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Glacial isostatic adjustment of the British Isles: new constraints from GPS measurements of crustal motion

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SUMMARY

We compared estimates of crustal velocities within Great Britain based on continuous global positioning system (CGPS) measurements to predictions from a model of glacial isostatic adjustment (GIA). The observed and predicted values for vertical motion are highly correlated indicating that GIA is the dominant geodynamic process contributing to this field. In contrast, motion of the Eurasian plate dominates the horizontal motion component. A model of plate motion was adopted to remove this signal in order to estimate intraplate horizontal motion associated with GIA. However, a coherent pattern of horizontal motion was not evident in the resulting velocity field. We adopted a recently published model of the British–Irish ice sheet to predict vertical crustal motion for a large number of spherically symmetric Earth viscosity models. Our results show that the adopted ice model is capable of producing a high-quality fit to the observations. The CGPS-derived estimates of vertical motion provide a useful constraint on the average value of viscosity within the upper mantle. Values of model lithospheric thickness and lower mantle viscosity are less well resolved, however. A suite of predictions based on an alternative ice model indicates that the vertical motion data are relatively insensitive to uncertainties in the ice loading history and so the constraints on upper mantle viscosity are robust.

Key words: Satellite geodesy; Plate motions; Dynamics of lithosphere and mantle.

1 INTRODUCTION

The on-going viscous response of the solid Earth to the melting of the last great ice sheets is currently a significant contributor to intraplate deformation around the globe, particularly in previously glaciated regions such as Canada and NW Europe. In these regions, this geodynamic process, more formally known as glacial isostatic adjustment (GIA), dominates the solid Earth deformation field (e.g. Johansson *et al.* 2002; Sella *et al.* 2007). A number of studies have considered the GIA of these previously glaciated (so-called 'near-field') regions through the measurement and interpretation of relative sea-level observations (e.g. Tushingham & Peltier 1991; Lambeck 1993a,b; Davis *et al.* 1999; Mitrovica 1996; Lambeck *et al.* 1998) and, more recently, space geodetic data (e.g. Milne *et al.* 2001; Tamisiea *et al.* 2007). Studies such as these have provided important constraints on both past ice extent and Earth viscosity structure.

There is a long history of sea-level observation in the British Isles (e.g. Sissons 1966; Sissons & Brooks 1971; Tooley 1974, 1982; Dawson 1984; Shennan *et al.* 1995; Selby & Smith 2007) and this has lead to the development of a high-quality and extensive regional data base (Shennan *et al.* 2006a). This data base has been employed to determine empirical estimates of GIA-induced land motion (Smith *et al.* 2000; Dawson *et al.* 2002; Fretwell *et al.* 2004) as well as constrain parameters in geophysical models of the GIA process (Lambeck 1993a,b, 1995; Lambeck *et al.* 1996; Shennan *et al.* 2000, 2002, 2006b; Peltier *et al.* 2002; Milne *et al.* 2006). While there has been a dramatic advance in the development of GIA models for the British Isles since the early 1990s, there remain significant differences in key parameters inferred by different groups and the model fits to the ever-expanding data base have been poor in certain areas (e.g. Shennan *et al.* 2006b). These discrepancies reflect the constant emergence of new geomorphological constraints on ice sheet extent and thickness as well as the high-degree of non-uniqueness inherent to modelling near-field GIA (Shennan *et al.* 2005, 2006b). This non-uniqueness is particularly acute for the British Isles given the relatively small size of the local ice sheet which results in the local isostatic component of the sea-level signal being of similar magnitude to that associated with the melt-water (eustatic) contribution from non-local ice sheets (e.g. Lambeck 1993b).

The application of the global positioning system (GPS) to monitor secular variations of the solid Earth has provided a new data set that can be applied to the GIA problem. The main advantages of GPS measurements over those of past sea-level changes is that a direct measurement of present-day vertical and horizontal crustal motion is recorded at each site and spatial sampling is not limited to coastal areas.

Results obtained from a regional GPS network in Fennoscandia demonstrated that continuous GPS (CGPS) was able to resolve both vertical and horizontal crustal motion at sub-millimetres per year precision (Johansson *et al.* 2002; Lidberg *et al.* 2007). Subsequent modelling studies have demonstrated that the Fennoscandian CGPS data provide useful constraints on GIA models that complement those imposed by the RSL observations and thereby reduce the level of non-uniqueness in inferring model parameters (Milne *et al.* 2001, 2004).

A CGPS network was first established across Great Britain in the late 1990s and there are currently around 130 receivers in operation (BIGF, 2007). The position time series for a subset of these receivers are now long enough to constrain 3-D crustal motion at the sub-millimetres per year level (Teferle *et al.* 2006; Milne *et al.* 2006). A primary aim of this study is to examine the CGPS-derived 3-D velocity field for Great Britain and to quantify the component of the motion associated with GIA. In addition, we aim to determine if the CGPS data will be useful in reducing the non-uniqueness of the British Isles GIA problem outlined above. Note that this study extends that of Milne *et al.* (2006) by: (1) considering a GPS solution based on longer time series and improved processing techniques; (2) adopting a recently improved British–Irish ice model that is consistent with a subset of the regional RSL data base as well as a variety of new geomorphological information on ice extent (Shennan *et al.* 2006b); and (3) by performing a much more complete Earth model parameter study (243 models compared to 7 in Milne *et al.* (2006)). In Section 2 we briefly introduce the CGPS data employed in this study and we present our modelling results in Section 3.

