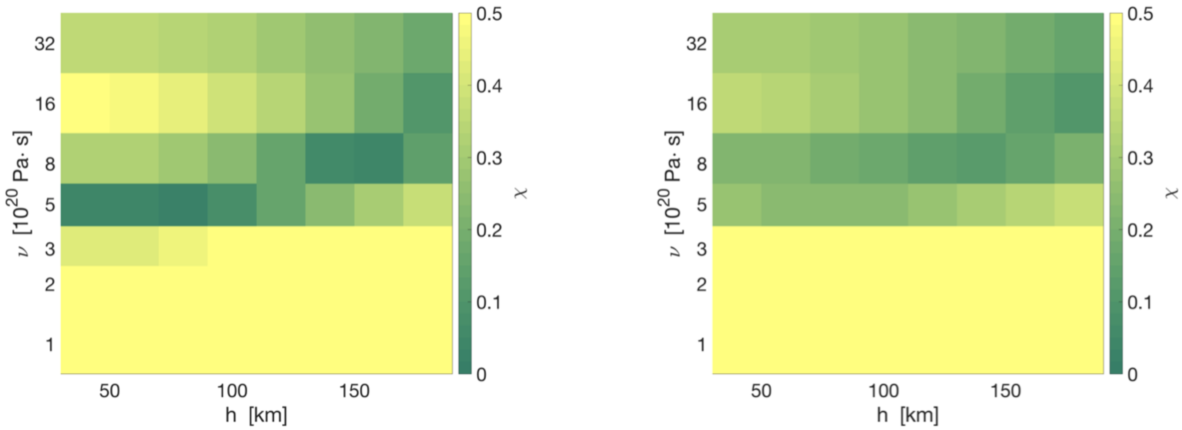

Figure R4: Normalised chi obtained using the maximum gravity rate (left) and by averaging the chi of points in Fennoscandia.

e) When concluding the factor of 2 viscosity difference between Barents Sea and Fennoscandia you should specify at which elastic thickness the comparison is made. If you use different elastic thickness for the different regions that should be explicit. Similarly for the seismic estimates of viscosity difference. These are at the same depth for the Barents Sea and Fennoscandia but to make a fair comparison it would be interesting with estimates of the seismic lithosphere thickness. How much does it matter for the comparison if there are differences in temperature and/or composition in the two regions?

As pointed out at b) the comparison is made for the best fitting upper mantle viscosity obtained for each lithospheric thickness. We add this explanation to the main text.

Differences in temperature are considered in the paper. With regard to composition, we now state that changes in composition are not considered but likely do not play as large a role as temperature, in the upper mantle (11,L28-30).

f) You should compare your inferred viscosity differences to other GIA studies of the Barents Sea area and Fennoscandia. The large number of varying results for Fennoscandia indicate that such a comparison is non-trivial. For the Barents Sea, Root et al. (2015a) indicates 4x10ˆ20 Pa s for the Barents and Auriac et al. (2016) has a very wide range of 2 - 20 x 10ˆ20 Pa s.

We try to put our results in context by comparing them to the results of Auriac et al. (2016) which found a similar bound for upper mantle viscosity in the Barents Sea. For Fennoscandia we refer to the recent work of Simon et al. 2018 that give an overview of different rheology estimates.

Moreover, we are also more cautious in our claim about upper mantle viscosity constrain as other GRACE studies (Root et al. 2015b, Steffen et al. 2014) show that the effect of lower mantle viscosity is not negligible and obtain optimal fits for higher lower mantle viscosities than the one used in this study (6,L21). We add this discussion in section 3.2 and stress this fact in the conclusions.

2. The four error estimates for the GRACE processing seem very appropriate, but I wonder:

(a) Is spherical harmonic degree 60 really enough for this study? And the filtering out of smaller wave-lengths seems to retain only very large scale features, on the order of the whole basin

See response given to the first comment of Reviewer 1.

b) The statement on page 5, line 12, about the independence of the estimates. It seems to me a little strange that the hydrological signal would be uncorrelated with the ice loss signal? Ice mass loss usually means melting, which surely influences the hydrology, both in time and magnitude. Is this not an issue?

See response given to the first comment of Reviewer 1. Moreover, land surface hydrology models have little skill in predicting trends and they do not contain permanent snow or glacier therefore not much correlation with ice melt is expected. We add this explanation below equation 1.

c) The estimate of ice mass loss from GRACE data does indeed seem circular, and a little difficult. A GIA model using GRACE data is used to estimate uncertainties in the GRACE data for GIA applications? On page 4, lines 21-22 the authors state that the current ice mass changes "...partly mask the GIA signal...", but on lines 33- 34 that "... the GIA model used to obtain the mass changes has a small effect in the recovered gravity rate trend...". This seems contradictory to me and need more detailed explanation. Do you use different GRACE filters here to capture the spatially smaller current deglaciations? Also, how are the error bars estimated from the range of ice and Earth models? Do you have a range of reasonable chiˆ2 or something for this error estimate?

We add a new Table showing the mass loss changes obtained using the different GIA models as well as some more details about rheology and ice sheet model affects the results (see response Reviewer 1). What we mean with "... the GIA model used to obtain the mass changes has a small effect in the recovered gravity rate trend.." is that the error introduced by using different GIA is similar as that given by GRACE measurement error (as shown in Table R1). We agree that this statement is confusing and thus we change it by (5,L7): " The error in the derived mass changes due to uncertainty in GIA is similar to GRACE's measurement error".

Minor comments

Including GPS data from Svalbard and northern Norway would have been helpful to constrain the models. Why was this not done?

The focus of this work was on what can gravity tell us about interior structure in this two regions. We agree that adding GPS and even RSL might help to constrain the models but we decided to focus on the gravity signal because the gravity rate is available in the center of the Barents Sea in the region of largest ice thickness where other measurements are absent.

Previous work (e.g., Auriac et al. 2016) have focused on GPS and RSL. To recognise the fact that we don't use these other data sets we add :

"We find that the ICE-5G, ICE-6G and UiT ice sheet models can be reconciled with GRACE observations provided the upper mantle viscosity is lower or the lithosphere thicker than in the VM2 model. The same conclusion is reached in Auriac et al. 2016 using GPS uplift measurements and RSL curves instead of gravity data."

It would be good to have a little bit more information on the ice reconstructions, especially with regard to the used Earth model physics for the non-GIA derived ice models. Do they have appropriate viscoelastic earths, or just simple hydrostatic adjustment, or...? Also, which time period do you use in the models? Just the deglaciation phase? If so, how are the ice sheets ramped up to the last glacial maximum?

We add more information about the different ice chronologies (see Section 2.2). We include both the build-up and deglaciation phase. We add (7,L21): "The only global ice sheet models are the ICE-5G and ICE-6G, for the other ice sheet models we use the ICE-6G ice model outside the EISC. We include the build-up and deglaciation phase of the last glacial cycle". Moreover, we add a new plot showing the deglaciation history in Fennoscandia (see Figure 2), and specify when the ice sheets start to build-up. We also include some information on how isostasy is implemented in the UiT model and S04 models.

Use regular non-italic font for units. Even in latex "Pa.s" can be made roman in math mode

We revise the text and make sure that Pa.s is non-italic.

1 3 You write insight into sub-surface structure. It is not really structure but rather rheology.

We follow the suggestion

2 30 Same as above

We follow the suggestion

1 4 Either spell out GRACE, or add "gravity" for clarity.

We follow the suggestion.

1 8 I would remove "deglaciation" here and describe the used time period in the paper.

For clarification we add "of the last glacial cycle" for clarification

1 16 Just to be clear, spell out GRACE or add gravity here the first time it is mentioned.

We follow the suggestion.

2 7 Here you could include the dynamic ice sheet model by Näslund et al. (2005): Näslund, J.-O., Jansson, P., Fastook, J. L., Johnson, J., and Andersson, L.: Detailed spatially distributed geothermal heat-flow data for modeling of basal temperatures and meltwater production beneath the Fennoscandian ice sheet, Ann. Glaciol., edited by: MacAyeal, D. R., International Glaciological Society, 40, 95– 101, doi:10.3189/172756405781813582, 2005.

We thank the reviewer for the suggestion but consider that the ice sheet models used in the study are enough to capture uncertainty in ice deglaciation chronology in the region.

3 7-8 "... best fitting models uplift rate measurements..." is difficult to understand. 3 29 "... we use ..." the software? There is an object missing in the sentence.

Indeed, we modify the sentence: “while best fitting models based on GPS uplift rate measurements have upper mantle viscosities”

4 6 Define "gravity disturbance rate" as opposed to "gravity anomaly rate".

Both terms are commonly used in physical geodesy, we add a reference to Hofmann-Wellenhof and Moritz physical geodesy book for clarification.

5 3 Reference to the ECCO model.

We add the reference to Forget et al. 2015 here.

5 4 What are the GAB products?

GAB files contain the ocean signal subtracted from GRACE and should be added back to restore GRACE’s full ocean mass mass variations. We add a reference to Flechtner et al. 2015 for the GAB products.

5 8 No italics.

We follow the suggestion.

5 17 Ocean bottom pressure changes in the Baltic can be neglected? Are they so much smaller than in the Barents, or just relatively smaller?

Both, they are smaller in the Baltic Sea and the signal there is higher (see Figure R5), so we decided to not consider the error

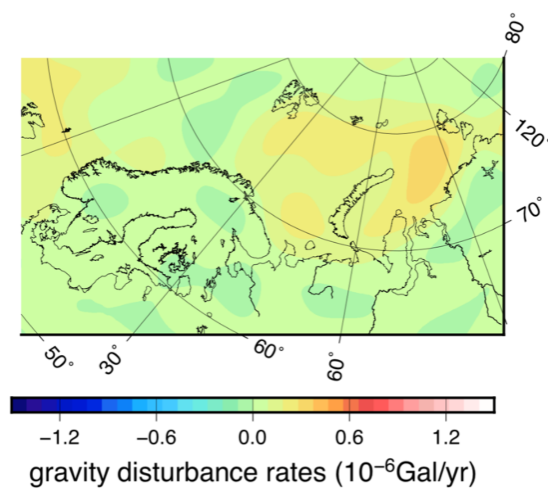

Figure R5: *Gravity disturbance rates due to ocean bottom pressure (from OMCT ocean model).*

6 28-29 This sentence need reformulation.

