In the RAS Bullerwell Lecture for 2012, Matt A King outlines ways to observe and model glacial uplift, highlighting the need for comprehensive data collection and more sophisticated Earth models.

Progress in modelling and observing Antarctic glacial isostatic adjustment

The Last Glacial Maximum (LGM) came to an end about 20 000 years ago when the ice sheets of North America, Greenland, Fennoscandia and Antarctica, among others (Peltier 2004), began to melt. Deglaciation had largely ended by about 6000 years ago and global mean sea-level had consequently risen between 120 and 130 m (Fairbanks 1989, Peltier 2004). This redistribution of ice/ocean mass represented a major reorganization of Earth's surface loads that altered Earth's gravity field, changed its rotation pole and speed, and produced a viscoelastic response within the solid Earth that induced surface deformation (Mitrovica *et al.* 2009). The viscosity of Earth's mantle governs the response time of solid-Earth adjustments and is high enough that deformation is still measurable at the present time, perhaps most prominently through geodetic measurements of Earth surface displacements and satellite and terrestrial gravity-field measurements. Thus, present-day measurements of these parameters can give insights into past ice-sheet configurations and sea-level changes, as well as Earth structure and rheology (e.g. Milne *et al.* 2001, King *et al.* 2010). Constraints on the time-history of this deformation and mass redistribution are provided by glacial geology and paleo sea-level data.

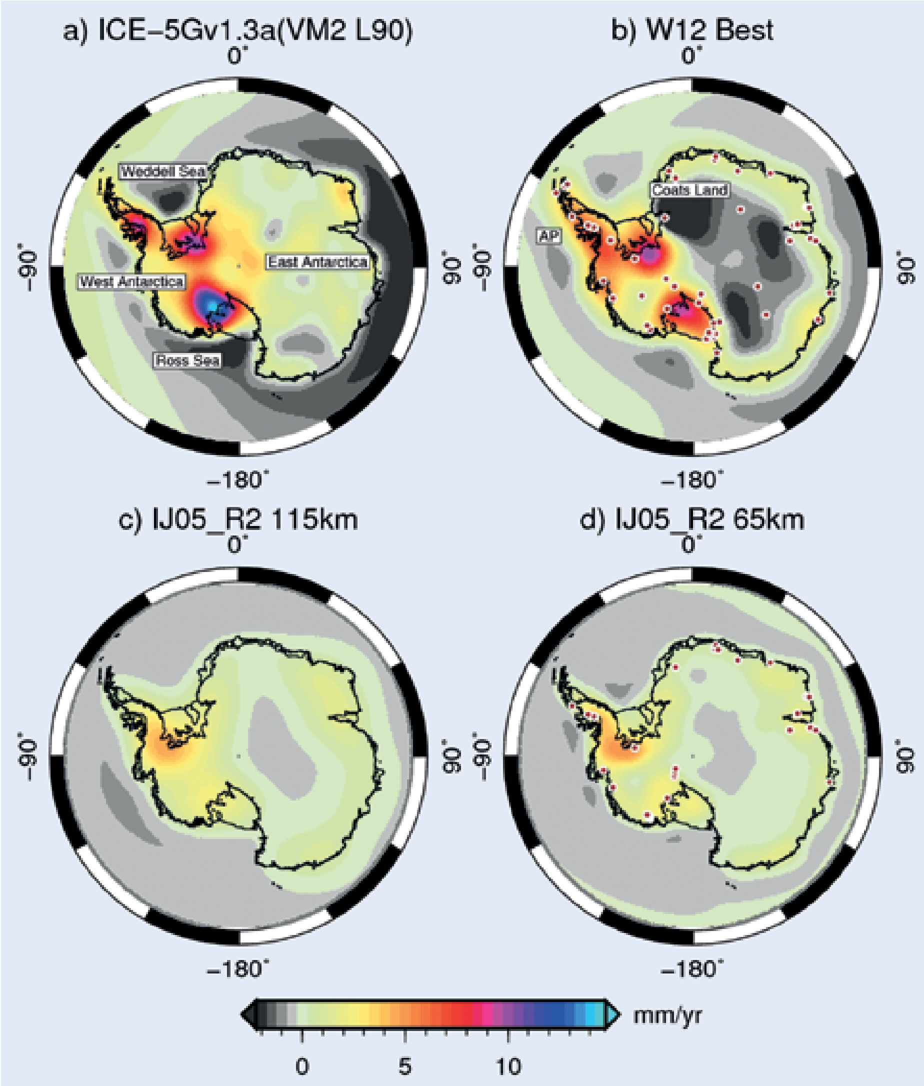
1: Field personnel deploying GPS and associated power equipment in the southern Antarctic Peninsula, including the antenna and monument (foreground) and solar/wind power system, met. station, batteries and GPS receiver (background). (Matt Burke)