2 CGPS DATA

The majority of CGPS stations in Great Britain were not deployed to monitor secular motions of the solid Earth and so most of the 130 sites now in operation are not suitable for this purpose. In a companion paper (Teferle *et al.* 2008), 44 of the 130 sites were initially considered after rejecting real-time kinematic (RTK) sites with time series less than 4 yrs in duration (non-RTK sites were not subjected to this criterion). This number was reduced to 24 after checking for consistency of results between two different processing strategies (sites were rejected when the discrepancy was greater than 1 mm yr^{-1}). For this analysis, we begin with these 24 stations and reduce the number further by applying the 4-yr time series criterion to all sites and removing stations at which: (i) the observed velocity is clearly not consistent with surrounding stations and therefore suspect, (ii) the data quality is poor, or (iii) monumentation is unsuitable. After applying these criteria, the number of sites considered was reduced to 16 sites for vertical motion and 21 sites for horizontal motion (Fig. 1a; Table A1). Despite this reduction in the number of sites, there remains an adequate spatial sampling of crustal motion across Great Britain.

The CGPS data were processed using the precise point processing (PPP) strategy (Teferle *et al.* 2007, 2008) implemented in the Bernese GPS software (version 5.0, Astronomical Institute of the Univ. of Bern, Bern, Switzerland) (Dach *et al.* 2007), to produce time series of daily position estimates for which the site velocities were obtained from a model-fit using maximum likelihood estimation (e.g. Williams 2003, 2008). Although all velocity errors are given to 1σ, these are of considerably larger magnitude, due to the proper handling of correlated noise in the time series during the model-fit, than those that would be obtained if the correlations were ignored. (Teferle *et al.* 2008). When processing the CGPS data to produce an estimate of vertical crustal motion, a systematic bias can be introduced due, largely, to uncertainty in the adopted terrestrial reference frame (see Teferle *et al.* 2008 for greater detail and plots of all position time series). To address this issue and attempt to improve the accuracy of the data, measurements of absolute gravity have been used to calibrate the CGPS-derived vertical rates (e.g. Teferle *et al.* 2006). Comparison of CGPS-derived vertical crustal motion to that inferred from absolute gravity measurements at Lerwick and Newlyn indicates that the CGPS rates are biased too high by around 0.5 mm yr^{-1} (Teferle *et al.* 2008). This result is somewhat tentative, however, given the small number of gravity measurements available in this region.

In the following section, we estimate model parameters using site-differenced rates for two reasons: (i) it removes, to a large extent, uncertainty associated with the reference frame and (ii) it isolates the signal associated with the BIIS (see next section). We chose to calculate motion relative to Sheerness (18 in Fig. 1a), as this site offers one of the longest observation periods (since March 1997) and the time series data are of high quality (Teferle *et al.* 2008). Since we model relative rates, the difference between the absolute gravity aligned and non-aligned versions of the CGPS vertical motion data is of no consequence.

Figure 1b illustrates the pattern of absolute vertical motion for the aligned CGPS data (with associated error) across Great Britain. Note that the non-aligned data produce the same pattern but with an increase in rates of 0.56 mm yr^{-1} at each site. The pattern, which comprises a region of maximum uplift centred over eastern Scotland and a zone of subsidence throughout most of England, is clearly correlated with the thickness extent of the most recent British–Irish ice sheet (BIIS) (e.g. Shennan *et al.* 2006b and references therein). The maximum uplift velocity is 1.07 ± 0.35 mm yr^{-1} at EDIN (site number 5) and the maximum subsidence velocity is 1.2 ± 0.40 mm yr^{-1} at LOWE (site number 13). The pattern of vertical motion estimated from the AG-aligned CGPS time series is broadly consistent with estimates of land motion obtained from late Holocene RSL observations (Shennan & Horton 2002; Teferle *et al.* 2008).

The horizontal velocity field (Fig. 1c) is characterized by motion directed towards the NE, with a magnitude of $\sim$15 mm yr^{-1}, and an average error of 0.29 mm yr^{-1}. This signal is dominated by motion of the Eurasian plate relative to the chosen reference frame. So, for each CGPS site, a modelled value of plate motion, estimated from the published ITRF2000 data set, was subtracted from the observed data to try and isolate a regional intraplate deformation pattern. The resulting velocity field (Fig. 1d) shows poor spatial coherence and the scatter in the rates is near the level of data precision.

The horizontal velocity field due to GIA predicted for this region is dominated by the deglaciation of the Laurentide ice sheet that covered most of North America at the last glacial maximum (e.g. Milne *et al.* 2006). The predicted signal is of magnitude 1–2 mm yr^{-1} directed predominantly NW (e.g. see Fig. 6 of Milne *et al.* (2006)). The isostatic component of horizontal motion associated with the BIIS—which has a radial geometry centred on western Scotland with a magnitude of $\sim$0.1 mm yr^{-1}—adds a minor perturbation to the Laurentide signal. We considered spatially differencing horizontal rates relative to a chosen reference station in order to isolate the BIIS signal. However, in contrast to a previous study (Milne *et al.* 2006), a correlation between the predicted and observed strain field was not apparent. We conclude, therefore, that