We reformulate the sentence.

7 1 Are you using central Fennoscandia? If so, where is this?

We do not use a specific point in Fennoscandia but find the point with maximum rate.

7 9 In Figure 1 it is the gravity signal after processing, not necessarily the GIA signal.

The reviewer makes a valid point here. We reformulate the paragraph: "A clear positive anomaly is evident both in Fennoscandia and the central Barents Sea where the main domes of the Scandinavian Ice Sheet and Svalbard-Barents-Kara Ice Sheet were presumably located (Figure 1), we assume this signal to be entirely due to GIA and call it the estimated GIA signal"

9 5 "A second set..." Which is the first?

We remove the 32*10^{20} upper mantle viscosity from our plot as it corresponds to a higher viscosity than that used for the lower mantle. Doing so the second subset is less evident and thus we decide to remove this paragraph.

9 9 ".. gravity rate which that is larger than..." Fix this.

Done

9 20 The authors should point out that 3D effects are indeed significant, e.g. Whitehouse et al. (2006), Steffen et al. (2006).

This is a good point. We modify the text accordingly which now reads. We decide to mention this in the method's section (6,L10-12)

" While this approach has been used in other GIA studies (Lambeck et al., 1998; Steffen et al., 2014), it has been suggested that far-field viscosity variations are relevant in Fennoscandia (Whitehouse et al. 2006)."

10 2 Which conclusion?

We add "of lateral viscosity changes between the two regions"

10 9 "...the reference model... a jump below 200 km" Please clarify which reference model and what the jump is, or refers to.

Changed to "a jump in the seismic velocity anomalies in the reference Earth models PREM and AK135

10 14 Stress for the flow law is taken from the GIA model. How accurate is this? Neglecting tectonics, topography, sediment loads etc surely distorts the "correct" stress state. How important is this?

Deviatoric stresses in the mantle from topography is small because of the long time-scale. Stresses from sediments are small as current uplift rates are small (van der Wal and Ijpelaar 2017). We add to the text (11,L33) "Background stresses due to mantle convection are neglected as recent work suggest little interaction between GIA and mantle convection "

10 23 Why did you choose the 1500 m contour?

We choose this contour to encompass the Scandinavian landmass and the Barents Sea, but avoid areas where the ice thickness was thin. A sentence is now added on page Section 3.3 (12,L9)

10 31 You should define "significant", or rather add uncertainties.

We delete "significant".

12 1 "... the GRACE misfit"? The GRACE GIA models?

We change the sentence which now reads as:

"This agrees very well with the results derived from the misfit of GIA models to GRACE data".

GRACE constraints on Earth rheology of the Barents Sea and Fennoscandia

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Abstract. The Barents Sea is situated on a continental margin and was home to a large ice sheet at the Last Glacial Maximum. Studying the solid Earth response to the removal of this ice sheet (Glacial Isostatic Adjustment, GIA) can give in-sight to~~in~~ the sub-surface rheology of~~in~~ this region. However, because the region is currently covered by ocean, uplift measurements from the center of the former ice sheet are not available.~~, but~~ The Gravity Recovery and Climate Experiment (GRACE) ~~GRACE~~ gravity data has been shown to be able to constrain GIA. Here we analyze GRACE data for the period $2003-2015$ in the Barents Sea and use it to constrain a GIA models for the region. We study the effect of uncertainty in non-tidal ocean mass models that are used to correct GRACE data and find that it~~ is not negligible and~~ should be taken into account when studying solid Earth signals in oceanic areas from GRACE. We compare the obtained gravity disturbance rates with GIA model predictions for different ice deglaciation chronologies of the last glacial cycle and infer a lower bound for the Earth's upper mantle viscosity of $2\cdot 10^{20}$ $\mathrm{Pa\cdot s}$. Following a similar procedure for Fennoscandia we find that the preferred upper mantle viscosity there is a factor 2 larger than in the Barents Sea for a range of lithospheric thickness values. This factor is shown to be consistent with the ratio of viscosities derived for both regions from global seismic models. The viscosity difference can serve as constraint for geodynamic models of the area.

1 Introduction

Ongoing viscous rebound of the solid Earth (Glacial Isostatic Adjustment, GIA) after the collapse of large ice sheets results in positive gravity disturbance rates in several regions of the Earth. The Gravity Recovery and Climate Experiment (GRACE)~~GRACE~~ satellite data has been used to constrain numerical models for GIA in North America (Tamisiea et al., 2007; Paulson et al., 2007; van der Wal et al., 2008; Sasgen et al., 2012) and Fennoscandia (Steffen and Denker, 2008; van der Wal et al., 2011; Simon et al., 2018). With longer time series it is now possible to observe weaker GIA signals such as that of the Svalbard-Barents-Kara Ice Sheet (SBKIS) in GRACE gravity data (Root et al., 2015a; Kachuck and Cathles, 2018; Simon et al., 2018). The use of GRACE data is especially relevant in this region as other geodetic observations normally used for GIA studies are only available from the islands surrounding the Barents Sea; in the periphery of the ice sheet that covered the region during the Last Glacial Maximum (LGM). This makes GIA-based ice sheet reconstructions such as ICE-5G and ICE-6G (Peltier, 2004; Peltier et al., 2015; Argus et al., 2014) uncertain.

Earlier work on the SBKIS proposed the existence of an extensive ice sheet spanning from the British Islands to the Kara Sea and extending further into mainland Russia (e.g., Grosswald, 1980, 1998) but more recent studies indicated a smaller ice sheet (e.g., Lambeck, 1995; Siegert and Dowdeswell, 1995; Svendsen et al., 1999, 2004; Mangerud et al., 2002). During the last decade, more geological and glaciological observations relevant for reconstructing the SBKIS have been obtained and compiled in the first version of the DATabase of Eurasian Deglaciation (DATED-1) resulting in new ice sheet limits for the whole Eurasian Ice Sheet Complex (EISC) (Hughes et al., 2016), but ice thickness variations can not be uniquely established.

Comparing the GRACE-derived gravity disturbance rates with those predicted for different palaeo-ice sheet configurations, Root et al. (2015a) conclude that the SBKIS contained less ice than previously thought. Kachuck and Cathles (2018) use GRACE data, along with Relative Sea Level (RSL) curves and GPS uplift measurements, to distinguish between two deglaciation histories: one with an ice sheet with a central dome in the Barents Sea and one with the Barents Sea marginally glaciated and domes in the surrounding Arctic islands. They show that the data is inconclusive in this regard.

Since the gravity disturbance rate signal in the Barents Sea region is small, it is important to thoroughly analyze the uncertainty in GRACE data. Here we present an extended analysis of GRACE data in the region and the different uncertainty sources. We focus on the gravity disturbance rate due to non-tidal mass variations in the ocean which influence the secular signal from GRACE data in oceanic areas (de Linage et al., 2009). In the processing chain to obtain Level 2 GRACE data, changes in ocean-bottom pressure are removed using the Ocean Model for Circulation and Tides (OMCT) forced with atmospheric data from the European Centre for Medium-Range Weather Forecasts (ECMWF). However, the OMCT secular signal is not reliable and should not be interpreted geophysically (Dobslaw et al., 2013). Lemoine et al. (2007) use a different ocean model in their GRACE data processing and find significant differences in the southern Arctic ocean.

We compare GRACE derived gravity disturbance rates to GIA model output to constrain the input of the GIA model. Because of uncertainty in solid Earth parameters and deglaciation history, it is difficult to uniquely constrain both. However, we can compare the GIA models for the Barents and Kara Sea areas with models for Fennoscandia constrained by the same data. In this way we can determine if there is a difference in Earth properties for both regions that is systematic for all deglaciation chronologies. Such constraints on variation in viscosity are useful for GIA modelling and geodynamic modelling in general, as viscosity maps derived from laboratory experiments and seismic velocities are not sufficiently constrained (e.g., Barnhoorn et al., 2011). Furthermore, the Barents Sea is located on a continental margin, and knowledge of the subsurface rheology~~structure~~ can help decipher its tectonic history. Our aim is to provide a lower bound on upper mantle viscosity for the Barents Sea region and Fennoscandia, focusing on the difference in viscosity between the two regions. We build on existing knowledge of Earth rheology and ice histories, which will be briefly reviewed in the following.

The rheology of the Barents Sea region is expected to be different from that of Fennoscandia, as it borders passive oceanic margins in the north and the west. Seismic tomography reveals lower seismic velocities in Barents Sea than below Fennoscandia (Levshin et al., 2007; Schaeffer and Lebedev, 2013), but not for all seismic periods and depths. 3D viscosity has been implemented in GIA models for the regions and has been found to affect sea level and uplift rates (Kaufmann and Wu, 1998). However, the difference in properties between Fennoscandia and Barents Sea has not been studied explicitly.