Glacial isostatic adjustment (GIA) is the term most widely used to describe the response of the solid Earth to changes in ice-ocean surface loading. Generally speaking, since the LGM deglaciation, mantle material has flowed into regions that have undergone a decrease in surface loading (the continents, where ice melted) and away from areas of increased loading (the oceans, which gained more water). In reality the patterns of change are more complex, partly due to feedbacks between mass redistribution, gravity and rotation (Mitrovica and Wahr 2011, Mitrovica *et al.* 2005). Substantial theoretical advances have been made in understanding the processes governing GIA (e.g. Milne and Mitrovica 1998, Dalca *et al.* 2013, Farrell and Clark 1976, Mitrovica and Milne 2003), and previously discordant GIA model outputs based on the same input are now in close agreement, or have well-understood differences (Spada *et al.* 2011, Peltier *et al.* 2012, Chambers *et al.* 2010, Chambers *et al.* 2012).

However, our ability to model ongoing GIA accurately is also dependent on our knowledge of the spatiotemporal evolution of past ice sheets and the rheology and spatially varying structure of Earth. Of all the LGM ice sheets, Antarctica represents the largest challenge because the presence of the ice sheet means that powerful geomorphological/geological evidence for determining the amount and timing of post-LGM ice retreat or thinning is limited to remote and spatially limited rock outcrops or offshore regions.

GIA, and particularly Antarctic GIA, has received renewed interest in the past decade due to the Gravity Recovery and Climate Experiment (GRACE) satellite mission (Tapley *et al.* 2003). GRACE data allow the production of maps of Earth's gravity field every 10–30 days with spatial scales of ~400–750 km (Rowlands *et al.* 2005, Chambers 2006), from which mass redistribution may be inferred. One of GRACE's primary goals was to resolve the long-standing debate and uncertainty surrounding the contribution of the Antarctic ice sheet to present-day sea-level change (Velicogna and Wahr 2002). However, GRACE data alone cannot distinguish mass change within the solid Earth from that on it (ice, oceans) or above it (atmosphere). Errors in models of atmospheric and oceanic mass-change signals over Antarctica are significantly smaller than those of GIA and initial GRACE estimates of Antarctic ice-mass change were swamped by the uncertainty in modelling GIA (Velicogna and Wahr 2006); indeed, even the uncertainty of the GIA models was largely unknown and some authors elected to not quantify it (e.g. Chen *et al.* 2006). As a consequence, substantial efforts have since been made to improve our understanding of Antarctic GIA.

Here, I review (a) some of the recent developments in modelling Antarctic GIA; (b) their application to the problem of determining present-day ice-mass balance; (c) available observational constraints; and (d) some of the areas of outstanding need with regard to new observations and model developments. Those interested in an overarching and thorough review of GIA are referred to Mitrovica *et al.* (2009) while those interested only in the effects of GIA on the spatially varying pattern of present-day sea levels are referred to introductory reviews by Milne (2008) and Tamisiea and Mitrovica (2011).



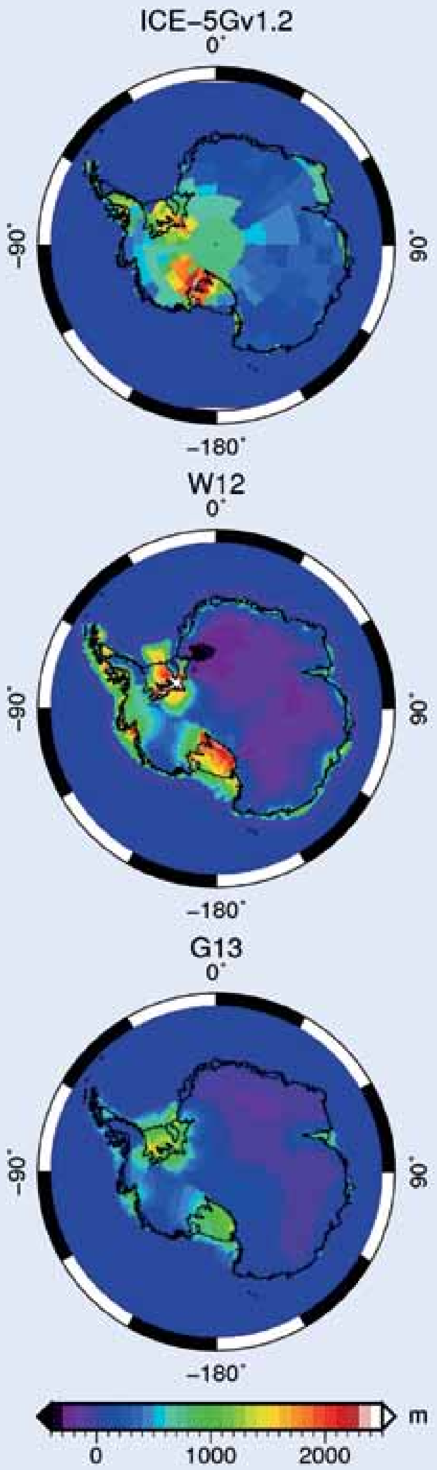
2: Present-day uplift rates due to glacial isostatic adjustment as predicted by four models. Black lines represent the present-day ice-sheet grounding line and ice-shelf fronts (Scambos *et al.* 2007). AP is Antarctic Peninsula. Brown circles represents onshore ice-extent constraints at (b) 20 kyr and (d) 5 kyr BP as tabulated by Whitehouse *et al.* (2012a).