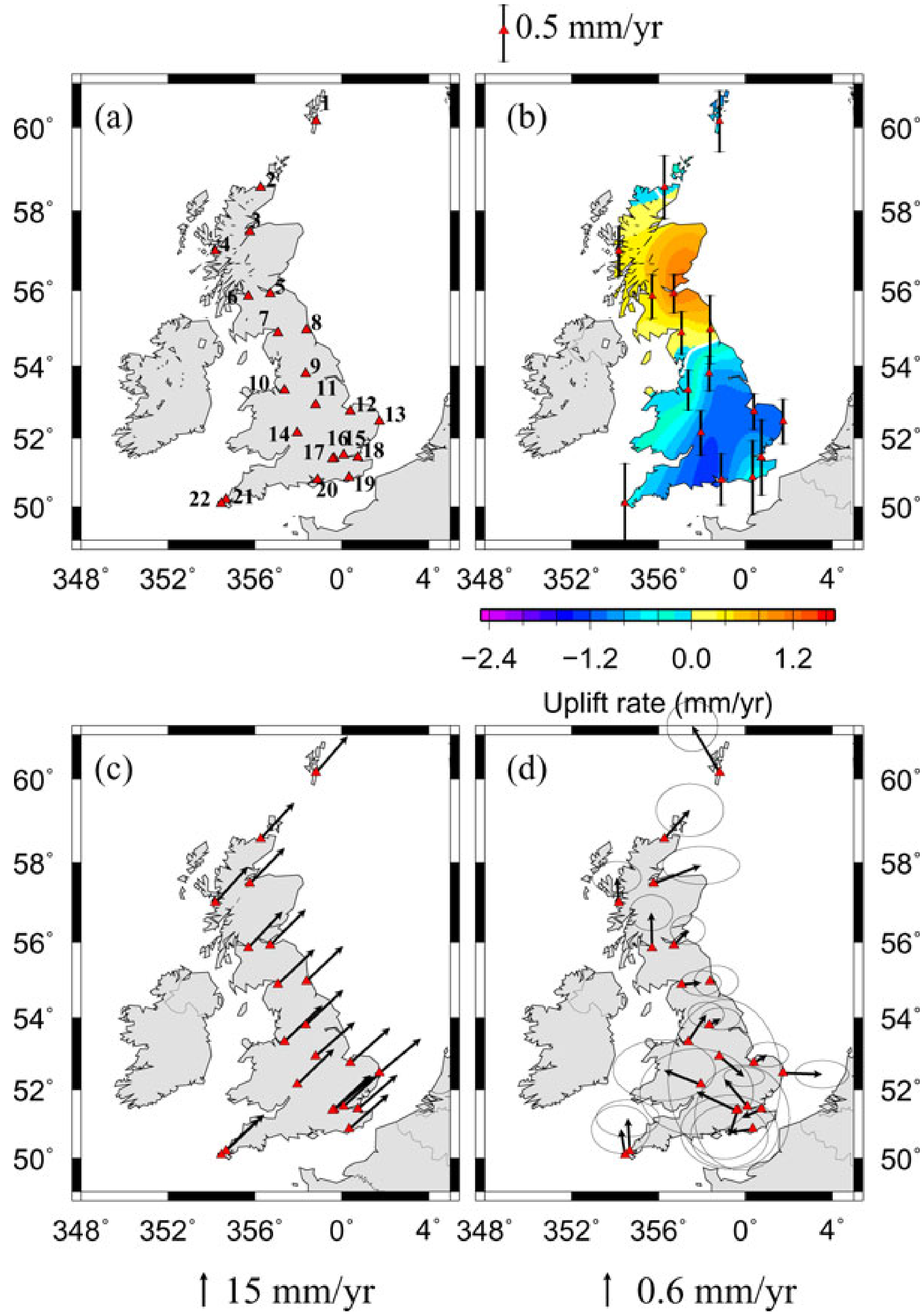

Figure 1. (a) Locations of GPS sites considered in this analysis. (b) Contour map of absolute vertical rates over the UK region interpolated from CGPS data for data aligned to absolute gravity. The error bars show the uncertainty at each of the 16 data sites. (c) Horizontal velocity vectors estimated from the GPS time series. (d) Same as (c) but with velocities from plate motion model removed. All uncertainties shown represent the 1σ range and are based on a post-processing scaling of the formal least squares values by a factor of 3 or more at most sites (see Teferle *et al.* 2008 for further information).

the current level of precision is not able to resolve a coherent pattern of horizontal motion associated with GIA. For this reason, the horizontal data will not be used in the remainder of this study. We will revisit the application of horizontal data in the future, when we would hope for a better correlation through the use of longer time series and a re-processing of the CGPS data, carried out using new and improved GPS satellite orbit and clock products (Steigenberger *et al.* 2006).

3 MODELLING RESULTS

The GIA model adopted in this study has three key inputs—a model of the Late Pleistocene ice history, an Earth model to reproduce the solid earth deformation resulting from surface mass redistribution (between ice sheets and oceans), and a model of sea-level change to calculate the redistribution of ocean mass (e.g. Farrell & Clark 1976).

