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Glacio-isostatic deformation around the Vatnajökull ice cap, Iceland, induced by recent climate warming: GPS observations and finite element modeling

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[1] Glaciers in Iceland began retreating around 1890, and since then the Vatnajökull ice cap has lost over 400 km^3 of ice. The associated unloading of the crust induces a glacio-isostatic response. From 1996 to 2004 a GPS network was measured around the southern edge of Vatnajökull. These measurements, together with more extended time series at several other GPS sites, indicate vertical velocities around the ice cap ranging from 9 to 25 mm/yr, and horizontal velocities in the range 3 to 4 mm/yr. The vertical velocities have been modeled using the finite element method (FEM) in order to constrain the viscosity structure beneath Vatnajökull. We use an axisymmetric Earth model with an elastic plate over a uniform viscoelastic half-space. The observations are consistent with predictions based on an Earth model made up of an elastic plate with a thickness of 10–20 km and an underlying viscosity in the range $4-10 \times 10^{18}$ Pa s. Knowledge of the Earth structure allows us to predict uplift around Vatnajökull in the next decades. According to our estimates of the rheological parameters, and assuming that ice thinning will continue at a similar rate during this century (about 4 km^3/year), a minimum uplift of 2.5 meters between 2000 to 2100 is expected near the current ice cap edge. If the thinning rates were to double in response to global warming (about 8 km^3/year), then the minimum uplift between 2000 to 2100 near the current ice cap edge is expected to be 3.7 meters.

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1. Introduction

1.1. Geological Setting

[2] Vatnajökull is the largest ice cap in Iceland [*Björnsson*, 1988], covering an area of about 8100 km^2 with a mean radius of 50.7 km and a maximum thickness of about 900 m [e.g., *Björnsson et al.*, 2002] (Figure 1). Vatnajökull includes a number of outlet glaciers that have been generally retreating since 1890, although they are also affected by periodic surges [*Björnsson*, 1979; *Björnsson et al.*, 2003]. The variable ice load induces crustal deformation. To fully understand it, one needs to consider the load changes taking place, the tectonic setting, the crustal structure, and other possible sources of deformation.

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[3] Changes in air temperature have a marked impact on the mass balance of Vatnajökull as the ice cap is temperate and the main part of the ice is located at low elevation. Since the end of the Weichselian glaciation about 10,000 ^{14}C years BP, Vatnajökull has experienced considerable ice volume changes [*Björnsson*, 1979; *Sigmundsson*, 2006]. At the time of the settlement in Iceland around 900 AD, the climate was relatively warm and Vatnajökull was smaller than today. Around 1300 the climate deteriorated and lowest temperatures were experienced during the Little Ice Age from 1600 to 1900. The ice cap appears to have advanced extensively until about 1750, with smaller fluctuations occurring until 1890, after which it has been retreating [*Björnsson*, 1979].

[4] Vatnajökull partly overlies volcanic zones at the boundary between the North American and the Eurasian plates in Iceland (Figure 1). The volcanic zones in Iceland are divided into volcanic rift zones associated with extensive crustal spreading, and volcanic flank zones where crustal spreading is negligible. The volcanoes located in the volcanic rift zone at northwestern Vatnajökull in the Eastern Volcanic Zone (EVZ) are seismically very active and many eruptions have occurred there in historical times, while a volcanic flank

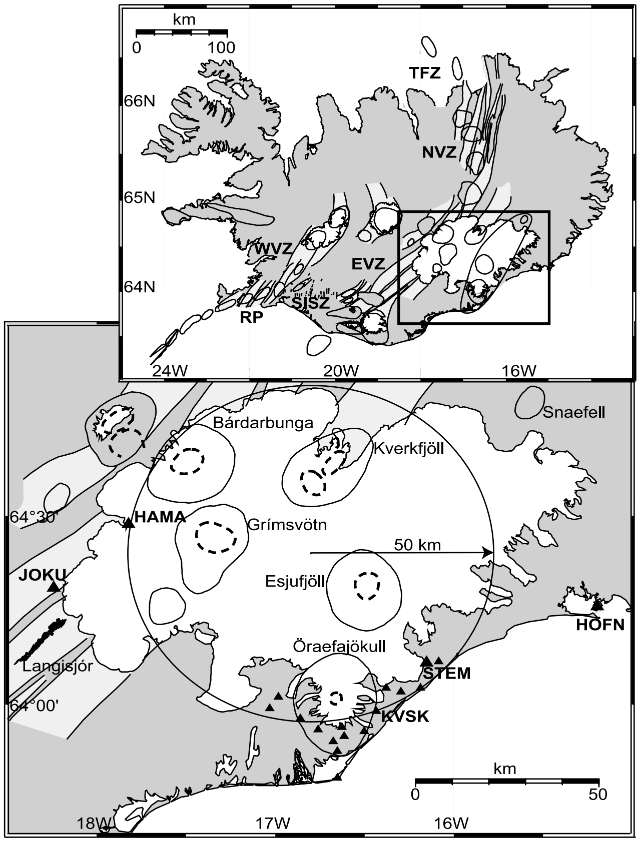

Figure 1. The Vatnajökull ice cap (white) and its tectonic setting. Overlaid are outlines of central volcanoes (solid oval outlines), calderas (dashed lines), and the outline of fissure swarms (solid lines) [after *Einarsson*, 1991]. Lake Langisjór is shown in black. The large circle gives the outline of the ice cap model. Black triangles mark the location of the GPS sites. The inset shows the tectonic setting of Iceland with the study area outlined by a box. Reykjanes Peninsula (RP), Western Volcanic Zone (WVZ), Eastern Volcanic Zone (EVZ), and Northern Volcanic Zone (NVZ), which are connected by transforms: South Iceland Seismic Zone (SISZ) and Tjörnes Fracture Zone (TFZ). Öraefajökull, Esjufjöll and Snaefell volcanoes form the Öraefajökull-Snaefell flank zone.

zone encompassing Öraefajökull, Esjufjöll, and Snaefell volcanoes is much less active (Figure 1).