Modelled Antarctic GIA

Figure 2 shows predictions of Antarctic present-day uplift rates from four recent GIA models. Figure 2a contains predictions based on the global ICE-5Gv1.3a ice history and the VM2 L90 radially symmetric Earth model (updated from Peltier 2004), hereafter referred to as ICE-5G. The ICE-5G ice history is constrained by global ice extent and relative sea-level data. Figures 2b–d show present-day uplift rates from models (W12 and IJ05_R2) that focus on Antarctica's ice history, replacing the Antarctic component of ICE-5G in GIA model computations. These models do not attempt to fit far-field sea-level data and hence should not be used for studies away from Antarctica where the related errors are minimized.

The W12 predictions (figure 2b; Whitehouse *et al.* 2012b) are based on a numerical model of the post-LGM Antarctic ice sheet (Whitehouse *et al.* 2012a) that attempts to fit Antarctic ice-extent data. This ice model is combined with a radially symmetric Earth model chosen to best fit Antarctic relative sea-level data. A variant of this model (W12a) has been created to address unmodelled late-Holocene thickening (last 1000 yrs) in the southern Antarctic Peninsula (Whitehouse *et al.* 2012b), but the unmodified version is shown here. Figure 2c shows a prediction from the IJ05_R2 model (Ivins *et al.* 2013), an update of IJ05 (Ivins and James 2005). IJ05_R2 ice history is based on a simpler approach, creating a set of ice-discs that attempt to fit a similar set of ice-extent data to W12. Extensive use of Global Positioning System (GPS) uplift rates was also made in constructing IJ05_R2; these GPS data are discussed later.

The differences between these three models



3: Thickness of the modelled LGM ice-sheet relative to modelled present-day in three different reconstructions. Black lines represent the present-day ice sheet grounding line and ice-shelf fronts.

(figures 2a–c) are striking in that the locations of signal maxima and their magnitudes vary substantially. Considering West Antarctica, ICE-5G predicts uplift rates centred on three domes with rates up to 16 mm/yr. The spatial pattern of W12 is relatively similar but with smaller maxima magnitudes (<11 mm/yr) and a different location for the dome of uplift in the southern Antarctic Peninsula. IJ05_R2 is, by contrast, much smoother and has much reduced uplift in the Ross Sea. In IJ05_R2 there is no more than 5 mm/yr uplift in the Weddell Sea/Antarctic Peninsula region.

In East Antarctica there is disagreement over the dominant sign of the uplift, with W12 predicting strong interior subsidence and coastal uplift, IJ05_R2 predicting less strong subsidence and ICE-5G predicting widespread uplift with isolated regions of subsidence. While the predicted signals are not as strong as those in West Antarctica, the much larger area of East Antarctica means that these differences represent a substantial uncertainty in modelling Antarctic GIA. Arguments for subsidence in the interior of East Antarctica stem from inferences of increased accumulation since the LGM, deduced from ice cores that cover this period (Lorius *et al.* 1984). Coupled with data that suggest a limited retreat of the ice margin (e.g. Mackintosh *et al.* 2011), these observations imply an ice-loading history that results in coastal uplift and interior subsidence (Whitehouse *et al.* 2012b, Ivins and James 2005, Ivins *et al.* 2013, Whitehouse *et al.* 2012a). Arguments for extensive East Antarctic uplift come from the need to fit the far-field sea-level data (Nakada and Lambeck 1988, Peltier 2004) and the limited spatial and temporal coverage of ice-extent data and, in some cases, uncertainty over their quality. A similar debate continues relating to the total ice volume coming from West Antarctica since the LGM (Bentley *et al.* 2010, Bentley *et al.* 2011, Clark 2011). A model that fits all local ice-extent data and the far-field sea-level data has thus far proved elusive and this remains an active area of research.