We adopt a recently developed ice model for the UK and North Sea that was constrained using both RSL and geomorphological data (Shennan *et al.* 2006b). This regional model is patched into a global model of ice extent that provides an optimal fit to equatorial sea-level records (Bassett *et al.* 2005). The Earth model is a spherically symmetric, self-gravitating Maxwell body, with the elastic and density structure taken from a seismic model (Dziewonski & Anderson 1981) with a depth resolution of 10 km within the crust and 25 km in the mantle. The viscous structure within the mantle is more crudely parameterized into three layers: a high viscosity (10^{43} Pa s) outer shell to simulate an elastic lithosphere, an upper mantle region of uniform viscosity extending from beneath the model lithosphere to the 660 km seismic discontinuity, and a lower mantle region extending from this depth to the core–mantle boundary. The thickness of the lithosphere and the viscosity within the upper and lower mantle are free parameters in the modelling. The initial reference earth models used is taken from 'Shennan *et al.* (2006b)', with a lithosphere thickness of 71 km, and an upper mantle

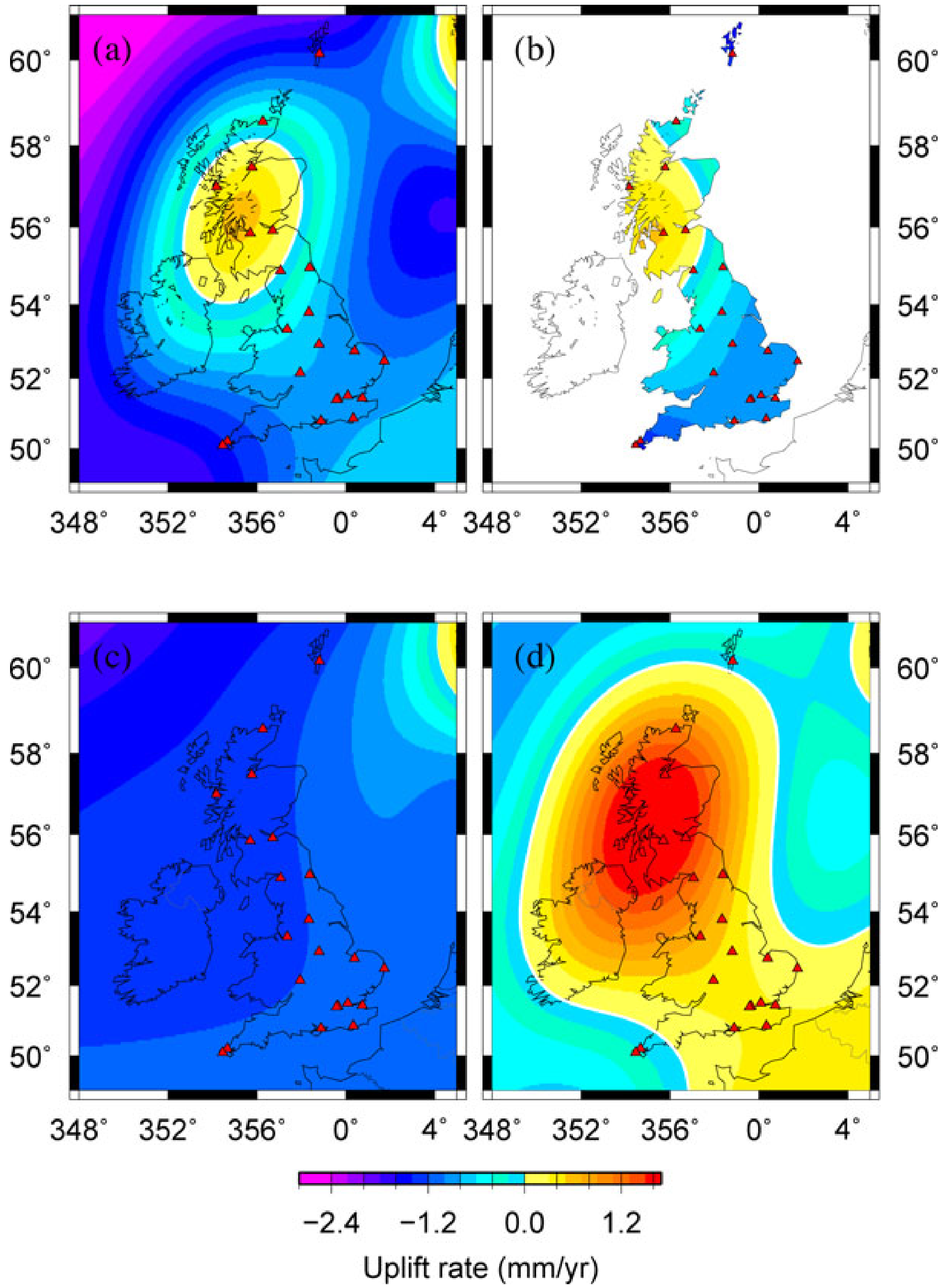

Figure 2. (a) Predictions of uplift rate on a 5 km by 5 km grid for the reference ice model and an Earth viscosity model that, when combined with the reference ice model, provides a good fit to the regional sea-level data base (Shennan *et al.* 2006b). This model adopts a 71 km lithosphere thickness, and a upper and lower mantle viscosity of 5×10^{20} Pa s and 1×10^{22} Pa s, respectively. The large positive uplift rates shown in the top right of the frame are associated with the deglaciation of the Fennoscandian component of the adopted ice model. (b) Same as (a) except that predicted velocity field sampled only at GPS site locations. (c) Component of total predicted signal (a) associated with non-local ice sheet loading. (d) Component of total predicted signal (a) associated with local (British–Irish) ice sheet and ocean loading. [Correction added after online publication 13 March 2009: the image for Fig. 2 was replaced with a corrected one.]

and lower mantle viscosity of 5×10^{20} Pa s and 1×10^{22} Pa s, respectively. This model provided a good fit to the regional sea-level database.