[5] Seismic and gravity studies [e.g., *Allen et al.*, 2002; *Kaban et al.*, 2002] suggest that crustal thickness in Iceland varies from 15 km in coastal areas to more than 40 km in central Iceland, where the center of a mantle plume under Iceland is located. A recent study by *Ágústsson and Flóvenz* [2005] of the maximum depth of earthquake hypocenters in Iceland argued for a seismogenic thickness in the range 10–16 km in the Vatnajökull region. They suggest that the base of the elastic, brittle crust is associated with a temperature of about 750°C, below which the lower crust and upper mantle deforms in a ductile manner. Lower temperatures have been suggested at the brittle ductile transition, for example, *Menke and Levin* [1994] propose a temperature of about 400°C.

[6] A GPS network lies near the ice-covered Öraefajökull stratovolcano at southern Vatnajökull. This volcano reaches 2110-m height, and has a summit caldera about 4 km in diameter. The volcanic edifice lies uncomformably on older plateau basalts that were probably formed in a rift zone [*Prestvik et al.*, 2001], with a rock series of transitional alkaline character, including basalts, intermediate lavas, and rhyolites. Analysis of compositional variation of rocks across the Öraefajökull-Snaefell zone suggests that it is an immature rift that did not fully developed as it was replaced by the EVZ about 0.7 million years ago [*Hards et al.*, 2000]. The Öraefajökull central volcano has produced two eruptions in historical times, in 1362 and in 1727. The 1362 eruption was purely explosive, occurring after at least 500 years of quiescence and produced about 10 km^3 of rhyolitic tephra, whereas the 1727 eruption was smaller with the composition of the eruptive products more basic (andesite) [*Thorarinsson*, 1958; *Larsen et al.*, 1999]. Both eruptions caused significant melting of glacier ice culminating in jökulhlaups (glacial outburst floods).

[7] The Öraefajökull-Snaefell volcanic flank zone is currently seismically quiet. Small swarms of earthquakes occurred in 1976, 1984, and 2002 in Esjufjöll. In 1976, 60 events were recorded and in 1984 a small swarm of 10 events occurred, with the largest magnitude reaching M_t 2.3 [*Einarsson*, 1988], with M_t being the magnitude based on the time duration of the recorded earthquakes. In 1983 an earthquake M_t 2.0 originated midway between Esjufjöll and Snaefell. In October to November 2002 over 100 events with the largest having a magnitude M_w 3.2 occurred in Esjufjöll and Öraefajökull (Icelandic Meteorological Office). Earthquakes occurring in the Öraefajökull-Snaefell volcanic flank zone under the Vatnajökull ice cap and reaching only magnitude ~3 are unlikely to cause any detectable deformation outside the ice cap edge. No geophysical evidence suggests the presence of any currently active magma chambers in the Öraefajökull-Snaefell zone, nor for any magmatic processes that may contribute to present-day surface deformation. Accordingly, we present a model of deformation taking place near the southern edge of Vatnajökull attributed solely to glacio-isostasy.

1.2. Studies of Glacio-isostatic Adjustment Around Vatnajökull

[8] Deglaciation in Iceland at the end of the Weichselian glaciation, about 10,000 ^{14}C years BP, was associated with rapid glacial rebound, apparently being completed in only about 1000 years in coastal areas. This exceptionally fast postglacial rebound suggests a viscosity under Iceland on the order of 10^{19} Pa s or less [e.g., *Sigmundsson*, 1991]. Such a low viscosity results in the rapid response of the Earth to contemporary changes in ice volume. Major ice retreat is currently ongoing at ice caps in Iceland in response to warmer climate. The ice volume loss at Vatnajökull between 1890 to 2000 is estimated to be about 435 km^3 (Helgi Björnsson, personal communication, 2006) and ongoing uplift around Vatnajökull has been reported by several geodetic studies. Previous studies of glacio-isostasy in Iceland are summarized by *Sigmundsson* [2006].

[9] In this paper we infer crustal deformation from GPS surveys conducted in 1996 2002, 2003, and 2004 around Öraefajökull as well as from annual GPS measurements at

several sites around the ice cap, providing longer and more detailed time series. Previous observations include lake level measurements, an alternate set of GPS observations, and measurements of temporal variation in gravity. Lake leveling measurements at Lake Langisjór at the SW edge of Vatnajökull (Figure 1) were performed in 1959, 1991, and in 2002. These measurements show a relative uplift rate of about 4 mm/yr between benchmarks spaced 15 km perpendicular to the ice edge [*Sigmundsson and Einarsson*, 1992]. In 1991 a GPS network of 10 points was established around the southeastern edge of Vatnajökull [*Sigmundsson*, 1992; *Einarsson et al.*, 1996]. In 1992, 1996, and 1999 the original 1991 GPS network was remeasured as well as eleven additional points included in the network in 1992 [*Sjöberg et al.*, 2000, 2004]. The uplift rate 1992–1999 was estimated to be about 5–19 mm/yr, decaying radially from the center of the ice cap. Repeated gravity measurements were conducted by *Jacoby et al.* [2001] at all GPS stations annually from 1991 to 2000, except in 1994. Gravity changes for points close to the ice cap are consistent with uplift rates up to 20 ±10 mm/yr.

[10] The observed uplift around Vatnajökull has been correlated with ice volume loss in order to constrain the underlying viscosity structure. The models include an elastic plate overlaying a viscous fluid or a viscoelastic material. The elastic plate represents the uppermost part of the Earth's crust that behaves elastically over long timescales. The thickness of the elastic plate is here referred to as elastic thickness. It should approximately correspond to the elastic, brittle part of the crust as inferred by seismic studies, which is suggested by *Ágústsson and Flóvenz* [2005] to be in the range 10–16 km in the Vatnajökull region.

[11] *Sigmundsson and Einarsson* [1992] used the lake level measurements at Lake Langisjór to constrain the viscosity structure, modeling the Earth as a Newtonian viscous fluid half-space overlain by an elastic plate. They considered a circular ice load and constant ice volumes before ice melting began in 1890. They used three different ice-thinning rates as a function of the distance from the ice cap center to mimic a realistic ice retreat, with higher thinning rates at the edge of the ice cap. Their modeling results suggested an elastic thickness of about 10 km and an underlying viscosity of 1×10^{18} to 5×10^{19} Pa s.