The effect of radially varying the Earth structure on GIA predictions is illustrated in the difference between figure 2c and 2d. These predictions were computed using identical IJ05_R2 ice histories, but figure 2c was produced using an Earth model with a lithospheric thickness of 115 km and upper and lower mantle viscosities of 2×10^{20} and 4×10^{21} Pas, respectively, whereas figure 2d was produced using values of 65 km, 2×10^{20} and 1.5×10^{21} Pas, respectively (Ivins *et al.* 2013). The weaker Earth model produces subtly stronger and more spatially concentrated uplift rates.

Seismic studies suggest a significant difference in Earth structure between East Antarctica and West Antarctica, with expectation of a thicker lithosphere in East Antarctica (Ritzwoller *et al.* 2001). Compared to the rest of West Antarctica, the Antarctic Peninsula has a much thinner lithosphere and lower mantle viscosity associated with ancient plate subduction in this region (Simms *et al.* 2012, Ivins *et al.* 2011, Yegorova *et al.* 2011). However, all of the models shown in figure 2 use a radially symmetric (one-dimensional) Earth structure and hence cannot reflect these spatial deviations. To date, the most common approach to deal with this limitation is the creation of multiple realizations of the GIA model with each considering a different 1D Earth model (Ivins and James 2005, Ivins *et al.* 2013, Whitehouse *et al.* 2012b). Lack of treatment of 3D Earth structure has been justified by the historical dominance of ice-history uncertainties on GIA model predictions.

Ice-sheet models

The most widely used GIA models include ice-sheet reconstructions that are not based on full numerical ice-sheet models, but there is growing evidence that such models offer some advantages over traditional (semi-)manual disc-based reconstructions. While numerical ice-sheet models have been used to reconstruct the post-LGM Antarctic ice sheet since the 1990s (Budd *et al.* 1998, Huybrechts *et al.* 2004), use of such models has attracted renewed interest recently due to recent advances in ice-sheet model realism and availability of community ice-sheet models (e.g. Rutt *et al.* 2009). The realism of a reconstructed ice sheet using such models depends on the completeness of the model physics, choice of parameterization and the accuracy of boundary conditions and input forcing data, including accumulation, relative sea level and ice–ocean interactions. The model physics ensures that the ice-sheet shape is physically reasonable, something that is not always exhibited in disc-based reconstructions. However, some degree of parameter tuning (e.g. to define the rate at which ice is allowed to slide over the bed) is required to fit ice-extent data (Whitehouse *et al.* 2012a) and there is likely to be more than one ice history that will fit the presently available observational constraints. This latter fact is useful in understanding ice-model uncertainties as discussed below.

Figure 3 shows LGM ice thickness relative to present day for three models: ICE-5G and the output of two recent and independent numerical ice model reconstructions. The latter two are an updated version of the model of Golledge *et al.* (2012), denoted here as G13 (Golledge *et al.* in review) and W12. Other forward models exist (Huybrechts 2002, Pollard and DeConto 2012, Briggs and Tarasov 2013), but these are shown as convenient and recent examples. In East Antarctica, thickness-changes predicted within W12 and G13 are in broad agreement, with both showing substantial regions where the LGM ice sheet was lower than present-day, although the spatial extent and magnitude of this is larger in W12. W12 also suggests a greater extension of the ice sheet onto the continental shelf at LGM, whereas G13 has limited extension from present-day with the notable exception of the Prydz Bay region. One important difference between the W12 and G13 models is that the W12 ice extent was forced to fit sparse offshore data, whereas G13 uses a freely evolving grounding line.

Table 1: Variations in estimates of Antarctic ice mass change

study	data span	GIA model	all Antarctica	West Antarctica	East Antarctica
King *et al.* (2012)	Aug. 2002 to Dec. 2010	W12a	−69±18	−118±9	+60±13
Shepherd *et al.* (2012)	Jan. 2003 to Dec. 2010	IJ05_R2+W12a	−81±66	−107±54	+56±76
Ivins *et al.* (2013)	Jan. 2003 to Jan. 2012	IJ05_R2	−57±68	–	–
Velicogna and Wahr (2013)	Jan. 2003 to Nov. 2012	IJ05_R2	−83±98	–	–

Estimated ice mass change (gigatonnes per year; Gt/yr). All uncertainties are specified as 2-sigma, assuming original 1-sigma uncertainties in Ivins *et al.* (in press) and Velicogna and Wahr (2013). Uncertainties of the King *et al.* (2012) estimates do not include GIA model uncertainties discussed in the main text. Estimates in Gt/yr may be converted to equivalent global-mean sea-level contribution in mm/yr by dividing by −362; these estimates suggest Antarctica has been contributing approximately 0.18–0.22 mm/yr over the data periods.