The sea-level model we adopt solves the generalized sea-level equation (Mitrovica & Milne 2003; Kendall *et al.* 2005) and includes the influence of GIA-perturbations to Earth rotation (e.g. Milne & Mitrovica 1998). The algorithm we adopt to calculate present-day crustal motion in response to a specified ice and ocean loading history is based on the spectral technique described by Mitrovica *et al.* (1994) and extended to incorporate the signal associated with GIA perturbations in Earth rotation (Mitrovica *et al.* 2001).

The predicted uplift pattern (Fig. 2a) is that of an ellipse, oriented approximately SSW–NNE, covering most of Scotland and a part of Ireland. The centre of uplift is located in central western Scotland with a magnitude of $\sim$0.6 mm yr^{-1}. The majority of England is subsiding with rates exceeding $\sim$1 mm yr^{-1} in places. While there is a broad similarity between predictions and observations (*cf.* Figs 1b and 2a), there are significant differences. Part of the difference is due to the contrasting spatial sampling of the two patterns in Figs 1b and 2a and so, to permit a more direct comparison, we show in Fig. 2b the pattern predicted when sampled at CGPS site locations only. On comparing Figs 1b and 2b, the most obvious difference is that the observed region of uplift is located further to the east and is of larger magnitude. Given the current precision and distribution of the data sites, it is not clear if this difference in geometry is necessarily a flaw in the adopted ice-earth model, which has the largest impact on this aspect of the prediction.

As the primary aim of this study is to employ the CGPS data from Great Britain to constrain parameters associated with the deformation driven by the local BIIS model, it is necessary to identify the contribution to the signal from non-local ice sheet loading in order to try and isolate the observed signal due to the BIIS and ocean loading only. The total predicted signal (Fig. 2a) was decomposed into two separate signals: that due to the non-local ice sheets only (Fig. 2c) and that due to the BIIS and ocean load only (Fig. 2d, note that the ocean loading is the sea-level change driven by all ice sheets). As expected, the signal due to the non-local ice sheets (predominantly the Fennoscandian ice sheet) is that of long wavelength subsidence of magnitude >1 mm yr^{-1} over the study area. The most straightforward way to isolate the signal associated with the BIIS and ocean load is to consider relative vertical rates. As described above, we chose to calculate motion relative to Sheerness (18 in Fig. 1a). The predicted rates relative to Sheerness associated with the non-local ice sheets (Fig. 2c) do not exceed –0.3 mm yr^{-1} and so this component of the signal is effectively removed by considering relative rather than absolute rates. [Correction added after online publication 13 March 2009: the preceding paragraph was amended in its discussion of Fig. 2.]

We performed a forward modelling analysis using the relative vertical data. The analysis considered values of upper and lower mantle viscosity in the range 0.08–5 $\times$ 10^{21} Pa s and 1–50 $\times$ 10^{21} Pa s, respectively, for assumed lithospheric thicknesses of 71, 96 or 120 km. The results for this suite of calculations are shown in Fig. 3.The solid black line in each frame indicates the χ^2 value below which all earth viscosity models fit the data to within the 95 per cent confidence level. There are a number of points to conclude from Fig. 3. First, there is a relatively mild trade-off between values of lithospheric thickness and upper mantle viscosity. As the lithosphere is thickened (going from frame a–c) the region of 'best fit' viscosity space migrates towards higher values of upper mantle viscosity. That is, an equally good fit can be achieved by raising or lowering the lithospheric thickness as long as the upper mantle viscosity is