Constraints from palaeoshoreline data on 1D GIA models resulted in best fitting upper mantle viscosities of $2-6\cdot 10^{20} Pa\cdot s$ in the Barents Sea region (Steffen and Kaufmann, 2005), while recent work based on RSL data find that best fitting upper mantle viscosity in the Barents Sea region is above $2\cdot 10^{20} Pa\cdot s$ (Auriac et al., 2016). For Fennoscandia, the best fitting upper mantle viscosity is found to be between $3-7\cdot 10^{20} Pa\cdot s$ based on RSL data and relaxation time spectra, while best fitting models based on GPS uplift rate measurements have upper mantle viscosities up to $15\cdot 10^{20} Pa\cdot s$, see the overview in Steffen and Wu (2011). More recent work summarized in Simon et al. (2018) shows an upper mantle viscosity in the range of $3.4-20\cdot 10^{20} Pa\cdot s$. Note that the lower bound for upper mantle viscosity in the Barents Sea is somewhat below that in Fennoscandia. Steffen and Kaufmann (2005) computed RSL misfit and find similar upper mantle viscosity for the Barents Sea and the Scandinavian mainland, but smaller lower mantle viscosity. However, the different studies used different ice histories and relied on multiple data sources, with substantially less coverage in the Barents Sea region. Therefore it is unknown if it can be concluded from previous 1D studies whether viscosity is indeed lower in the Barents Sea than in Fennoscandia.

In this study we analyze GRACE data in the Barents Sea region and Fennoscandia to obtain the GIA signal there, focusing on the first region where the signal to noise ratio is lower. We compare the estimated~~observed~~ signal with 1D GIA model output to infer upper or lower bounds in viscosity for different ice deglaciation chronologies. From comparison between the best fitting models for the two regions we draw conclusions on the variation in Earth rheology between the Barents Sea and Fennoscandia.

2 Methodology

2.1 GRACE Data Processing

Temporal variations of the Earth's gravity field measured by GRACE are related to mass transport within the Earth system due to different geophysical processes, such as hydrology, ongoing cryospheric mass changes, GIA and~~,~~ (post-) seismic signals (e.g., Wouters et al., 2014). To study GIA, other geophysical signals that mask the GIA signal should be removed. Additionally, GRACE data is affected by instrumental noise and the anisotropic sampling of the signal due to the satellites' orbit (Wahr, 2007; Flechtner et al., 2016). Different data-processing techniques have been developed to increase the GRACE signal-to-noise ratio (e.g., Han et al., 2005; Swenson and Wahr, 2006; Kusche et al., 2009). In the following, we detail the post-processing used to analyze GRACE data in the Barents Sea and Fennoscandia with focus on the Barents Sea as it presents additional difficulties due to the smaller magnitude of the signal.

In our analysis we use the University of Texas Center for Space Research (UTCSR) release 5 (RL05) (Bettadpur, 2012) up to spherical harmonic degree 60. We use data for the $2003-2015$ period. We substitute the degree two coefficients with those obtained from satellite laser ranging (Cheng et al., 2013). To increase the signal-to-noise ratio in the Barents Sea we follow the strategy of Root et al. (2015a). We use a Gaussian filter to filter-out the noisy short wave-length gravity data and reduce GRACE's correlated errors which are evident as north-south stripes in the Level 2 data. We also use a high-pass filter in the Barents Sea to remove the long-wavelength signal that contains unmodelled long wave-length phenomena such as global sea-level rise. We adopt the half-widths used in Root et al. (2015a) which were tuned to optimize the signal-to-noise ratio in the Barents Sea (see Root et al. (2015a) Supplementary Material). The low-pass filter half-width ranges from 200 to 300 km while the high-pass filter half-width ranges from 500 to 700 km. As the signal in Fennoscandia is larger and has a larger wavelength we only use the low-pass filter there with a half-width also ranging from 200 to 300 km.

We compute gravity disturbance rate (gravity rate in the following) ~~as opposed to the gravity anomaly rate~~ (Hofmann-Wellenhof and Moritz, 2006). We use the least square method to obtain the secular, annual and semiannual signals of each time series of Stokes' coefficients. We estimate GRACE measurement errors (σ_{GRACE}) using the residuals after the secular, annual and semiannual signals are removed from the signal (Wahr et al., 2006).

After processing the signal as explained above, the GIA signal is evident as a positive gravity rate in Fennoscandia and the Barents Sea (Figure 1). However, this signal cannot be directly interpreted as it contains the trend of other geophysical processes as well, one of them being hydrology. Secular changes in land water storage result in gravity trends that should be subtracted when analyzing GRACE data in continental areas. The long-term hydrology signal in Fennoscandia is probably small, as demonstrated by the good agreement between GIA signal derived from GRACE and GPS (van der Wal et al., 2011). However, the hydrology signal of the Russian Arctic Archipelago (Novaya Zemlya, Franz Josef Land and Severnaya Zemlya) can leak into oceanic areas. We subtract the hydrology signal using the GLDAS hydrology model (Rodell et al., 2004). Because its reliability for the islands of the Arctic Archipelago is not well-known, we follow Matsuo and Heki (2013) and take the amplitude of its trend in the Barents Sea as an indication of the uncertainty in the hydrology signal in these polar regions ($\sigma_{hydrology}$).

Present-day changes in the cryosphere and the resulting present-day solid Earth response can also mask the GIA signal. In particular, the glaciers of the islands Svalbard and the Russian Arctic Archipelago are experiencing significant mass changes evident in GRACE observations which partly mask the GIA signal in the Barents Sea region (see Figure 1). Independent data on mass changes in Svalbard and the Russian Arctic Archipelago is limited. Moholdt et al. (2012) derived trends using ICESat for the 2003-2009 period using altimetry; other authors (e.g., Schrama et al., 2014; Matsuo and Heki, 2013) have used GRACE data. For the period 2003-2008, GRACE estimates are lower than altimetry estimates but agree within uncertainty (Root et al., 2015a). In Simon et al. (2018) ice mass loss estimates from altimetry and glaciology for a longer period were shown to be much larger than GRACE estimates in Svalbard, Franz Josef Land and Novaya Zemlya, and the former were scaled down in that study. Here we follow Root et al. (2015a) and use ice loss corrections obtained using the mascon method of Schrama et al. (2014) (see Table 1) to remove the ice loss signal taking into account elastic loading (Wahr et al., 1998).

To obtain the present-day mass changes from GRACE, a GIA correction needs to be first applied ~~; an ensemble of ice deglaciation chronologies and Earth rheological parameters is used and differences in obtained mass changes are accounted for in the mascon error budget.~~ As our aim is to quantify the GIA signal in the central Barents Sea, the problem seems circular.

However, the GIA model has a relatively small effect on the derived present-day mass changesHowever, the GIA model used to obtain the mass changes has a small effect in the recovered gravity rate trend in the central Barents Sea and is included in uncertainty estimate.We account for uncertainity in mass loss estimations due to GIA by employing an ensemble of ice deglaciation chronologies and Earth rheological parameters. We use the ICE-5G model and two runs of the GSM (Tarasov et al., 2012) with maximum and minimum ice sheet extents combined with the VM5a Earth model (Peltier, 2004) and an Earth model with a stronger mantle, as well as the W12 ice model (Whitehouse et al., 2012) with a strong mantle. Mass loss changes obtained using the different GIA models are shown in Table 1:, more massive ice sheet models and stronger mantles result in higher mass loss rates. The error in the derived mass changes due to uncertainty in GIA is similar to the GRACE measurement error. We use the error bars of the estimated mass changes for Svalbard and the Russian Arctic Archipelago to estimate the error in the recovered GIA gravity rates due to uncertainty in mass loss changes in the region (σ_{ice}). Finally, for the Barents Sea, the Greenland mass loss is already filtered when using the high-pass filter, but for Fennoscandia we need to remove it. To do so we use ICESat mass changes from Sørensen et al. (2011).

We account for the uncertainty in non-tidal ocean changes by using the ECCO ocean model (Forget et al., 2015) as alternative for the ocean model used in standard GRACE level 2 processing. In that case we first add back the GAB products to restore the full GRACE ocean mass signal (Flechtner et al., 2015; Yu et al., 2018)(Yu et al., 2018) before subtracting the ECCO ocean model. The ECCO model is a dynamically consistent ocean model constrained with observations from altimetry, Argo floats and GRACE (Forget et al., 2015). The model has been shown to correctly capture long-term bottom pressure variability in the Arctic Ocean and Adjacent Seas (Peralta-Ferriz, 2012). The version of the ocean model we use is the ECCOv4-llc270 compilation. This compilation covers the period 2001-2015 which means the GRACE time-series that we use in the Barents Sea is limited to this period. We obtain gravity rates in the central Barents Sea using the UTCSR GRACE solution corrected with both the OMCT and the ECCO ocean models respectively. The differences between these two solutions are used as an indication of the uncertainty in non-tidal ocean changes (σ_{ocean}).

We estimate the total error in the gravity trends by assuming that the different error sources are uncorrelated:

$$\sigma = \sqrt{\sigma_{ice}^2 + \sigma_{GRACE}^2 + \sigma_{ocean}^2 + \sigma_{hydrology}^2}. \quad (1)$$

The assumption that errors are uncorrelated requires further discussion. GRACE data is assimilated in the ECCO ocean model. However, GRACE is only one of the 40 data sets used in the inversion process and the final product does not fit GRACE data well (Yu et al. 2018). Therefore there will be only a weak correlation with the GRACE data used in our estimation. Correlation between land surface hydrology models and present-day ice melt is not expected, because hydrology models have little skill in predicting trends and do not model areas of permanent snow. Finally, ice loss changes errors (σ_{ice}) arise due to uncertainty in the GIA model and GRACE measurement error, we cannot rule out that the second error component might be correlated with σ_{GRACE}.

For the Barents Sea we consider the four terms; while for while when analyzing the gravity signal in Fennoscandia we only consider GRACE measurement errors, as the ice loss changes in the Arctic Archipelago and ocean bottom pressure changes have a very small effect on the gravity trends recovered in Fennoscandia, and~~ the~~ Greenland's mass loss signal is ~~relatively~~ well known from altimetry measurements.