In West Antarctica the Weddell and Ross Sea regions show post-LGM thinning, although W12 shows substantially more. W12 also suggests >1500 m extra ice over the Amundsen Sea region and Antarctic Peninsula at the LGM, approaching double the thickness of that in G13 for the same regions. Indeed G13 predicts very little change in the southern Antarctic Peninsula. As already discussed, ICE-5G has the opposite sign in East Antarctica, but in West Antarctica the pattern of change more closely resembles W12 but with important differences at the level of hundreds of metres of ice thickness. The IJ05_R2 disc-based reconstruction shows a different pattern again (Ivins *et al.* 2013, their figure 3d). Overall, the differences between W12 and G13 suggest there is substantially more work to be done before the output of numerical ice-sheet models is fully understood.

Unlike Earth-model uncertainties, ice-model uncertainties are not generally produced by GIA modellers due to the deterministic manner in which the ice models are constructed. Most GRACE-based studies of ice-mass balance, for instance, resort to model-differencing to obtain GIA uncertainties representative of both the Earth and ice models (e.g. Shepherd *et al.* 2012, Velicogna and Wahr 2006). The W12/W12a model is currently the exception to this rule, in that it provides lower and upper bounds on predicted GIA uplift and geoid rates, generated using a suite of ice models used in turn in the GIA model computation. This approach was further developed by King *et al.* (2012), who considered conservative bounds on W12a geoid rates before applying them to GRACE time series. Briggs and Tarasov (2013) performed an investigation into Antarctic ice-extent data uncertainties and their implementation in ice-sheet models and these could, in the future, be propagated to GIA predictions.

GIA and GRACE

The models shown in figure 2 have each been applied to GRACE data in order to determine Antarctic ice-mass balance, with estimated trends and 2-sigma uncertainties listed in table 1. The first three studies listed use GRACE release 4 (RL04), whereas Velicogna and Wahr (2013) use RL05 and they note that RL05 ocean and atmosphere models that are subtracted from GRACE data are sufficiently different from those provided within RL04 to explain the difference between the results of Ivins *et al.* (2013). The uncertainties on the King *et al.* (2012) estimates are substantially smaller than the other estimates because GIA and other potentially systematic errors were expressed as systematic bounds in that study; the presented uncertainties relate just to the uncertainty of fitting a linear term to the ice-mass change time-series. The bounds on their estimate, largely due to W12a GIA error bounds, were [−126, −29] Gt/yr, with West Antarctica having a smaller range due to a smaller GIA uncertainty [−128, −103] Gt/yr ,and East Antarctica a larger range [+7, +89] Gt/yr.

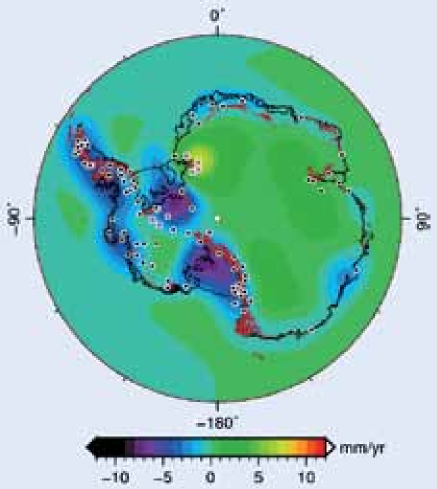
The four separate GRACE estimates, based on two different and independent GIA models, give values in close agreement, although with large uncertainties or bounded errors. We note that the four studies have small differences in data spans which will introduce some real differences in mass loss since nonlinear changes in mass have been observed in some regions (Velicogna 2009, Chen *et al.* 2009). Given their similar overall equivalent sea-level contribution since the LGM, application of other recent ice histories (e.g. G13) in GIA models would likely produce magnitudes of ice-mass change that are similar to the three studies summarized here (Ivins *et al.* 2013); they all produce rates of mass-loss systematically smaller by 60–90 Gt/yr than studies based on earlier GIA models such as ICE-5G.