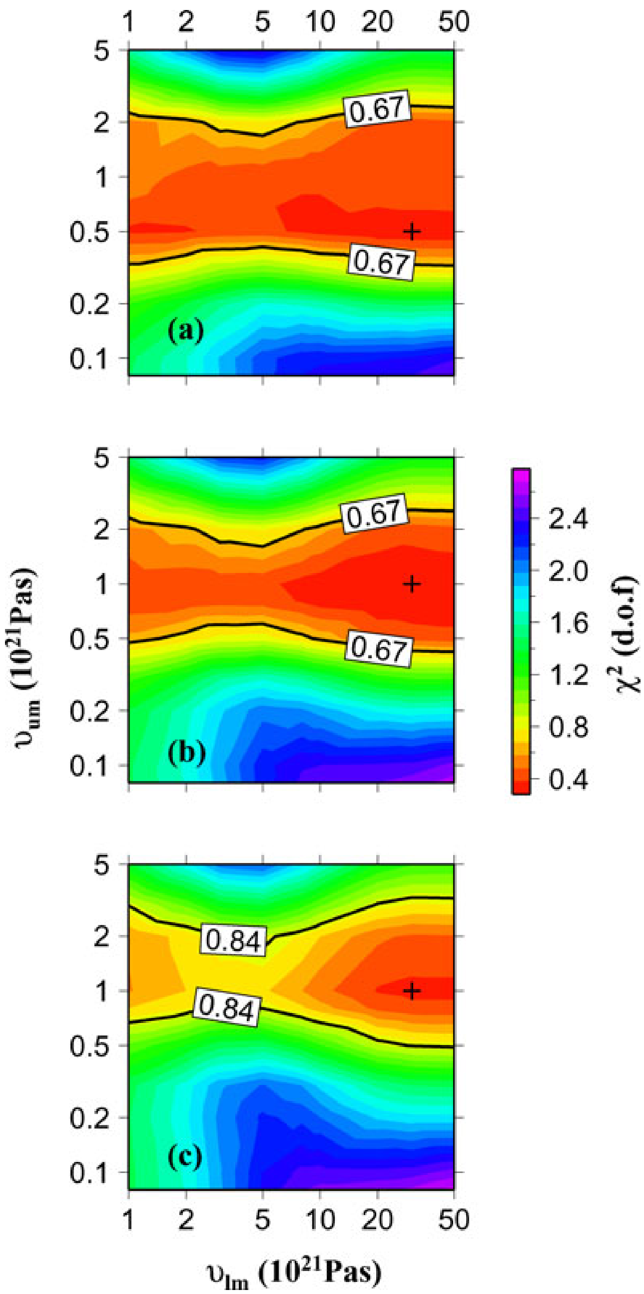

Figure 3. χ^2 values of data-model fit for a range of values for upper and lower mantle viscosity (υ_{um} and υ_{lm}, respectively) and lithospheric thicknesses of (a) 71 km, (b) 96 km and (c) 120 km. The black line and cross in each frame represent, respectively, the χ^2 value below which the quality of fit is equivalent at the 95 per cent confidence level and the minimum value found.

also raised or lowered to compensate. This parameter trade-off has been demonstrated in previous studies (e.g. Lambeck *et al.* 1996; Milne *et al.* 2004).

Second, the data rule out upper mantle viscosity values outside the (approximate) range 0.3–2 $\times$ 10^{21} Pa s (for the range of lithospheric thicknesses considered). Viscosity values less than the lower bound of this range permit the uplift to proceed too rapidly following the end of the main phase of deglaciation ($\sim$15 kyr BP), which results in predicted uplift rates that are dramatically lower than those observed. A similar constraint on upper mantle viscosity was reached using the vertical CGPS rates from Fenoscandia (e.g. Milne *et al.* 2004). Finally, the data do not provide a useful constraint on lower mantle viscosity or lithospheric thickness. The former is evident in the result that an equally good fit (to within 95 per cent confidence) can be achieved for all values of lower mantle viscosity considered. This reflects the fact that the deformation field sampled by the relative rates does not extend significantly into the lower mantle. A similar result is evident for lithospheric thickness: the quality of fit is equally good to within the specified confidence limit for the range considered.

The range of viscosity models preferred by the CGPS data from Great Britain is compatible with those inferred using relative sea-level data from this region (Lambeck *et al.* 1996; Shennan *et al.* 2006b; Brooks *et al.* 2007). We note, however, that the sea-level data display a clear preference for a relatively thin lithosphere (e.g. Lambeck *et al.* 1996) and so a joint inversion of sea-level and CGPS data will likely provide a better constraint on Earth viscosity structure.

The above viscosity inference will depend, to some extent, on the adopted local ice sheet history. The greater this dependence, the more sensitive (and therefore less robust) the viscosity inference will be to errors in the ice model. We explored this issue by generating predictions for the Earth models considered in Fig. 3 and a significantly altered ice model that is still compatible with current field constraints. Given recent discussion on the extent and thickness of ice in Ireland and the Irish Sea (Ballantyne *et al.* 2006; Hiemstra *et al.* 2006; Roberts *et al.* 2007), we considered a model that is significantly thickened and extended within the Irish Sea Basin and out to the continental shelf, with a more rapid deglaciation after 21 kyrs (Brooks & Edwards 2006). Spatial plots of ice thickness for the two ice models are shown on Fig. 4 for key time steps. The χ^2 plot for this model, assuming a lithospheric thickness of 96 km, is shown in Fig. 5. Comparison between Figs 3b and 5 shows that the results are insensitive to the changes made to the ice model and so we conclude that the viscosity inference based on the CGPS data is insensitive to plausible variations in the ice history. This insensitivity was also evident for the other values of lithospheric thickness considered. This result is most likely related