2.2 GIA Modelling

We compare GRACE derived gravity rates with those predicted by GIA models. To compute the gravity trends the sea level equation is solved, using the pseudo-spectral approach presented in Mitrovica and Peltier (1991). We use the same code as Barletta and Bordoni (2013). To be able to run calculations for many different Earth parameters and ice models we assume that solid Earth properties only vary radially, which~~. This still~~ allows to compute GIA response for different regions separately with different viscosity profiles~~ as done by, (e.g., Lambeck et al., 1998; Steffen et al., 2014; Nield et al., 2014; Barletta et al., 2018)~~, but neglects effects of viscosity changes in surrounding regions. While this approach has been used in other GIA studies (Lambeck et al., 1998; Steffen et al., 2014), it has been suggested that far-field viscosity variations are relevant in Fennoscandia (Whitehouse et al., 2006).

We neglect the loading effect due to sediment transport during deglaciation, as the effect is small and well below that of the unknown ice thickness (0.01 to 0.05 $\mu\mathrm{Gal/yr}$ in Fennoscandia, and below 0.014 $\mu\mathrm{Gal/yr}$ in the Barents Sea as shown in van der Wal and IJpelaar (2017) ~~(van der Wal and IJpelaar, 2017)~~). To study the effect of the ice deglaciation history on the present gravity rates we start by using a reference Earth model based on the averaged VM2 model which is similar to the VM5a model (Peltier, 2004; Argus et al., 2014). The model consists of a 90 km lithosphere, a 570 km upper mantle with a viscosity of $0.5 \cdot 10^{21}\ \mathrm{Pa \cdot s}$ and a 2216 km lower mantle with an average viscosity of $2.6 \cdot 10^{21}\ \mathrm{Pa \cdot s}$. The elastic properties of the Earth are based on the PREM model (Dziewonski and Anderson, 1981). To investigate the effect of the Earth's rheology, we vary the upper mantle viscosity between $0.1 - 1.6 \cdot 10^{21}\ \mathrm{Pa \cdot s}$~~$0.1 - 3.2 \cdot 10^{21} Pa \cdot s$~~ and the lithospheric thickness between $40 - 180$ km (Table 2). We do not change the lower mantle as its viscosity cannot be constrained uniquely from data in Fennoscandia (Steffen and Kaufmann, 2005). However, gravity rates are influenced by the lower mantle viscosity which is discussed in the comparison with other GIA studies for the region.

We use an ensemble of ice histories that reflects the uncertainty in the deglaciation history of the European Ice Sheet Complex (EISC), the amount of ice in the SBKIS and the Scandinavian Ice Sheet (SIS) for the different ice deglaciation scenarios is shown in Figure 2 ~~(Figure 2)~~ .The ice sheet models that we use can be divided in two main categories: (1) empirical ice sheet models based on~~models based on GIA models that have been constrained using different~~ GIA observables and empirically-determined ice extents, and (2) those based on numerical ice-sheet modeling forced under different palaeo-climate scenarios and tuned to fit different constrains. A fundamental difference between these two kinds of models is that GIA-based palaeo-ice sheet models are explicitly associated with a specific Earth model. The first set of models is represented by the ICE-5G and ICE-6G models (Peltier, 2004; Peltier et al., 2015; Argus et al., 2014). Both models start the ice build-up 122 ka BP. T~~; t~~he second set consists of three models obtained using the Glacial System Model (GSM) for Northern Europe (Tarasov et al., 2012), the University of Tromsø Ice Sheet Model (UiT ISM) (Patton et al., 2016, 2017), and the S04 ice sheet model (Siegert and Dowdeswell, 2004), which are further described below.

The three ice sheet models obtained using the GSM model are a subset of a bigger ensemble used in Root et al. (2015a) which showed good agreement with GRACE observations. The ensemble was obtained using a Bayesian calibration ofconstraining GSM runs with RSL curves, present-day ground velocities and ice deglaciation margins from the DATED-1 project (Hughes et al., 2016). The VM5a rheology model was employed as reference during the calibration process, however, errors introduced by the rheology model were accounted for during the calibration process which implies that this model is not as strongly biased by a single viscosity profile as the ICE-5G and ICE-6G models. The three selected models consist of a late deglaciation model, labeled nn45283 in Root et al. (2015a) and two early deglaciation models, nn56536 and nn56597, with different maximum ice volumes. The build-up phase is faster than for the ICE-5G and ICE-6G models, build-up starts 28 ka BP. The models will be labeled T1 (nn45283), T2 (nn56536) and T3 (nn56597) to simplify the notation.

The University of Tromsø Ice Sheet Model is based on a 3D thermomechanical ice model which uses an approximation of the Stokes equations forced by climatic and eustatic sea level perturbations to simulate the evolution of the EISC. The model is constrained using different geophysical and geological data sets including geomorphological flow sets, moraine and grounding zone wedge positions and isostasy patterns and is consistent with the DATED-1 ice sheet margins. Isostatic loading is implemented using the elastic lithosphere/ relaxed astenosphere model of Le Meur and Huybrechts (1996). The model has no ice in the region before 37 ka BP.

Finally, we consider an ice sheet model which gives a lower bound for the mass present in the Barents Sea during the LGM, the S04 model (Siegert and Dowdeswell, 2004). The model is based on the continuity flow equations coupled with a model of water, basal sediment deformation and transportation. The model is forced with eustatic sea level curves of the last 30 ka and palaeo air temperatures and precipitation and assumes an ice-free scenario before 32 ka BP. Bedrock topography is adjusted for isostasy using the method of Oerlemans and van der Veen (1984).

The only global ice sheet models are the ICE-5G and ICE-6G, for the other ice sheet models we use the ICE-6G ice model outside the EISC. We include the build-up and deglaciation phase of the last glacial cycle. All ice sheet models are sampled in a grid with a spatial resolution corresponding to a 128 degree Gaussian grid and the output of the model is truncated at degree 60 and processed using the same filters used to process the GRACE data.

2.3 Model Performance Assessment

We assess the fit of the modelled and estimatedobserved gravity rates for different combinations of ice deglaciation history and rheologyWe assess the fit of each combination of ice deglaciation chronology and rheology by comparing the GRACE data after filtering and correcting for other signal with the corresponding filtered GIA model signal. As GRACE's resolution is of the same order of magnitude as the extension of the SBKIS we cannot resolve the differences in the shape of the ice sheet in the data. Thus we assess the model fit only by comparing the maximum modelled (m_i) and estimatedobserved (e_i) gravity rate in the central Barents Sea and Fennoscandia and normalize this difference using the observation error (σ_i). In order to make the results as independent of the filter parameters as possible, we compute the average of the misfit obtained using different filter configurations:

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

where N is the number of filter settings. For each ice sheet model we define a 95% confidence interval for the solid Earth parameters as all models with a $\Delta\chi^2$ smaller than 5 (Press et al., 1992):

$$\Delta\chi^2 = \chi^2 - min(\chi^2) \quad (3)$$

The low pass-filter is varied between 200 and 300 km half-width in 20 km intervals. Additionally, in the Barents Sea the high-pass filter is varied between the 500 and 700 km half-width in 100 km intervals.

3 Results

3.1 GRACE GIA signal in Fennoscandia and the Barents Sea

We use the methods presented in Section 2 to obtain the gravity ratesthe GIA signal over Fennoscandia and the Barents Sea. A clear positive anomaly is evident both in Fennoscandia and the central Barents Sea where the main domes of the Scandinavian Ice Sheet and Svalbard-Barents-Kara Ice Sheet were presumably located (Figure 1), we assume this signal to be entirely due to GIA and call it the estimated GIA signal. The melting of ice in Svalbard and the islands of the Russian Arctic Archipelago is also evident as a negative gravity trend. After removing the mass loss signal as explained in Section 2, we observe that most of the signal of Novaya Zemlya, Svalbard and Franz Josef Land is indeed removed (Figure 1). However, there is still a negative gravity rate left over Severnaya Zemlya, indicating that our ice loss changes might be underestimated for this island. We do not observe a clear positive signal in the Kara Sea, which indicates that if it was glaciated during the LGM the amount of ice present there was much smaller than that located in the Barents Sea. This fact advocates against the larger ice sheets in Denton and Hughes (1981); Grosswald (1998); Grosswald and Hughes (2002) and further confirms the results of the DATED-1 (Hughes et al., 2016) and QUEEN projects (Svendsen et al., 2004).

We obtain the maximum gravity rate in the Barents Sea for different filter configurations using the OMCT and ECCO ocean models. Figure 3 shows the maximum gravity rates for a 500 km high-pass filter and different low-pass filter half-widths. As expected, we observe that the maximum gravity signal reduces with increasing filter half-width and so does the error. The gravity rates recovered using the ECCO ocean model are systematically higher that those obtained with the OMCT model. We also show a breakdown of the error (Figure 4) for different low-pass filter half-widths. We observe that the hydrology signal leaking into the Barents Sea is very small and the error budget is dominated by the uncertainty in present-day ice changes, the GRACE measurement error and the non-tidal ocean signal. Moreover, we observe that while the other error sources decrease with increasing filter half-width the ocean error does not. This implies that it has a wavelength similar to that of the GIA signal we want to resolve.

3.2 Implications for viscosity and ice sheet chronology

We perform three experiments. In the first experiment we only study the effect of the ice history on the model misfit. We use the reference Earth model (see Table 2) and compare the fit of the predicted gravity rates for different ice deglaciation models with the GRACE derived gravity rate. In the second experiment, we change the Earth rheological parameters to obtain the subset of ice deglaciation histories and Earth rheological parameters that best fit the GRACE observations. Thirdly, we repeat the second experiment for Fennoscandia and compare the optimal solid Earth parameters for both regions to detect possible variations in rheological parameters.