Despite this encouraging agreement, and agreement with independent non-GRACE estimates (Shepherd *et al.* 2012), the bounds on East Antarctic ice-mass change mean that it is not possible to assess confidently if this region is slightly losing or substantially gaining mass at present. Further, the substantial differences in spatial detail in ice-sheet reconstructions (figure 3) and GIA models (figure 2) that are especially evident in West Antarctica suggest that spatial patterns of ice-mass change from GRACE are far from certain. As explained, these uncertainties stem from uncertainties in Earth and ice models and these results strongly suggest that new and spatially extensive observations are required to constrain ice and Earth models.

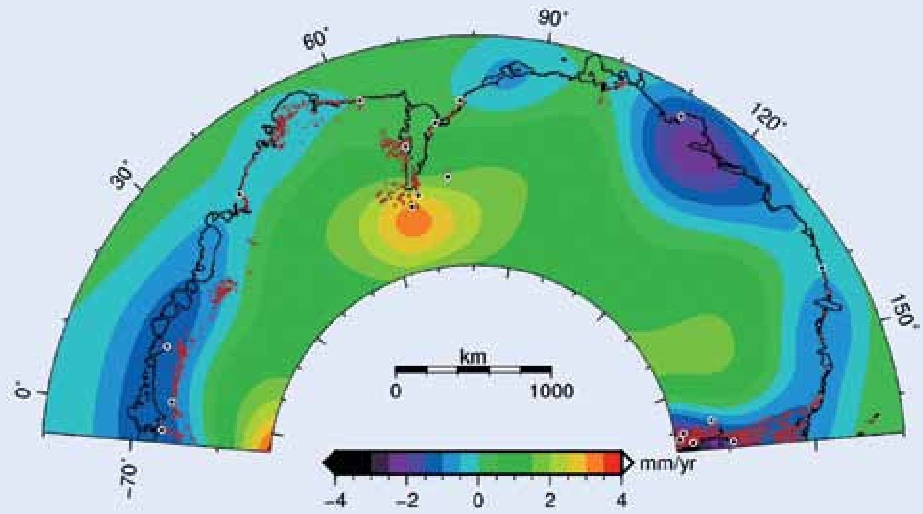
Observations and GIA modelling

Given the continuing dominance of uncertainty in ice history, the most valuable data types for constraining GIA modelling are those that provide dated limits on ice extent, such as cosmogenic exposure dating of rock outcrops (Bentley 2010), ice cores (Lorius *et al.* 1984) and marine geophysics (Cofaigh *et al.* 2008, Livingstone *et al.* 2012). Figure 2b also shows locations of existing LGM ice-extent data as tabulated by Whitehouse *et al.* (2012a); figure 2d shows the locations for the 5 ka Before Present (BP) time slice. A similar database has been established by Briggs and Tarasov (2013). Examination of figure 2b reveals that the limited data coverage available for the LGM ice sheet is further reduced at subsequent times, as illustrated by the 5 ka time slice (figure 2d). Offshore, high-resolution ("multibeam") bathymetric surveys further constrain previous ice flow and retreat, but such surveys are quite rare in East Antarctica, where most of its coastline is covered by just one multibeam bathymetric profile (Arndt *et al.* 2013).

The limited data relating to ice-sheet changes in the last few thousand years is particularly concerning because GIA decays with time, and hence present-day GIA is highly sensitive to recent loading changes. In the Antarctic Peninsula, where mantle viscosity is relatively low, much of the present-day GIA may be governed by quite recent ice-load changes (Simms *et al.* 2012, Ivins *et al.* 2000). Indeed, increases in ice loading by several tens of metres have occurred in the Antarctic Peninsula since the 1850s with potentially important levels of subsidence being induced (Nield *et al.* 2012), while at the same time low-elevation glaciers have retreated and thinned (Cook *et al.* 2005, Kunz *et al.* 2012). Constraints on the period before that are poor (Ivins *et al.* 2011). However, the importance of late-Holocene changes extend to other regions, as illustrated by the substantial subsidence



4: Difference in predicted present-day uplift rates (IJ05_R2 minus W12). The version of IJ05_R2 with a 115km thick lithosphere is used. Black circles indicate known existing continuously operating GPS sites and/or sites with uplift rates published by Thomas *et al.* (2011). Magenta squares indicate known funded GPS sites not yet installed at the time of writing. Brown lines indicate rock outcrops. Black lines represent the present-day ice-sheet grounding line and ice-shelf fronts.