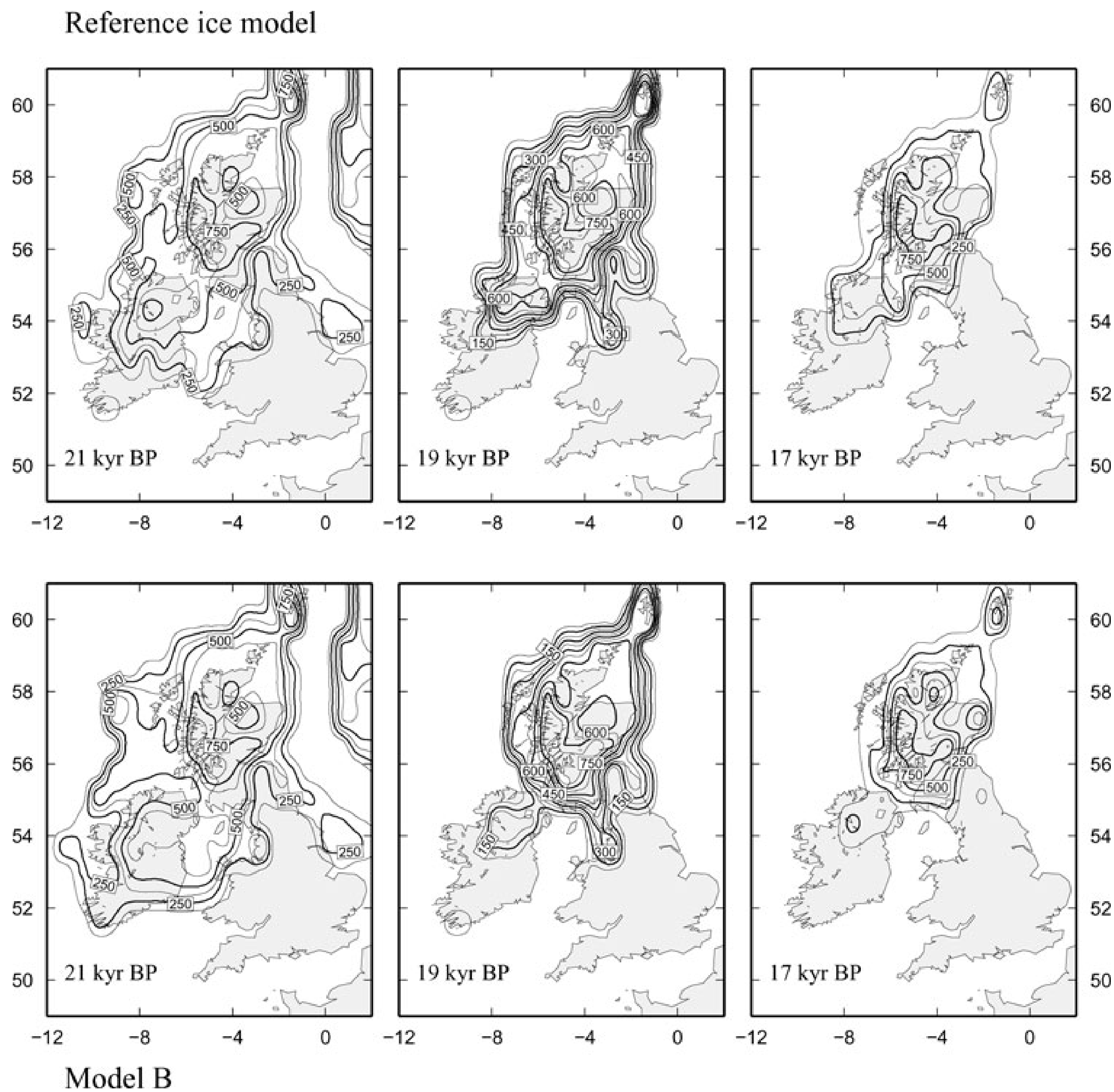

Figure 4. Spatial plots of the contoured ice thickness for the reference ice model (Shennan *et al.* 2006b) and comparison ice model (Model B) taken from Brooks *et al.* (2007). Note the varying contour interval between each time slice.

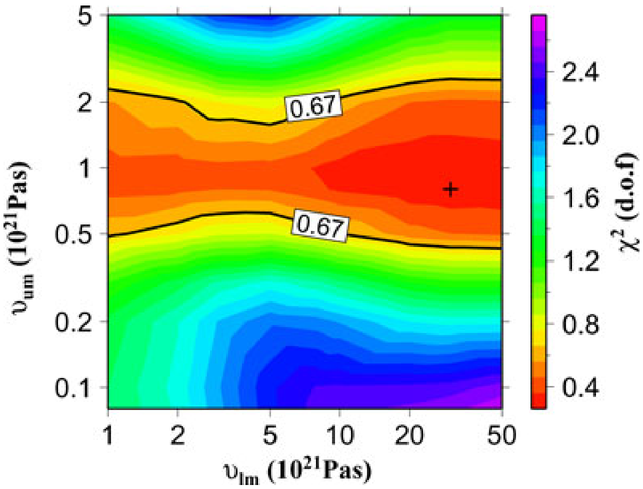

Figure 5. Same as (b) in Fig. 3 except that an alternative ice loading history was adopted (see main text for details).

to the fact that the region has been free of extensive ice cover since ∼15 cal kyr BP.

Adopting an optimum earth model from Fig. 3a (we chose a thin lithosphere to be consistent with constraints from sea-level data), we present in Fig. 6 a comparison between predicted and observed relative vertical rates at each site. Inspection of Fig. 6 shows that the predictions agree with the observations at all sites to within the estimated $1-\sigma$ uncertainty.

4 CONCLUSIONS

We considered CGPS-derived estimates of crustal velocities within Great Britain to constrain a model of GIA for the British Isles. Only the vertical component of the motion was employed to constrain model parameters since a coherent pattern of horizontal motion was not evident once a model of plate motion was removed. The observed vertical signal is strongly correlated with predictions based on a recent GIA model constructed to fit geomorphological constraints of ice extent as well as a subset of the extensive regional sea-level data base (Shennan *et al.* 2006a). A detailed forward modelling search, using this ice model and observations of relative vertical motion, was performed to locate optimum parameters in a three-layer Earth viscosity model. The results of this analysis indicate that the (relative) vertical CGPS rates are able to place a useful constraint on the average value of viscosity within the upper mantle but are not able to constrain effectively the lithospheric thickness or lower mantle viscosity. Furthermore, parameter trade-off between lithospheric thickness and upper mantle viscosity indicates that a unique constraint on upper mantle viscosity is not possible without the consideration of additional data (e.g. sea-level). A viscosity inference based on an alternative and plausible ice model demonstrate that the data are relatively insensitive to possible errors in the adopted ice-loading model. We conclude, therefore, that the CGPS data provide an important constraint on the Earth component of the GIA model and so will aid in reducing the non-uniqueness inherent to the British Isles GIA problem when combined with the sea-level and geomorphological data.