Figure 3 compares the maximum present-day estimatedobserved gravity rates in the Barents Sea with those given by for the different ice sheet models. It must be noted, that the maximum gravity rates produced by each ice history are not only related to the maximum ice volume attained during LGM, but also its geographical distribution and the onset of the deglaciation process. As an example, we find that while the T2 model has more ice in the Barents Sea than the T1 model, it results in lower gravity rates. This is because deglaciation starts earlierlater in the T2 model than in the T1 model when the sensitivity of the present gravity rates to mass changes is higher as shown in Figure 5. Similarly, the highest gravity rates are associated with the UiT ISM even though it has less ice than the ICE-5G and ICE-6G model. This is because the UiT ice sheet model has more ice in the central Barents Sea during the last phase of deglaciation. In fact, the model includes an ice bridge between Svalbard, Franz Josef Land and Novaya Zemlya with ice thickness as large as 2000 m at 14.5 ka BP which does not disappear until 12 ka BP. This is not present in either the ICE-6G or the ICE-5G models.

When we compare the modelled and estimated gravity rates with GRACE observations we find that, for the reference Earth model, the T1, T2 and T3 ice sheet models are the closest to observations. The S04 ice sheet model performs worse; the model does not have enough ice in the region. This result is in accordance with Auriac et al. (2016) who found poor agreement between the S04 model and RSL curves. The more massive ICE-5G, ICE-6G and UiT models result in gravity rates that are too high. However, the discrepancy between these models and GRACE derived estimations observations is reduced if we use the ECCO ocean model instead of the OMCT. Furthermore, the GRACE data can be reconciled with the UiT ISM if the maximum volume of mass in the model is reduced by around 1 m of equivalent sea level rise or if deglaciation started 1 kyr earlier.

Next, we study the effects of changing the solid Earth rheology in the Barents Sea. Figures 6 and 7 (left column) show the misfit of the different ice sheet models to the estimatedobserved maximum gravity rates in the Barents Sea for different rheology models. We see that there is a large subset of Earth rheological parameters for which the modelled gravity rate is within the 95% confidence interval2σ interval of the GRACE estimatedobserved gravity rate.

The T1, T2 and T3 ice sheet models present a good fit to the observations for a large subset of Earth models including the reference Earth model ($\nu = 5 \cdot 10^{20}$ $\mathrm{Pa \cdot s}$, $h = 90$ km). For the less massive S04 model the confidence interval extends from $\nu = 5 \cdot 10^{20}$ to $\nu = 1.6 \cdot 10^{21}$ $\mathrm{Pa \cdot s}$. In contrast, for the more massive ice sheets (ICE-5G, ICE-6G and UiT ISM) the subset of Earth models which present a good fit to the Barents Sea observations is smaller and does not contain the reference

Earth model. These models, however, fit the observations either for a less viscous~~lower~~ upper mantle~~ viscosity~~ or for a thicker lithosphere when upper mantle viscosity is fixed. If a less viscous~~lower~~ upper mantle viscosity is used the relaxation time of the solid Earth is decreased and the sensitivity to mass changes that occurred during the LGM decreases (see Figure 5). On the other hand, a thicker lithosphere acts as a low pass filter that smooths the gravity signal, reducing its maximum value.

Our results for the UiT ISM are consistent with those obtained by Patton et al. (2017) who inferred an upper mantle viscosity of $2 \cdot 10^{20}$ Pa·s based on RSL data. The lower bound obtained with the other ice models is similar as models with a low viscosity have little sensitivity to mass changes during the early deglaciation phase, where differences between ice models are more manifest (Figure 2). Overall, . ~~Finally, using all the ice sheet models we can infer a lower bound for the upper mantle viscosity in the Barents Sea of~~ ~~$2 \cdot 10^{20}$ Pa·s~~ for a lower mantle viscosity of $2.6 \cdot 10^{21}$ Pa·s, we obtain a lower bound for the upper mantle viscosity of $2 \cdot 10^{20}$ Pa·s, which agrees with the range of possible upper mantle viscosity found in Auriac et al. (2016) using RSL curves and GPS uplift measurements. We refrain from drawing conclusions on the preferred lithosphere thickness from the misfit plots because the lithosphere has a large influence on the shape of the gravity rate pattern which was not used as constraint here. A higher lower mantle viscosity can result in a lower bound as shown in Steffen et al. (2010); Root et al. (2015b).

~~A second set of good fitting models is found for some ice models. For example, for ice models T1, T2 and T3 a good fit is obtained for upper mantle viscosity of~~ ~~$3.2 \cdot 10^{21} Pa \cdot s$~~ ~~in the Barents Sea, and in Fennoscandia for ice model T1. For ice models T2 and T3 this set could exist for even higher upper mantle viscosity, outside the range studied here. This result is found more often in GIA studies studies (e.g. Lidberg et al. (2010), Root et al. (2015c)). The explanation is that a certain viscosity results in maximum gravity rate which that is larger than the estimatedobserved gravity rate, and both increasing and decreasing viscosity could reduce the gravity rate, and hence result in a good fit. To exclude one of the two sets, an extra dataset is required. Based on other best fitting models in literature for the Barents Sea (Auriac et al., 2016) and Fennoscandia (Steffen and Wu, 2011) the good fitting models of large upper mantle viscosity are not likely and we do not consider them further in this study.~~

We follow the same procedure for Fennoscandia to obtain the subset of Earth rheological parameters and ice sheet deglaciation histories with an acceptable agreement with the GRACE observations (see right column of Figures 6 and 7). It must be noted that the values of the χ^2 are higher for Fennoscandia than the Barents Sea and thus the confidence interval is smaller. The reason is twofold: the observation error is smaller as compared with the Barents Sea, where uncertainty from~~form~~ mass changes in the glaciers of the surrounding islands and non-tidal ocean changes increase the error bars; and the GIA signal is higher in Fennoscandia than in the Barents Sea (see Figure 1). Nevertheless we can compare the best-fitting models for both regions.~~ This assumes that 1D models can be used to represent each area separately and there are no 3D effects, an approach that is followed in other GIA studies as well (Lambeck et al., 1998; Steffen et al., 2014)~~.

We observe that, contrary to what we got for the Barents Sea, the combination of the ice sheet models ICE-5G~~, ICE-6G~~ and UiT with the reference lithospheric thickness and upper mantle viscosity have a good fit~~n optimal~~ fit. (Figures 6 and 7). As already mentioned, the ICE-5G and ICE-6G models have been constrained using GIA observations, which are abundant in Fennoscandia. As we are using these models with an Earth rheology similar to its~~their~~ reference rheology it is not surprising that the ICE-5G model presents a good fit~~ is obtained for~~in this region, however the ICE-6G model performs better with a more viscous mantle due to its lower ice volume. T On the other hand, the T1-3 models do not fit the estimated GIA signal with the observations with the reference Earth model and require a more viscous mantle. The early deglaciation of the S04 model results in low gravity disturbance rates that do not fit the GRACE estimated gravity disturbance rates. For Fennoscandia we find a lower bound for the upper mantle viscosity of $5 \cdot 10^{20}\ \mathrm{Pa \cdot s}$ which is consistent with current estimates (Simon et al., 2018).

We can infer lateral rheology changes by comparing the optimal Earth rheological parameters obtained for both regions. For each ice deglaciation chronology, we compare the two confidence intervals as well as the best fitting upper-mantle viscosity obtained for each lithospheric thickness. We observe that for the UiT, ICE-5G and ICE-6G model both the confidence intervals as well as the best fitting models systematically prefer a less viscous upper upper mantle in the Barents Sea as compared with Fennoscandia. This is also the case for the T1,T2 and T3 models when the best fitting models are compared, although there is an overlap of models of high upper mantle viscosity and thick lithospheres with a good fit in both regions. We observe a systematic difference between the optimal Earth rheological parameters in Fennoscandia and the Barents Sea for all the ice sheet models. The different ice sheet models have a preference for a less viscouslower upper mantle viscosity (around a factor of two smaller) in the Barents Sea as compared with Fennoscandia, or for a thicker lithosphere if the upper mantle viscosity is equal in both regions. This systematic difference is likely evidence of lateral variation in Earth rheology.

3.3 Lateral viscosity variation

To strengthen the conclusion of viscosity differences between the two regions from the previous subsection, we derive viscosity estimates in an independent way, based on seismic velocity anomalies and experimentally derived flow laws. The absolute viscosity values obtained in this way contain large uncertainty, but the relative difference resulting from the seismic models should represent real change in temperature or composition. Therefore we focus on the ratio between the viscosities beneath Fennoscandia and the Barents Sea and check whether it agrees with the outcome of the GIA model misfit.

To take uncertainty in seismic velocity anomalies into account we use two global seismic tomography models: S40RTS Ritsema et al. (2011) and Schaeffer and Lebedev (2013) (labeled SL) which has higher spatial resolution but reduced sensitivity with depth. For both, the reference model is adjusted to account for a jump in the seismic velocity anomaly in PREM (Dziewonski and Anderson, 1981) and AK135 (Kennett et al., 1995) reference model below 200 km. Shear wave velocities are converted to temperature using relations from geochemistry (Goes et al., 2000; Cammarano et al., 2003) for primitive mantle composition and accounting for anelasticity (anelastic correction model Q4 from Cammarano et al. (2003)). Differences in composition between the Barents Sea and Scandinavia could play a role, but is unlikely to reverse the temperature contrast, due to the first order effect of temperature on seismic velocities in the upper mantle (Goes et al., 2000). To compute viscosity we follow the procedure in Wal et al. (2013) and insert temperatures in the olivine flow laws of Hirth and Kohlstedt (2013). The flow laws for diffusion and dislocation are added, which means the viscosity depends on grain-size and stress. Stress is taken from a 3D GIA model which uses the ICE-5G ice load. Background stresses due to mantle convection are neglected as recent work suggest little interaction between GIA and mantle convection (Huang, 2018). Grain size is chosen to be 4 mm or 10 mm.

4 mm gave best overall fit to GIA data in and 10 mm grain size resulted in the best fit with the observed maximum uplift rate (Wal et al., 2013).