[12] *Thoma and Wolf* [2001] used lake level measurements of Lake Langisjór from 1959 to 1991 and GPS measurements from 1992 to 1996 [*Sjöberg et al.*, 2000] to constrain the viscosity structure. A compressible, self-gravitating, spherical Earth model with Maxwell viscoelasticity and an elliptic ice load was used, together with two different load history models. In the first model they assume a constant ice volume before the beginning of ice melting in 1890, and in the second case they consider a linear increase in ice volume between 1200 to 1750, followed by a constant ice volume from 1750 to 1890, when ice retreat begins. The modeling results suggest a lower crust/upper mantle viscosity in the range of 7×10^{16} to 3×10^{18} Pa s and an elastic thickness of 10–20 km. However, the short GPS observation sessions and lack of constraints on the uplift rate at their GPS reference station in Höfn may bias their conclusions. They showed that uplift predictions are not affected by the two different ice models if the viscosity is lower than about 1×10^{18} Pa s, in which case, the Earth's response is primarily controlled by ice volume loss after 1890.

[13] *Sjöberg et al.* [2004] derived vertical velocities around the southern edge of Vatnajökull from GPS measurements in 1992, 1996, and 1999, assuming an uplift rate at their reference station in Höfn of 10 mm/yr. They concluded that the vertical GPS velocities can be fit by a model with an elastic thickness of 10–20 km and an underlying viscosity perhaps as low as 1×10^{17} Pa s. However, they do not show the model prediction for these parameters. Their interpretation is based on the models of *Wolf et al.* [1997a, 1997b], which assume ice retreat continuing only until 1980. Hence the uplift rates for 1992–1999 would most likely be affected by later ice thinning not accounted for in the models.

[14] *Jacoby et al.* [2001] measured gravity changes between 1991 and 2000 and compared them to the gravity changes predicted by the model of *Sigmundsson and Einarsson* [1992]. The gravity data was interpreted in terms of a lower crust/upper mantle viscosity on the order of 10^{18} Pa s and an elastic thickness of about 10 km.

1.3. Ice Retreat History

[15] Glacier retreat began in Iceland around 1890 due to a mean temperature increase of 1–2°C over the last century. The area of Vatnajökull as estimated from a Landsat satellite image in 1973 was 8300 km^2 [*Björnsson*, 1978, 1979]. The area of Vatnajökull in 1890 was on the other hand estimated to have been 8600 km^2, using maps of Iceland by the Danish General Staff from 1904–1973 and extrapolated ice retreat rates (Helgi Björnsson, personal communication, 2006). The total ice volume loss from 1890 to 1973 was initially estimated to be 182 km^3 [*Sigmundsson and Einarsson*, 1992]. Regular measurements of the surface mass balance of Vatnajökull started in 1991 [*Björnsson et al.*, 1998]. Because of climatic variability, the ice cap's mass balance was positive from 1991 to 1994, close to zero in 1995 but negative from 1996 to 2001, giving a total ice loss between 1991 and 2001 of about 21 km^3 [*Björnsson et al.*, 2002]. By compiling data from various sources, a revised estimate of the ice volume loss between 1890 and 2003 results in a value of 435 km^3 (Helgi Björnsson, personal communication, 2006).

2. Deformation Data

[16] The main data set used in this study consists of vertical and horizontal velocities derived from GPS measurements in 1996, 2002, 2003, and 2004 (Figures 1, 2, and 3). A GPS network of 15 stations was measured in 1996 around the Öraefajökull volcano at the southern edge of the Vatnajökull ice cap. In 2002, four GPS points were remeasured. In 2003, all the stations from the 1996 campaign were remeasured. In 2004 three GPS points were remeasured. The GPS station Kvisker (KVSK) was used as a reference station for all campaigns, collecting data throughout the surveys. Other GPS measurements around Vatnajökull used in this study are at the sites HAMA, JOKU, and STEM (Figure 1). HAMA and JOKU, at the western edge of Vatnajökull have been measured annually from 1997 to 2004, except in 1998. The GPS station STEM at the southern edge of the Vatnajökull ice cap has also been measured yearly from 2002 to 2005. The

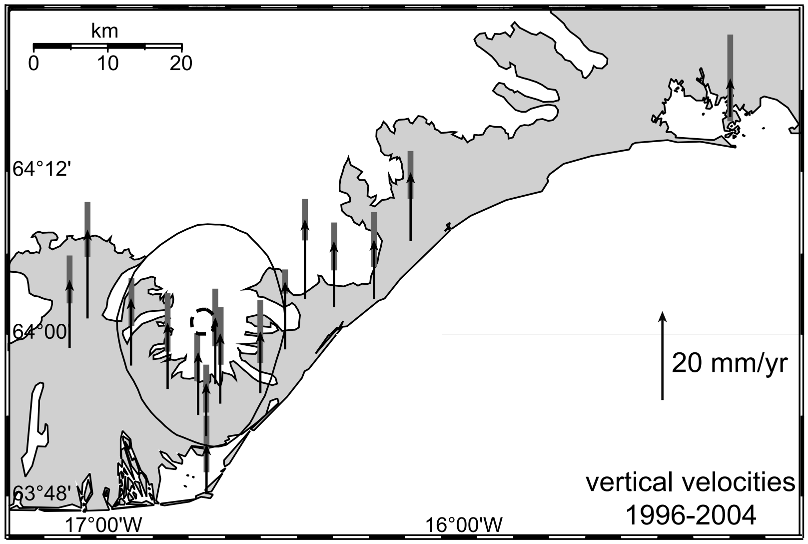

Figure 2. GPS vertical velocities 1996–2004 (mm/yr) relative to ITRF00 stable Eurasia reference frame. Velocities vary from 9 to 20 mm/yr with uncertainties between 2–4 mm/yr. Gray bar shows inferred 95% confidence interval. Overlaid is the outline of the Öraefajökull central volcano (solid line) and its caldera (dashed line).

continuous GPS station HOFN in eastern Iceland was installed in 1997 and it has been continuously operating since then.

[17] Our model predictions are also compared to lake leveling measurements at Lake Langisjór at the SW edge of Vatnajökull.

2.1. GPS Measurements

[18] The GPS data from the 1996, 2002, 2003, and 2004 campaigns around Öraefajökull have been processed with the GAMIT/GLOBK GPS software, version 10.2 [*Herring et al.*, 2006]. We used the GAMIT software to estimate the positions of our campaign GPS sites and all continuous GPS sites in Iceland (including REYK and HOFN) together with about 20 nearest stations of the IGS global network, located on the Eurasian and the North-American plates. We then used the GLOBK software to impose the International Terrestrial Reference Frame (ITRF) 2000 relative to stable Eurasia to our results. The GPS data set had previously been processed with the Bernese GPS software, version 4.2, and the resulting velocities [*Pagli*, 2006] are comparable to results obtained with GAMIT/GLOBK.