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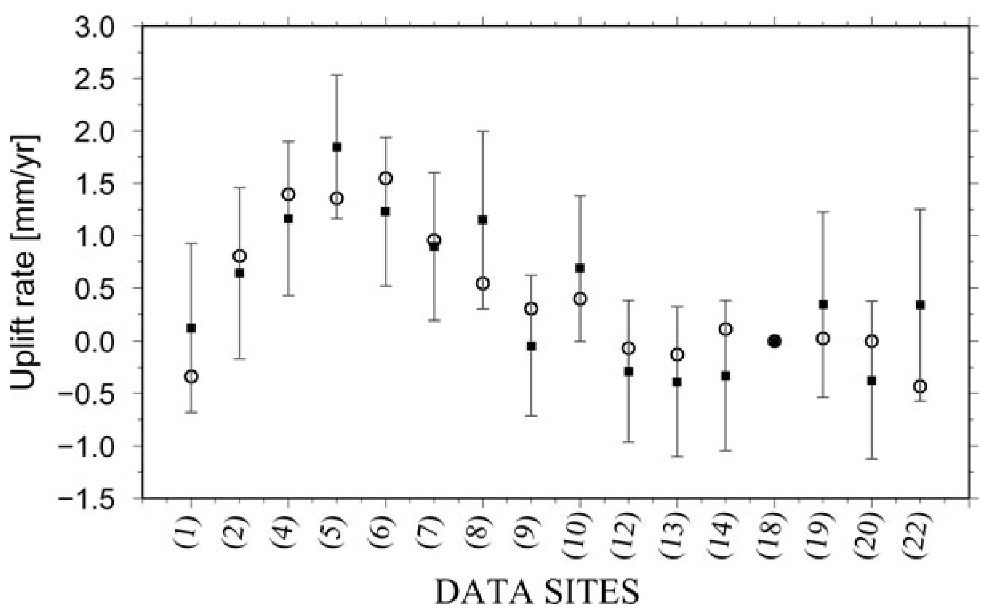

Figure 6. Observed (triangles) and predicted (circles) uplift rates relative to Sheerness (site 18) for the sites located in Fig. 1a. The predictions are based on the best-fitting viscosity model with an assumed lithospheric thickness of 71 km (Fig. 3a). Error bars indicate 1σ precision.

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APPENDIX A

Table A1 displays the estimated crustal velocities (mm yr^{-1}) at the CGPS stations shown in Fig. 1. All errors are to 1σ uncertainty

Table A1. Locations and velocities for the CGPS sites considered in this study. The time series solutions are taken from Teferle *et al.* (2008): Table 5 for the horizontal velocities and Table 9 for the AG-aligned vertical velocities

NUMBER	NAME	Longitude	Latitude	East	Error	North	Error	Vertical	Error
1	LERW	358.82	60.14	13.65	0.18	15.94	0.18	−0.65	0.54
2	THUR	356.27	58.58	14.53	0.24	15.72	0.19	−0.13	0.56
3	INVE	355.78	57.49	15.18	0.27	15.55	0.13	/	/
4	MALG	354.17	57.01	14.13	0.17	15.70	0.11	0.39	0.44
5	EDIN	356.71	55.92	15.34	0.12	15.47	0.10	1.07	0.35
6	GLAS	355.70	55.85	14.89	0.15	15.84	0.13	0.45	0.39
7	CARL	357.06	54.90	15.86	0.14	15.22	0.09	0.12	0.38
8	NEWC	358.38	54.98	15.89	0.16	15.15	0.11	0.37	0.61
9	LEED	358.34	53.80	16.35	0.14	15.27	0.08	−0.82	0.31
10	DARE	357.36	53.34	16.41	0.13	15.66	0.09	−0.09	0.36
11	IESG	358.81	52.94	16.96	0.14	14.78	0.07	/	/
12	KING	0.40	52.75	17.16	0.15	15.22	0.09	−1.07	0.32
13	LOWE	1.75	52.47	17.96	0.19	14.99	0.10	−1.17	0.40
14	PERS	357.96	52.15	15.97	0.36	15.42	0.22	−1.11	0.40
15	BARK	0.10	51.52	16.85	0.40	15.56	0.57	/	/
16	NPLD	359.66	51.42	17.03	0.27	14.65	0.22	/	/
17	SUNB	359.58	51.40	16.49	0.40	15.44	0.30	/	/
18	SHEE	0.74	51.45	17.07	0.34	14.91	0.37	−0.77	0.67
19	HERS	0.34	50.87	17.09	0.35	15.02	0.26	−0.43	0.65
20	PMTG	358.89	50.80	/	/	/	/	−1.15	0.46
21	CAMB	354.67	50.22	16.51	0.22	15.83	0.13	/	/
22	NEWL	354.46	50.10	16.45	0.21	15.71	0.17	−0.43	0.70

The following sites have been removed due to processing problems and data errors: ABER, ABYW, DUNK, HURN, LIVE, MORP, NPLD, NSTG – for the vertical velocity estimates, and ABER, ABYW. DUNK. HURN, LIVE, MORP, LOND, BLAK, COLC, NSTG, PMTG, OSHQ for horizontal velocity estimates.

Note that a model of plate motion has been adopted to remove the tectonic signal from horizontal velocities (see Fig. 1 frames c & d and related discussion).