To be able to compare against viscosity for the upper mantle in the previous section we use viscosity averaged between 225 and 325 km. This depth is a trade-off; shallower layers have lower temperature and small viscous deformation during the glacial cycle, while for deeper layers the seismic models are less accurate. The depth range is also close to the depth to which the gravity rate in Fennoscandia is most sensitive, see the sensitivity kernels in van der Wal et al. (2011). The viscosity maps are plotted in figure 8. In principle all viscosity values around the ice load play a role in the GIA process, but the highest sensitivity is to values directly underneath the ice load (Paulson et al., 2005; Wu, 2006). We compute the average of viscosities for the locations where LGM ice heights are above 1500 m which covers most of the land mass of Scandinavia, and most of the Barents Sea (see dashed brown contour). Viscosity is computed separately for the region below 71° latitude for Fennoscandia ~~black line~~ and above ~~and including 71° latitude~~ for the Barents Sea. We find that the average viscosity below Fennoscandia is a factor of 2.3 to 2.4 times higher than that in the Barents Sea. This agrees well with the change in best fit upper mantle viscosity that can be seen in the misfit figures 6 and 7. There could still be an effect of 3D structure that is not captured by modelling both regions with 1D models, such as lateral variations within Fennoscandia (Steffen et al., 2014) or the influence of viscosity from outside each region.

4 Conclusions and Discussion

In this study, we analyse GRACE data in the Barents Sea to constrain the Earth rheology in the region. We compare the fit of different GIA models in Fennoscandia with that for the Barents Sea to find if there is a ~~significant~~ difference in viscosity between the two regions. We investigate several deglaciation chronologies of the SBKIS, some of which are not explicitly tied to a viscosity model. We use GRACE data for the period 2003 – 2015 and process it to reveal the GIA signal. The ice loss signal from the Svalbard and the Russian Arctic Archipelago is removed using mass change values obtained from GRACE using the mascon method. We observe a positive gravity anomaly in the Barents Sea but no significant anomaly in the Kara Sea, which shows that the ice cover at LGM was considerably thinner there than in the Barents Sea, in agreement with recent studies.

The Barents Sea GIA signal is in a region now covered by sea; therefore, the gravity trends might be affected by non-tidal oceanic mass changes. We correct GRACE gravity rates in the Barents Sea using either of two ocean models, the OMCT and ECCO ocean model, and find higher gravity rates using the ECCO model. The difference in the ocean signal according to the two models is large in the Barents Sea. This uncertainty has not been considered in previous studies of the GIA signal in the region (e.g., Root et al., 2015a; Simon et al., 2018; Kachuck and Cathles, 2018) and thus the errors bars in those studies were provably underestimated. This result has also implications for GRACE studies of non-oceanic mass changes, such as post seismic deformations, in ocean areas (e.g., Han and Simons, 2008; Wang et al., 2012) which possibly have higher uncertainty than previously thought due to errors in the ocean model.

We compare the GRACE derived gravity rates with modelled ones to infer geophysical constraints for the Earth rheology and ice sheet chronology in the Barents Sea region. For a three-layer average of the VM2 viscosity profile (Peltier, 2004) we find, as Root et al. (2015a), that thick ice sheet models (ICE-5G, ICE-6G and UiT) do not fit GRACE observations, while the less massive ice models (T1,T2 and T3) do. Upper mantle viscosity and lithospheric thickness was varied for each ice sheet chronology between $0.1 \cdot 10^{21} - 32 \cdot 10^{21}$ Pa · s and $40 - 180$ km. We find that the ICE-5G, ICE-6G and UiT ice sheet models can be reconciled with GRACE observations provided the upper mantle viscosity is lower or the lithosphere thicker than in the VM2 model. The same conclusion is reached in Auriac et al. (2016) using GPS uplift measurements and RSL curves instead of gravity data.

The interplay between ice deglaciation chronology and Earth rheology makes it difficult to constrain the ice deglaciation chronology in the Barents Sea (Kachuck and Cathles, 2018). Root et al. (2015a) used GRACE data to conclude that the SBKIS had less ice than previously thought ($5 - 6.3$ m of equivalent sea level versus 8.3 m). To do so, they used ICE-5G and ICE-6G and showed that they do not obtain the estimated~~observed~~ gravity rate when these ice models are combined with their corresponding Earth rheology model. However, here we use the UiT ISM which does not come with an *a priori* Earth rheoloy model and which contains around 7.5 m of equivalent sea level rise and show that it can fit GRACE observations provided the upper mantle viscosity is around $3\ 10^{20}$ Pa · s if the lithophere is thinner than 130 km. However, we are able to place a constraint on upper mantle viscosity. From the misfit of all investigated ice chronologies and using a lower mantle viscosity of $2.6 \cdot 10^{21}$ Pa · s, we infer a lower bound for the upper mantle viscosity of $2 \cdot 10^{20}$ Pa · s, which agrees with previous constraints derived from RSL and GPS uplift observations Auriac et al. (2016).

We also study the misfit of GRACE observations to the GIA models in Fennoscandia. For a $2.6 \cdot 10^{21}$ Pa · s lower mantle viscosity, w~~W~~e obtain a lower bound of $5 \cdot 10^{20}$ Pa · s for the upper mantle viscosity, which is consistent with current estimates. Given all the ice sheet deglaciation chronologies we find that the lower bound for the upper mantle viscosity is a factor of two smaller in the Barents Sea (or, alternatively, the lithosphere thickness should be increased there). Unless all the tested ice deglaciation chronologies are biased in the same direction, this result is evidence of lateral changes in viscosity in between the two regions.

To strengthen the finding of viscosity difference between the two regions, we compare our results with viscosity derived from global velocity anomalies and flow laws for mantle material and find that the average viscosity in the Barents Sea is a factor of 2.4 lower than in Fennoscandia. This agrees very well with the results derived from the misfit of GIA models to GRACE data~~the GRACE misfit~~, and strengthens the conclusion that there is a small but significant difference in average upper mantle viscosity between the two regions. This findings have implications for ice sheet models inverted with just one viscosity profile (e.g., ICE-5G, ICE-6G) and advocates in favour of including lateral Earth rheological parameters in GIA models. The constraints on viscosity variations can be also used to calibrate other geodynamic models of the regions.

Code and data availability. Gravity rates for the different ice sheet models and Earth rheology models as well as GRACE maximum disturbance rates for Fennoscandia and the Barents Sea are provided at http://doi.org/10.4121/uuid:424126e6-b5d3-4ac9-b5cd-f495c8ad6939. The GIA code used for the simulations is available upon request from VRB.

Author contributions. All authors contributed to the discussion and commented on the manuscript. M.R.N and W. v.d.W led the writing of the article. V.R.B contributed with her GIA code. M.R.N analysed GRACE data and ran the GIA simulations. W.v.d.W. provided the 3D viscosity maps. All authors contributed to the interpretation of the results.

Competing interests. The authors of this manuscript declare that they do not have any conflict of interest.

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Table 1. Ice loss changes in Svalbard the Islands of the Russian Arctic Archipelago between 2003 and 2015 in $\mathrm{Gt/yr}$ obtained for different GIA models. The ICE-5G model and two runs of the GSM with maximum (GLAC2) and minimum (GLAC1) ice sheet extents that comply with RSL and GPS observations combined with the VM5 Earth rheological model or a model with stronger mantle, labelled M2, with $\mu_{UM} = 1.6 \cdot 10^{21}$ $\mathrm{Pa \cdot s}$ and $\mu_{LM} = 5.12 \cdot 10^{22}$ $\mathrm{Pa \cdot s}$. Additionally, the W12 ice model with $\mu_{UM} = 1 \cdot 10^{21}$ $\mathrm{Pa \cdot s}$ and $\mu_{LM} = 1 \cdot 10^{22}$ (M3) is also used. The last row indicates the average value and uncertainty due to GRACE measurement error and uncertainty in the GIA model

Ice Model	**Rheology**	**Novaya Zemlya**	**Svalbard**	**Franz J.Land**	**Servernya Zemlya**
GLAC1	M2	4.71 ± 0.42	5.05 ± 0.49	1.12 ± 0.19	0.76 ± 0.10
GLAC1	VM5a	4.94 ± 0.42	4.96 ± 0.49	1.10 ± 0.19	0.63 ± 0.10
GLAC2	M2	4.60 ± 0.42	4.90 ± 0.49	0.93 ± 0.19	0.92 ± 0.10
GLAC2	VM5a	4.57 ± 0.42	4.85 ± 0.49	0.85 ± 0.19	0.67 ± 0.10
ICE-5G	M2	5.87 ± 0.42	5.77 ± 0.49	1.68 ± 0.19	0.80 ± 0.10
ICE-5G	VM5a	4.54 ± 0.42	5.16 ± 0.49	1.03 ± 0.19	0.70 ± 0.10
W12	M3	6.13 ± 0.42	5.34 ± 0.49	1.64 ± 0.19	0.46 ± 0.10
-	-	5.15 ± 0.58	5.05 ± 0.79	1.19 ± 0.38	0.70 ± 0.18

Table 2. Solid Earth rheological parameters for this study: lithosphere thickness (h_l), upper mantle viscosity ν_{UM} and lower mantle viscosity ν_{LM}

Parameter	**Reference Model**	**Range**
h_l (km)	90	$40 - 180$
ν_{UM} (10^{21} $\mathrm{Pa \cdot s}$)	0.5	$0.1 - 1.6$
ν_{LM} (10^{21} $\mathrm{Pa \cdot s}$)	2.6	2.6