[19] The GPS data from the sites HAMA, JOKU, and STEM were processed with the Bernese software, version 4.2. The velocity vectors were initially computed relative to Reykjavík (REYK), and then we corrected for the subsidence of REYK at a rate of 3.7 ± 0.4 mm/yr relative to ITRF00 [*Árnadóttir et al.*, 2006].

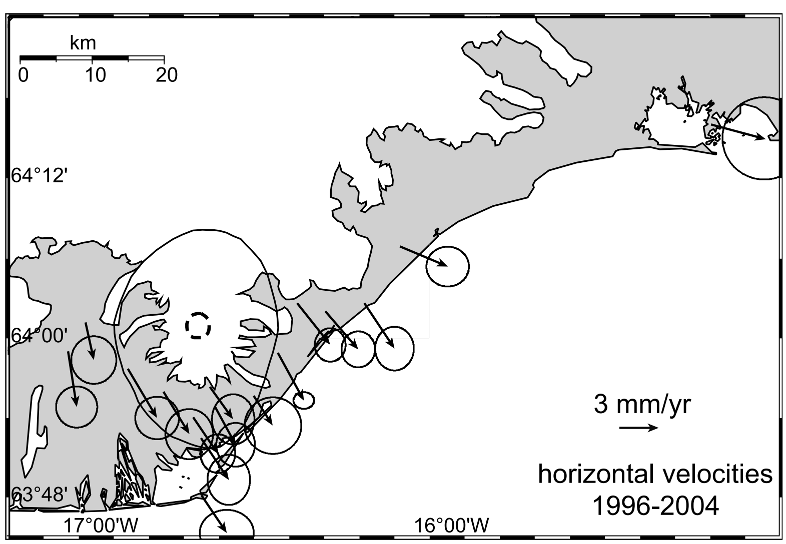

Figure 3. GPS horizontal velocities 1996–2004 (mm/yr) relative to ITRF00 stable Eurasia reference frame. Velocities vary between 3 and 4 mm/yr with uncertainties between 1 and 2 mm/yr. Error ellipses show inferred 95% confidence interval.

[20] Our measurements complement the work of *Sjöberg et al.* [2000, 2004]. In their 1992 survey, measurements were collected at 30 s intervals and the observation time was about 9 hours per session, with each station being occupied about three times. In their 1996 survey, data were sampled at 15 s intervals with 9 hours observation time per session. The time span of their observations was one 12 hours session per day in 1999 and data were sampled at 15 s intervals. In our 1996–2004 GPS campaigns, the time span of the observations at each site was at least one complete 24-hour session, starting at 00:00 GMT (Greenwich Mean Time) and ending at 23:59 GMT, with data collected every 15 s. The long observation time in our 1996–2004 campaigns enables us to reduce the uncertainty in the derived GPS displacement rates compared to earlier studies. Furthermore, the data have now been analyzed with the GAMIT/GLOBK software, allowing simple evaluation of the deformation relative to the stable Eurasian plate.

2.2. Velocity Field

[21] The vertical velocities for 1996 to 2004 around Öraefajökull vary from 20 mm/yr at the stations next to the ice cap, decaying to 9 mm/yr at our farthest station (HOFN), with inferred uncertainties between 2–4 mm/yr (Figure 2). Horizontal velocities do not exceed 4 mm/yr with inferred uncertainties between 1 and 2 mm/yr (Figure 3). Vertical velocities are exceptionally high compared to the horizontal velocities, as characteristic for areas exhibiting glacio-isostatic adjustment [e.g., *Pinel et al.*, 2007]. Magmatic movements associated with the Öraefajökull volcano are an unlikely source of the deformation signal we observe, as the resulting deformation field would be different, with a more rapid decay in uplift rates away from the volcano. A pressure increase at depth (i.e., a Mogi source) would cause both high vertical and horizontal velocities, which are not observed at Öraefajökull. All GPS vertical velocities measured in eastern, southern, and western Vatnajökull (Figures 2 and 4) are higher close to the ice cap edge and decay radially away from it, consistent with glacio-isostatic adjustment around Vantajökull. While, if the deformation signal was due to magmatic activity at Öraefajökull, the vertical velocities would be higher around the Öraefajökull volcano than elsewhere, which is not the case. Thus we infer that the deformation we observe is due to glacio-isostatic adjustment caused by thinning of the Vatnajökull ice cap.

[22] The vertical velocities of HAMA and JOKU are 19 ± 5 and 25 ± 5 mm/yr relative to ITRF00, respectively (Figure 4). The vertical velocity of STEM is 24 ± 6 mm/yr relative to ITRF00 in the time span 2002–2005 (Figure 4).

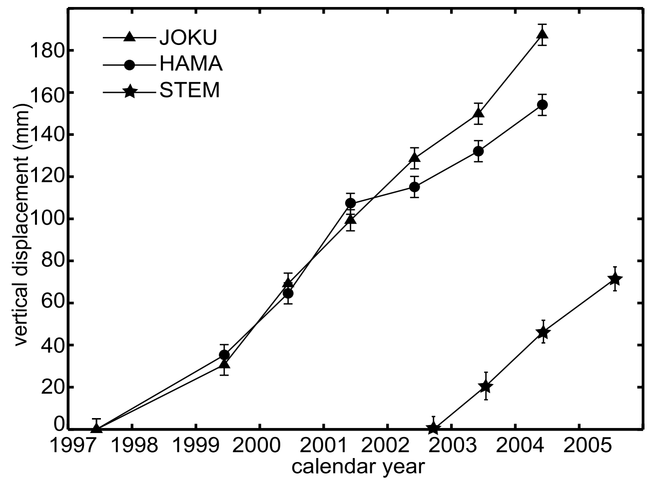

Figure 4. GPS time series around Vatnajökull relative to ITRF00. GPS vertical displacements at JOKU and HAMA 1997–2004, and STEM 2002–2005 (Figure 1). Average vertical velocity is 25 ± 5 mm/yr at JOKU, 19 ± 5 mm/yr at HAMA and 24 ± 6 mm/yr at STEM.