5: As for figure 4 but showing detail in East Antarctica. Note the different colour scale.

predicted by W12 in Coats Land (figure 2) – this is due to the modelled ice-sheet thickening over the last few thousand years. There is not yet ice-extent data to test this (figure 2d).

Observations of ongoing uplift rates using GPS or similar techniques can constrain GIA models, or be used to validate them, although these data are ambiguous in that differences from modelled uplift rates may be due to errors in ice or Earth models or both. The different spatial patterns of such errors means that they can be substantially decoupled using observations with a high spatial density (Milne *et al.* 2001) and this is possibly feasible in West Antarctica where rock outcrops are relatively prolific.

Several Antarctic GPS uplift-rate datasets have now been published (Argus *et al.* 2011, Bevis *et al.* 2009), with the most comprehensive to date being that of Thomas *et al.* (2011). That study compared the uplift rates with predictions from ICE-5G and IJ05 (Ivins and James 2005) and found that the GPS rates were universally lower near to modelled signal maxima. This finding, together with new glacial geology data, contributed to the revision of IJ05 (Ivins *et al.* 2013) shown in figure 2. Comparing the same uplift rates to the W12 and W12a models, Whitehouse *et al.* (2012b) found a factor 2 reduction in misfit compared to the earlier models, suggesting a substantial improvement in model accuracy.

It is important to note that comparison of GIA models with GPS uplift rates first requires consideration of (i) elastic effects due to present-day ice mass changes and (ii) the difference in the definition of reference-frame origin between the GPS velocities and GIA models. Thomas *et al.* (2011) provided maps of elastic rebound computed using two different present-day ice-mass change fields and found that the effect on their site distribution was small, with most sites being well away from regions of large present-day mass change. The exception was the Antarctic Peninsula, where rapid uplift was identified following the Larsen B Ice Shelf breakup and subsequent rapid glacier mass-loss into the oceans (Scambos *et al.* 2004). There was suggestion that the rapid uplift may have been driven by elastic effects only, but subsequent elastic modelling that expanded the load change to higher spherical harmonic degree shows the elastic signal to be more spatially concentrated than that presented by Thomas *et al.* (2011) (R Riva, personal communication, 2011; Spada *et al.* 2012). Very high GPS uplift rates have also been identified in the Amundsen Sea region (Groh *et al.* 2012), and it is not yet clear if these are purely elastic or have a viscous component. The latter would imply low-viscosity mantle and/or long-term mass loss in this region.

Figure 4 shows the locations of known continuous GPS sites, either installed or funded, overplotted on the difference between W12 and IJ05_R2. The current distribution of GPS sites is partly governed by rock outcrop availability, but the dominant factor in their distribution is actually logistics provision, especially regarding access to remote regions of East Antarctica where inter-model differences are easily measurable by GPS (figure 5). Co-location of geological sampling and GPS sites represents both efficient use of logistics and powerful joint constraints on GIA models.

While GRACE observations have been used outside Antarctica to constrain GIA (e.g. Hill *et al.* 2010), the presence of the Antarctic ice sheet precludes this approach. One exception is the application of GRACE data offshore, including over the floating ice shelves, where mass changes are expected to be small. GIA model errors have been identified using this approach (King *et al.* 2012, Sasgen *et al.* 2007), including the observation of little ongoing GIA under the Ross Ice Shelf (King *et al.* 2012), in disagreement with the predictions of W12, ICE-5G (figure 2a–b) and older models. This could be due to recent loading changes, very light grounding of the extensive LGM ice sheet in this region, or limitations in 1D Earth models, among others, and requires further investigation.

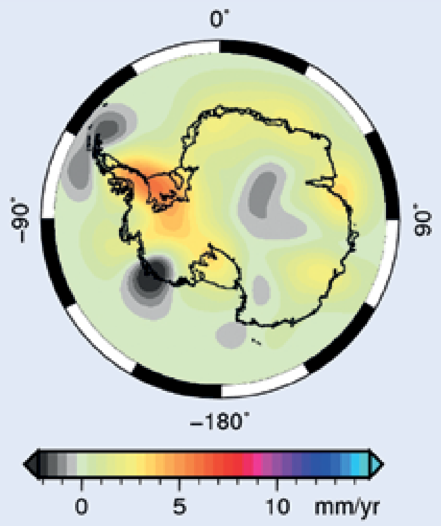
Ground-based absolute gravimetry is a powerful technique, but it is yet to produce substantial results (Amalvict *et al.* 2009, Makinen *et al.* 2007). However, the combination of geodetic datasets has been used to produce spatially continuous empirical estimates of present-day uplift (Riva *et al.* 2009, Sasgen *et al.* 2007, Wu *et al.* 2010). Thomas *et al.* (2011) found relatively close agreement between the estimate of Riva *et al.* (2009) and GPS uplift rates. Comparing figure 6 with figure 2 reveals remarkable similarities between IJ05_R2 uplift rates and those estimated by Riva *et al.* (2009), suggesting this pattern of present-day uplift may be relatively robust.