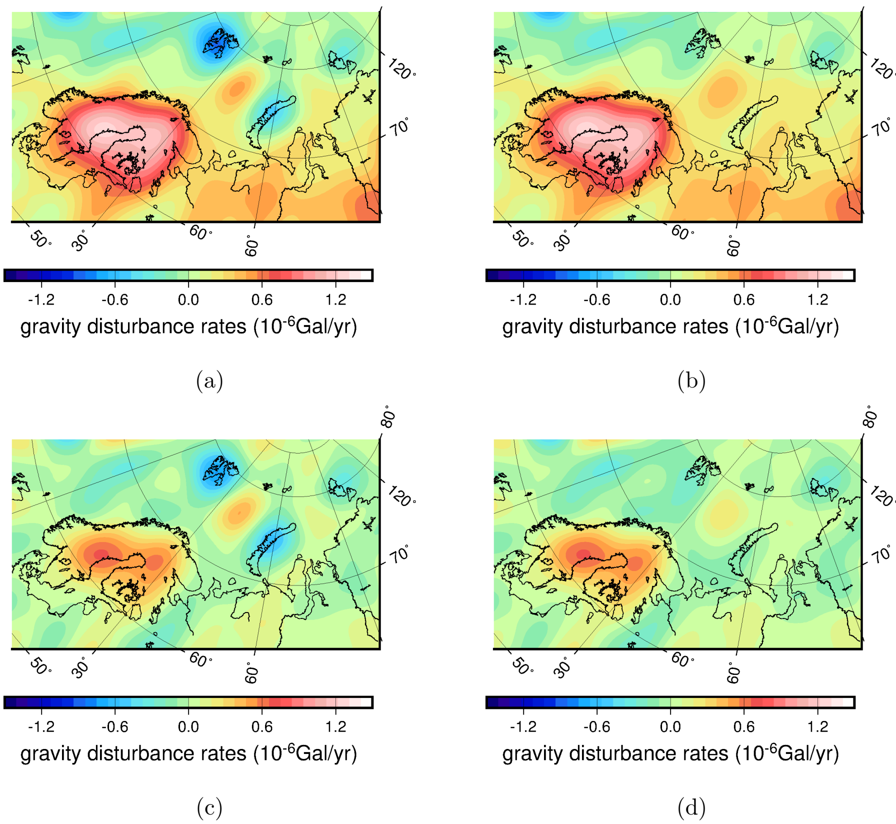
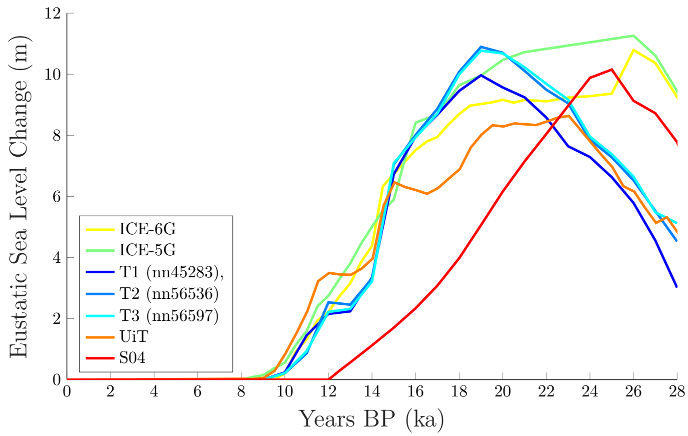
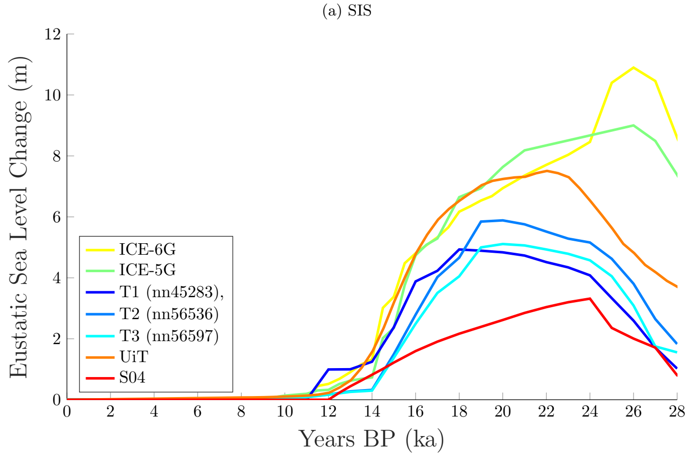


Figure 1. Gravity signal in Fennoscandia and the Barents Sea for the period 2003-2015. (a) and (b) show the gravity disturbance rate filtered with a 200 km low-pass filter while in (c) and (d) the data is additionally filtered with a 600 km high-pass filter to remove long wavelength signals. The mass loss signal of the Arctic Archipelago islands has been removed in (b) and (d).



(a) SIS



(b) SBKIS

Figure 2. Volume of ice present in the SIS (a) and SBKIS (b) during the last glacial period given in equivalent eustatic sea level rise for different ice sheet reconstructions. Six different deglaciation chronologies are shown: the GIA-constrained models ICE-5G and ICE-6G (Peltier, 2004; Peltier et al., 2015; Argus et al., 2014); three models obtained using the Glacial System Model (GSM) (Tarasov et al., 2012), the T1, T2 and T3 chronologies; the University of Tromsø Ice Sheet Model (UiT) (Patton et al., 2017); and the S04 ice sheet model (Siegert and Dowdeswell, 2004). The divide between both ice sheets is taken to be the 70° parallel.

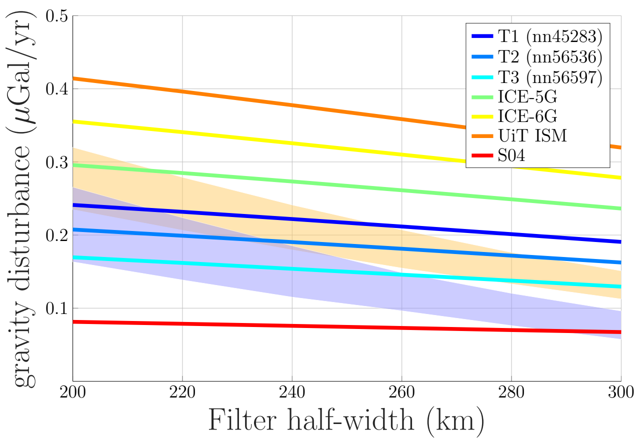


Figure 3. Maximum gravity rate in μGal/yr recovered in the central Barents Sea using GRACE, after removing the ocean signal with the OMCT ocean model (blue) or ECCO ocean model (orange) for different low-pass filter half-widths and a 600 km half-width high-pass filter. The GIA signal for different ice deglaciation histories with a reference Earth model is also shown.

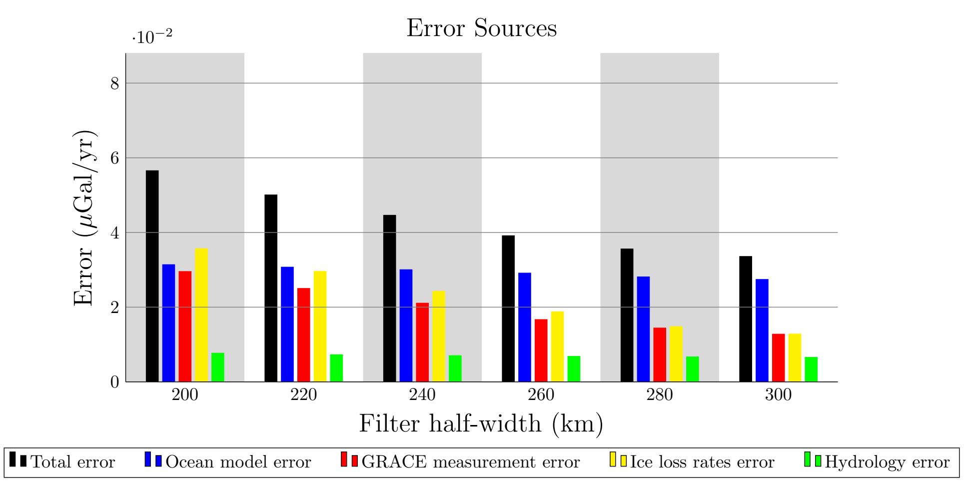


Figure 4. Error in μGal/yr in the maximum gravity rate in the central Barents Sea from different sources. The magnitude of the error is given for different low-pass filter half-widths and a high-pass filter half-width of 600 km.

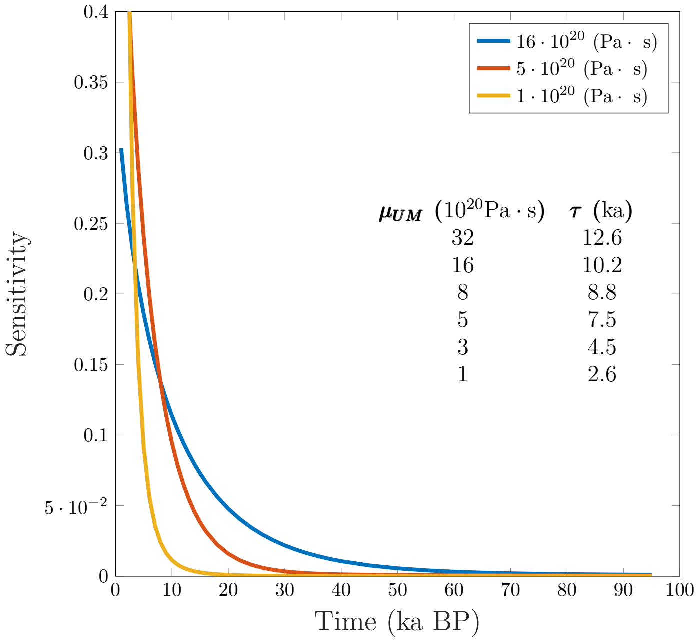
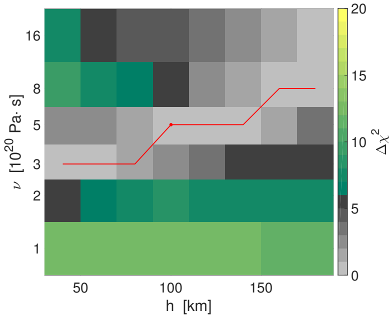
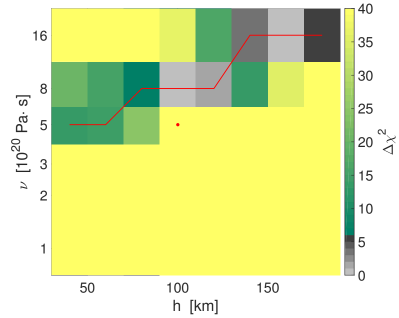


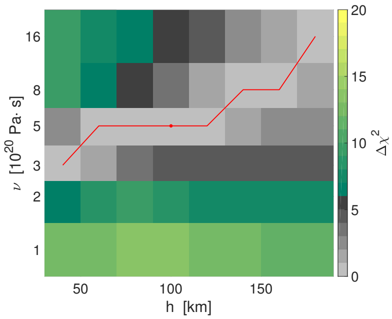
Figure 5. Present gravity disturbance rate induced by a uniform mass change in the Barents Sea at a given epoch for three different upper mantle viscosity. The results have been normalized using the maximum gravity disturbance rate obtained with $\mu_{UM} = 1 \cdot 10^{20}$ Pa · s. Inset: relaxation times for different upper mantle viscosity.