3. Numerical Model

[23] We have constructed an axisymmetric Finite Element model, using the ANSYS 7.1 software [*ANSYS Incorporation*, 2005], to constrain the viscosity structure in the Vatnajökull region. The dynamic response of the Earth to the gradual thinning of an axisymmetric ice load with a rectangular profile is evaluated. In our main model we describe Vatnajökull using a circular ice cap, centered at latitude 64.4°N and longitude 16.8°W, with a 50-km radius, similar to the work of *Sigmundsson and Einarsson* [1992]. Such a model facilitates the calculations, although it is a relatively simple approximation of the real ice cap. The areal extent of the ice load at the eastern and western parts of the ice cap is underestimated, but it is overestimated in the northern and southern parts of Vatnajökull. Consequences of using a circular ice cap are evaluated in section 7. We assume ice thinning since 1890, with isostatic equilibrium conditions prior to that. Models incorporating earlier ice volume fluctuations, as well as a different ice load geometry, were also tested. We model the Earth as an isotropic, incompressible, Maxwell viscoelastic half-space with a constant gravitational acceleration g = 9.81 m/s^2, overlaid by an elastic plate. The viscoelastic substrate corresponds to the lower crust and upper mantle and the elastic plate to the elastic, brittle crust.

3.1. Mathematical Model for Viscoelastic Deformation

[24] If the long-term rheological response of the Earth is approximated by a viscous fluid, then its equilibrium surface deflection due to an applied load is balanced by buoyancy forces, as stated by the Archimedes' principle. A more realistic model of the Earth is a Maxwell viscoelastic solid (a mechanical analog is a spring and a dashpot connected in a series), responding elastically over short timescales but viscously over long timescales. The equation of motion for the viscoelastic problem is commonly transformed to the corresponding elastic problem according to the correspondence principle of linear elasticity and using the Laplace Transform [*Cathles*, 1975, equation II-59; *Flugge*, 1975]. Assuming that the Earth is flat, incompressible and nonself-gravitating, the equation of motion for a continuous elastic material can be written as [*Wu*, 2004]:

$$\vec{\nabla}\boldsymbol{\sigma} - \rho_0 g_0 \vec{\nabla} w = 0 \tag{1}$$

where $\boldsymbol{\sigma}$ is the stress tensor, ρ_0 and g_0 are the reference density and gravitational acceleration, respectively, and $w = \vec{u} \cdot \hat{z}$ is the vertical component of the displacement vector ($\vec{u}$ displacement vector, $\hat{z}$ unit vector in the vertical direction, positive upwards). The first term is the divergence of stress and the second term is the advection of pre-stress that corresponds to the restoring force of isostasy [*Wu*, 2004]. By neglecting the second term, there will be no viscoelastic gravitational relaxation.

[25] Finite element (FE) packages are usually not designed for geophysical applications and the force of isostasy is not considered, in that the equation of motion is written as $\vec{\nabla}\boldsymbol{\sigma} = 0$. Therefore some modifications need to be applied to the FE packages to study the deformation in a viscoelastic Earth. We follow the same approach as described by *Wu* [2004]. We define a new stress tensor

$$\mathbf{s} = \boldsymbol{\sigma} - \rho_0 g_0 \vec{w}\mathbf{I} \tag{2}$$

where $\vec{w} = (0, 0, w)$ and $\mathbf{I}$ is the identity matrix. Equation (1) becomes $\vec{\nabla}\mathbf{s} = 0$. This is the same equation as used in FE packages, except that the new stress tensor is used. Because of the modifications to equation (1), the boundary conditions need to be modified accordingly. The term $\rho_0 g_0 w$ can be accounted for in finite element models by placing an elastic foundation (also called the Winkler foundation) on the free surface and at each density contrast across the material interfaces. The foundation acts as an elastic spring with a spring constant k, equal to the density contrast across the interface times the gravitational acceleration. The elastic foundation has an important physical meaning. As an example, at the free surface, if we assume $k = \rho_0 g_0$, the foundation will exert a pressure k on the surface of the model for each unit of displacement in the vertical component, w. Therefore the contribution of the elastic foundation at the free surface will be $\rho_0 g_0 w$, which is the buoyancy force.

[26] The constitutive equation of an isotropic viscoelastic material can be written in terms of the relaxation function of the deviatoric stress tensor, assuming an incompressible Earth. This tensor is a function of the shear modulus, which for a Maxwell material may be represented by a finite series of exponentials, called a Prony series [*Belytschko et al.*, 2000]. The Prony representation corresponds to the solution of the classical Maxwell model of viscoelasticity. Using one Prony series, the shear modulus can be written as:

$$G = G_0 \exp\left(-\frac{t}{\lambda}\right) \tag{3}$$

where G_0 is the shear modulus at the fast load limit, t is the current time and $\lambda = \eta / G_0$ is the Maxwell relaxation time for the Prony component G with viscosity η. The constitutive equation of an isotropic viscoelastic material can be written as:

$$\boldsymbol{\sigma} = \int_0^t 2G_0 \exp\left(-\frac{(t-\tau)}{\lambda}\right) \frac{d\boldsymbol{\varepsilon}}{d\tau} d\tau \tag{4}$$

where $\boldsymbol{\sigma}$ is the stress tensor, $\boldsymbol{\varepsilon}$ is the deviatoric strain tensor and τ is past time [*Belytschko et al.*, 2000; *ANSYS Incorporation*, 2005]. This convolution integral can recover the fully elastic behavior at the fast load limit ($t = 0$) and the viscous flow at the slow limit ($t = \infty$). The equation is a solution of the constitutive equation of a Maxwell material, commonly written as:

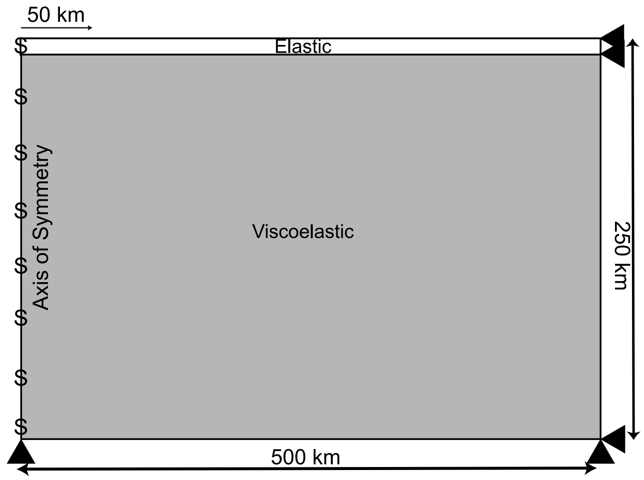

Figure 5. Model geometry and boundary conditions. The black triangles indicate that zero displacements both in the x and y directions are imposed at the far field boundaries.

$$\dot{\boldsymbol{\sigma}} + \frac{\boldsymbol{\sigma}}{\lambda} = 2G_0 \dot{\boldsymbol{\varepsilon}} \tag{5}$$

where $\dot{\boldsymbol{\sigma}}$ and $\dot{\boldsymbol{\varepsilon}}$ are the time derivatives of the stress and strain tensors, respectively.