Discussion and conclusions

The focus in this brief review has been on the dominant error-source in present-GIA models: ice history. Greater understanding of the appropriate Earth model will be required as uncertainties of Antarctic ice history are further reduced. This will include moving beyond 1D Earth models to more fully represent the laterally heterogeneous structure present beneath Antarctica. This is presently a challenge due to the absence of appropriate regional datasets. Preliminary 3D-modelling studies have been conducted, with the conclusion that considering lateral structure either produces major (Kaufmann *et al.* 2005) or moderate (A *et al.* 2012) differences, depending on the assumed structure. Importantly A *et* *al.* (2012) suggest that failure to consider 3D structure may substantially affect the tuning or validation of 1D GIA models with GPS uplift data (e.g. Ivins *et al.* 2013, Whitehouse *et al.* 2012b). Application of 3D models importantly means that horizontal GPS velocities, with a precision 2–3 times that of vertical velocities, will become useful for the first time in constraining Antarctic GIA models. The extensive collection of seismic data within the POLENET project (http://www.polenet.org) should soon yield new information to allow 3D scenarios to be examined with greater confidence of realism. While structure inferred from seismic waves is not necessarily directly applicable at GIA timescales, such data are critical for the setup of 3D Earth models (Whitehouse *et al.* 2006).

In contrast to the time-dependent rheologies now widely used in Earth modelling in other disciplines (Wang *et al.* 2012), all widely used GIA models to date adopt a linear Maxwell rheology. While the GIA modelling community has explored the concept that mantle viscosity is transient (Sabadini *et al.* 1985), little attention has been paid to it since linear rheologies were identified that satisfied large-scale geophysical constraints (e.g. Nakada and Lambeck 1989). However, transient rheology has recently been raised again (Morrow *et al.* 2012, van der Wal *et al.* 2013) and this uncertainty needs to be resolved. Such time-dependent effects may be important, for example, in conjunction with late-Holocene ice-loading changes in West Antarctica and the Antarctic Peninsula.

Figures 2 and 3 demonstrate that recent GIA models and ice-sheet reconstructions represent a significant revision to earlier studies. They have not yet, however, reached the stage of being definitive, as is shown by the substantial inter-model differences that exist and the dearth of observations with which to constrain and/or validate the models. The new models are, however, in general agreement with ice-extent data, where they exist, that imply a relatively low contribution to post-LGM sea level from Antarctica – of the order of 8–12 m (Ivins *et al.* 2013, Bentley *et al.* 2010); the means of closing the paleo sea-level budget with a reduced Antarctic contribution is an ongoing debate. To address the remaining Antarctic GIA model uncertainty, new spatially extensive observations of ice extent (onshore and offshore) that cover all periods following the LGM are required (figures 2b, d), as are new bedrock uplift data (figures 4 and 5). Geological and GPS data are required in the coastal margin, on offshore islands and further inland. Ice-sheet reconstruction studies adopting numerical ice-sheet models require further advances in understanding paleo accumulation and ocean temperature, and sea-level forcing (Gomez *et al.* 2012, Mackintosh *et al.* 2011). Given GPS sites require near-annual maintenance visits, their deployment requires substantial investment, but the need for GPS and geological data in remote regions suggests that this investment is necessary. For substantial regions of the deep interior of East Antarctica, where no rock outcrops exist, new ways of constraining GIA and ice history are required (e.g. Siddall *et al.* 2012). ●



5: Empirical present-day uplift rate estimates of Riva *et al.* (2009)

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Acknowledgments. Nick Golledge, Pippa Whitehouse and Erik Ivins kindly supplied model output. ICE-5Gv1.3a grids were helpfully made available by Dick Peltier at http://www.atmosp.physics.utoronto.ca/~peltier/data.php. I thank the BGA for the invitation to write this paper associated with a Bullerwell Lecture at the EGU in 2012. Christopher Watson and Pippa Whitehouse provided helpful comments on a draft of the manuscript. Matt King is a recipient of an Australian Research Council Future Fellowship (project number FT110100207).
● *The Bullerwell Lecture is awarded annually to an outstanding young British geophysicist, by the British Geophysical Association (http://www.geophysics.org.uk)*

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