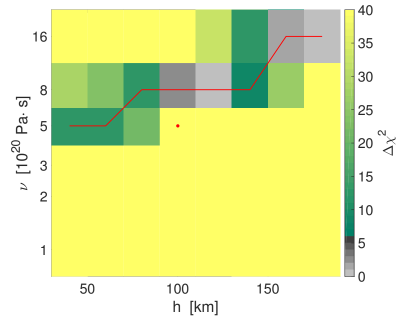
(a) T1 Barents Sea



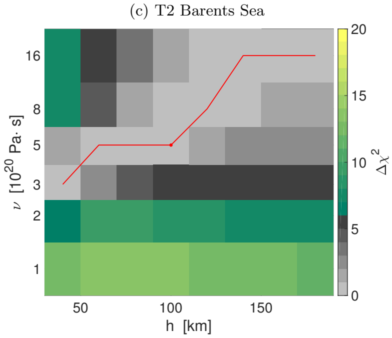
(b) T1 Fennoscandia



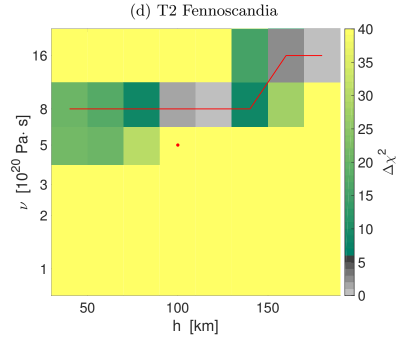
(c) T2 Barents Sea



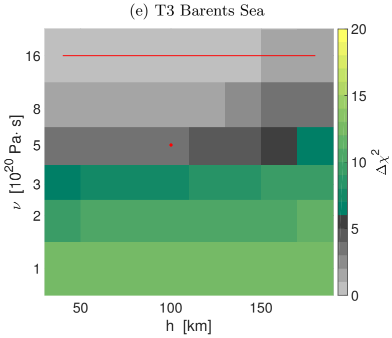
(d) T2 Fennoscandia



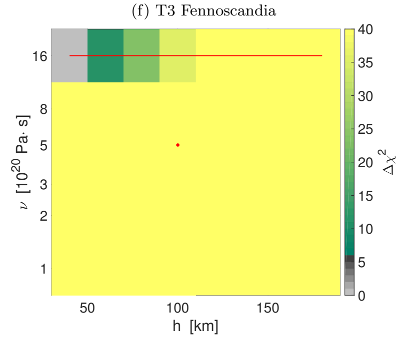
(e) T3 Barents Sea



(f) T3 Fennoscandia



(e) S04 Barents Sea



(f) S04 Fennoscandia

Figure 6. Misfit (χ^2) of the T1,T2,T3 and S04 ice deglaciation chronologies to GRACE observations for different values of upper mantle viscosity (ν) and lithospheric thickness (h) in the Barents Sea (left column) and Fennoscandia (right column). The fit is given in terms of the $\Delta\chi^2$. The circle indicates the reference model and the red line shows the best fitting model for each lithospheric thickness

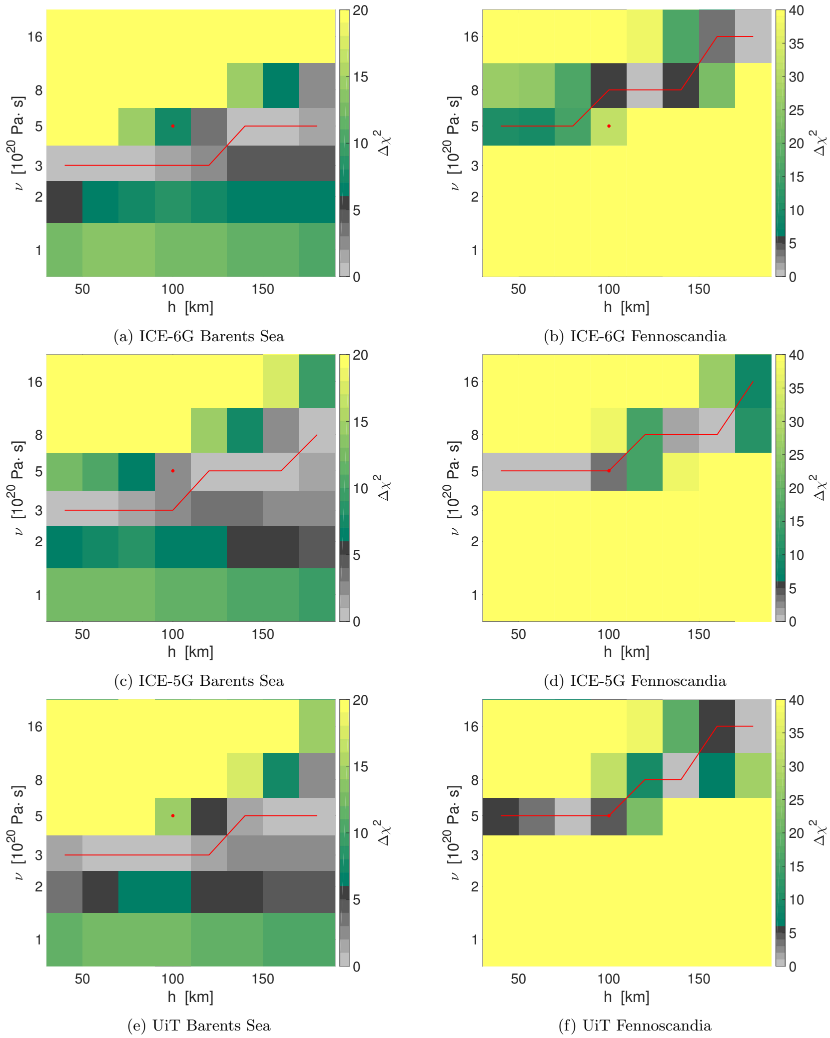


Figure 7. Same as Figure 6 but for the ICE-6G, ICE-5G and UiT ice sheet models.

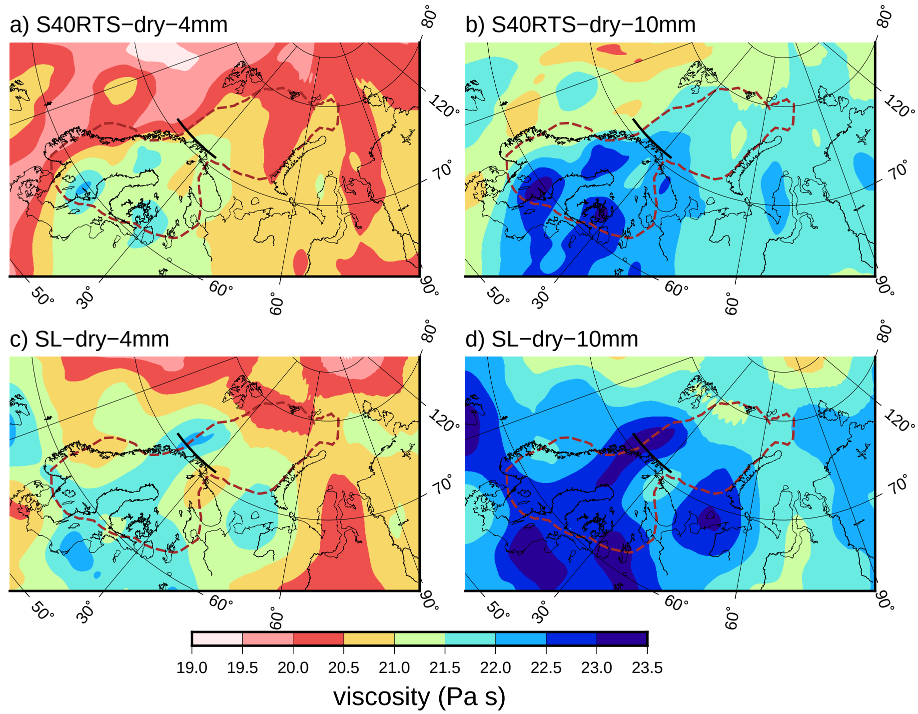


Figure 8. Viscosity between 225 and 325 km depth derived from seismic models S40RTS (Ritsema et al. 2011) (a and b), and Schaeffer and Lebedev (2013) (c and d), and for different flow law parameters: 4 mm grain size (a and c) and 10 mm grain size (b and d). The brown line denotes the 1500 m ice height contour at LGM in the ICE-5G model; the black line denotes 71° latitude which separates the areas used for computing the viscosity for Fennoscandia and the Barents Sea.