[27] The validity of this method and our application of it has been tested by conducting the same FE tests as done by *Wu* [1992]. It was shown that the surface vertical displacements for a two-dimensional model of a viscoelastic half-space, undergoing an impulsive boxcar load, is identical to results obtained numerically using a Fast Fourier Transform (FFT). Our FE model test runs are in agreement with the results of *Wu* [1992]. Additional FE studies of glacio-isostatic adjustment using this method include *Wu* [1993], *Kaufmann et al.*, [2000], and *Lund* [2005].

3.2. Earth Model Parameters

[28] We approximate the Earth as an infinite half-space. Radial symmetry of the Earth and ice model is assumed, with the axis of symmetry at the center of Vatnajökull and the model boundaries far from the center of the ice cap (Figure 5). The size of the model needs to be sufficiently large such that the far-field boundaries, where displacements are not allowed, do not affect the results. The model geometry presented in this paper is therefore the result of a number of tests to determine how large modeling space is required. *Lund* [2005] suggested using a model geometry with a factor of at least 10 between the radius of the applied load and the width of the model, and a factor of five between the radius of the applied load and the thickness of the model. Our Earth model is therefore 500 km wide and 250 km thick (Figure 5). For the thickness of the elastic layer we used 10 and 20 km. The thickness of the underlying viscoelastic layer was thus 240 and 230 km, respectively.

[29] Both the elastic and viscoelastic layers are meshed with eight-node elements with quadratic displacement behavior on the element sides, called PLANE183 in the

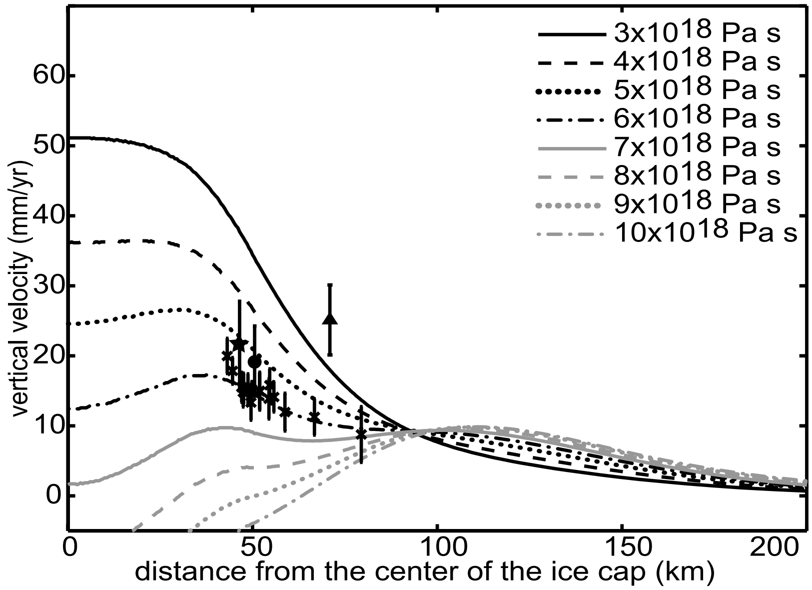

Figure 6. Modeled vertical velocities (lines) and observed GPS vertical velocities (symbols) plotted as a function of distance from the center of the ice cap, using different viscosities. Black crosses are the vertical velocities of the GPS sites as in Figure 2 and the other symbols are the vertical velocities of the JOKU-HAMA-STEM sites as in Figure 4. We use a 50-km radius ice cap model, a 10-km elastic thickness and we assume a sudden emplacement of the ice cap in the year 900 (5800 km^3), a uniform ice accumulation phase from 900 to 1750, corresponding to an increase in ice volume of 1265 km^3, followed by stable ice conditions until 1890 when ice thinning begins. Only the GPS site JOKU (black triangle) appears displaced with respect to the other GPS points. This effect is caused by assuming a circular ice cap, which underestimates the extent of the ice in western Vatnajökull. The JOKU site is located at the edge of the real ice cap but it is 20 km away from the edge of the model with a radius of 50 km (Figure 1).

ANSYS 7.1 software. Each node has two degrees of freedom, namely translations in the x and y directions. The elements have both elastic and viscoelastic capabilities. The elastic layer has a Poisson's ratio ν = 0.5 and Young's modulus E = 90 GPa (corresponding to a shear modulus μ = 30 GPa). The density of the elastic layer is ρ_0 = 2800 kg/m^3. The viscoelastic layer is modeled by a Maxwell material with one Prony component [equations (2) and (3) apply]. Poisson's ratio ν = 0.5, Young's modulus E = 90 GPa and several viscosities η are tested in the modeling. The density of the viscoelastic layer ρ_1 = 3200kg/m^3. The Winkler foundations have been included using the element SURF153, which is placed at the free surface and also at the interface between elastic and viscoelastic layers. At the free surface, the elastic foundation constant, $k = \rho_0 g_0$, is 27,440 Pa/m. At the interface between the elastic and viscoelastic layers, the elastic foundation constant, $k = g_0(\rho_1 - \rho_2)$, is 3920 Pa/m.

[30] The model space was meshed with quadrilateral elements of different sizes. The mesh was finer at the free surface and at the edges of the ice cap where more detailed results are needed. In the horizontal direction, the element size in the region 0 to 50 km from the ice cap's center is 2 km, in the region 50 to 150 km the element size increases linearly from 2 to 4 km, in the region 150 to 250 km it is 5 km, and in the region 250 to 500 km it is 7 km. In the vertical direction, the element size between 0 to 10-km depth is 1.25 km and between 10- to 250-km depth the size increases linearly from 1.25 to 8 km. The axis of symmetry is located in the center of the ice cap (Figure 5). Zero displacements both in the x and y directions are imposed at the model space's edge (Figure 5).

4. Ice Load Models

[31] In all our calculations the total volume of ice lost between 1890 to 2003 at Vatnajökull is 435 km^3. The main ice model assumes isostatic equilibrium in 1890 and ice loss thereafter. Mass balance studies of Vatnajökull indicate that most of the ice thinning occurs at the edge of the ice cap [*Björnsson et al.*, 1998]. Hence we define the thinning at the edge of Vatnajökull between 30 to 50 km distance from the center of the ice cap to be 62 cm/yr, with loss over the rest of the ice cap set to 25 cm/yr. We also tested ice models with ice volume changes prior to 1890 in order to understand how such changes would affect current deformation rates. We calculated current deformation rates, using an ice model that includes a sudden emplacement of the ice cap in the year 900 (5800 km^3), a uniform ice accumulation phase from 900 to 1750, corresponding to an increase in ice volume of 1265 km^3 [*Sigmundsson and Einarsson*, 1992], followed by stable ice conditions until 1890 when ice thinning begins. Results show that a viscosity of 4 $\times$ 10^{18} Pa s, and an ice cap growth phase as explained earlier, predicts current velocities significantly larger than the observed one (Figure 6). Such a growth phase is indicated by historical records in Iceland, but its details are uncertain. In any case, deviation from isostatic equilibrium in the year 1890 due to a previous ice accumulation phase would bias

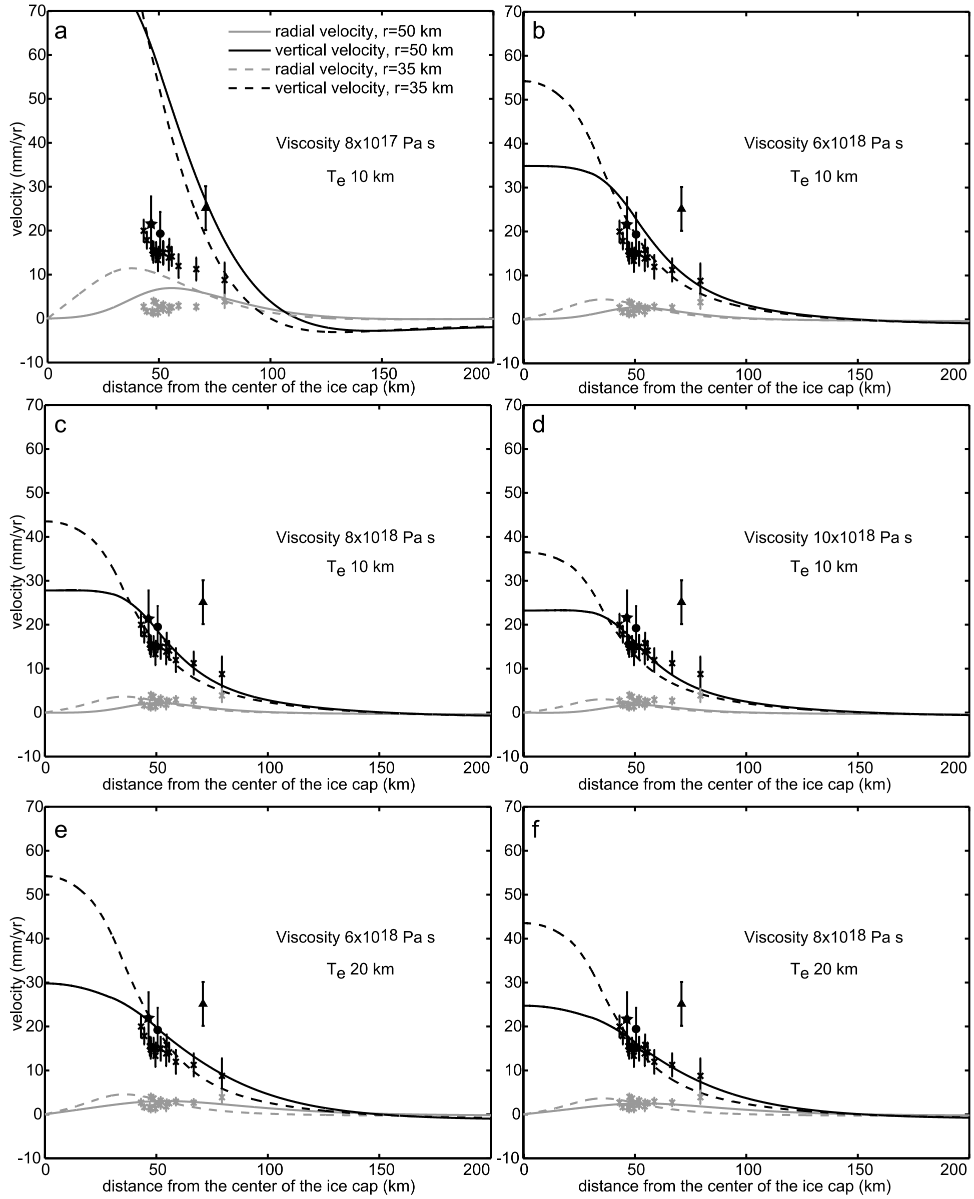

Figure 7. Modeled velocities (lines) and observed GPS velocities (symbols) plotted as a function of distance from the center of the ice cap. Black crosses are the vertical velocities and grey crosses are the radial velocities of the GPS sites as in Figure 2 and the other symbols are the vertical velocities of the JOKU-HAMA-STEM sites as in Figure 4. Modeled velocities (mm/yr) using an ice cap radius (r) of 50 km are shown with solid lines and of 35 km with dashed lines. Figures 7a–7d show the results assuming an elastic thickness T_e of the 10 km and using viscosities of 8×10^{17} Pa s, 6×10^{18} Pa s, 8×10^{18} Pa s and 1×10^{19} Pa s, respectively. Figures 7e and 7f show the results assuming an elastic thickness T_e of 20 km and using viscosities of 6×10^{18} and 8×10^{18} Pa s, respectively.

our viscosity estimates derived in the next chapter in such a way that they represent an upper limit of the real viscosity. On the basis of the modeling presented in Figure 6, and allowing for even larger amount of ice accumulation in earlier times, we infer 4×10^{18} Pa s as a lower limit for the viscosity. In the following, we consider then only the models that assume isostatic equilibrium in 1890, deriving an upper limit on the viscosity.

[32] The geometry of the ice model has also to be considered. Some GPS stations that are located close to the edge of the ice cap are underneath the circular ice model, affecting the predicted velocities. In order to assess the effect of the ice load geometry, an ice cap model with a radius of 35 km has been tested for comparison. In this model, the thinning is focused at the edges between 21- to 35-km distance from the center of the ice cap at a rate of 130 cm/yr, while in the central part ice thinning is 52 cm/yr. The modeled current velocities at the distance range of the GPS sites are similar although slightly lower with respect to the results when a 50-km ice cap radius was used. We conclude that our assumptions of the ice cap load history and the model geometry both influence to some extent the viscosity estimate. If a smaller ice cap or ice

Table 1. Modeling Results (see Section 5 for Discussion)

Elastic Thickness (km)	Viscosity (Pa s)	Ice Cap Radius 50 km		Ice Cap Radius 35 km	
		RMS (mm)	χ^2	RMS (mm)	χ^2
10	8×10^{17}	29	6340	27	6596
10	6×10^{18}	5	239	3	133
10	8×10^{18}	3	78	2	42
10	1×10^{19}	2	52	3	61
20	6×10^{18}	3	93	3	133
20	8×10^{18}	2	26	2	42

growth before 1750 is considered, then a lower value for the viscosity will be inferred. However, for the rest of the modeling, we will assume equilibrium conditions before ice thinning begins in 1890. This will provide us with an upper limit for the viscosity. We will also use both the 50- and 35-km radii ice models to fit the data.

5. Results of Modeling Current Deformation

[33] We compare the 1996 to 2004 vertical and radial GPS velocities to predictions of glacio-isostatic deformation around Vatnajökull for different values of viscosity and elastic thickness. The radial velocities are the components of the horizontal velocities along the direction connecting the center of the ice cap and each GPS site. The data were compared using models with an elastic thickness of 10 and 20 km, and viscosities, η, of 8×10^{17}, 6×10^{18}, 8×10^{18}, and 1×10^{19} Pa s. We plot the predictions of current vertical and radial velocities as a function of distance from the center of the ice cap. For each viscosity value, we plot the velocity predictions arising from both the 50- and a 35-km ice model radius (Figure 7). In order to evaluate the fit of the model, we use both the GPS vertical and radial velocities. The fit of the vertical velocities is expressed by the RMS and the $\chi^2 = (r^{\mathrm{T}}\Sigma^{-1}\mathrm{r})$, where r is the residual between the observed velocities and the model prediction and Σ is the data covariance matrix. Modeling results, using an elastic thickness of 10 km, show that a viscosity of 8×10^{18} Pa s gives the best fit for the 35-km ice model radius, and 1×10^{19} Pa s for the 50-km ice model radius (Figures 7a–7d and Table 1). Similar results are obtained using an elastic thickness of 20 km (Figures 7e and 7f). Lower RMS values are achieved with a viscosity of 8×10^{18} Pa s than for a viscosity of 6×10^{18} Pa s (Table 1). The velocity of the GPS site JOKU appears displaced with respect to the other GPS points (black triangle in Figures 6 and 7). This effect is caused by assuming a circular ice cap, which underestimates the extent of the ice in western Vatnajökull. The JOKU site is located at the edge of the real ice cap but it is 20 km away from the edge of the model with a radius of 50 km (Figure 1). As a consequence, we do not use this GPS site to constrain the viscosity.

6. Future Deformation

[34] On the basis of our current knowledge of the viscosity structure beneath Iceland, we can attempt to predict the future uplift at Vatnajökull caused by ice reduction in the coming decades due to global warming. We use the ice model with a 50-km radius, and assume that the past average thinning rate will continue throughout this century (a thinning rate of 25 cm/yr applied to the central part of the ice cap and 62 cm/yr between 30 to 50 km). The elastic thickness was set to 10 km and our preferred viscosity of 8×10^{18} Pa s is used. The resulting prediction is a vertical velocity of about 30 mm/yr in 2001 (Figure 8a) and at least 2.4 m of uplift between 2000 and 2100 at the ice cap edge (Figure 8c). However, the future evolution of the Vatnajökull ice cap is uncertain with predicted climate warming this century possibly leading to a large increase in the rate of ice retreat [*Marshall et al.*, 2005; *Flowers et al.*, 2005]. Therefore we also estimate the uplift assuming that the thinning rates are twice the current value between 2000 and 2100. This model predicts a vertical velocity of at about 45 mm/yr in 2100 (Figure 8b) and 3.7 m of predicted uplift between 2000 and 2100 at the ice cap edge (Figure 8d).

7. Discussion

[35] The ice load model together with the parameters of the Earth model determine the predicted deformation field. Uncertainties in both temporal evolution and spatial distribution of the ice load model need to be considered. Results from the finite element modeling indicate that a viscosity of 8 to 10×10^{18} Pa s provides the best fit to the GPS velocities in our study area if isostatic equilibrium in the year 1890 and an elastic thickness between 10 and 20 km are assumed. The uncertainties in both the temporal and spatial variation in the ice load influence our preferred viscosity, such that this value is an upper limit, assuming that significant ice retreat started in 1890 as suggested by the ice retreat history [*Björnsson*, 1979]. Modeling results indicate that for viscosities of 10×10^{18} Pa s, changes in the ice retreat history, such as ice volume increase prior to 1750, affect our predictions of the current velocities. Including this effect would thus lead to a lower viscosity estimate. Tests using various models indicate that a viscosity of 4×10^{18} Pa s is the lower limit. For lower viscosity values all models give too high uplift rates, even if the likely ice accumulation phase from 900 to 1750 is considered.

[36] Lake leveling measurements from Lake Langisjór are in agreement with a viscosity of 8×10^{18} Pa s, when the effect of the ice load extent is included. Our circular ice cap model underestimates the extent of the ice next to Lake Langisjór. We correct this effect by comparing the Lake Langisjór measurements of predicted uplift as a function of distance from the ice cap edge, rather than from the center of the ice cap.

[37] Glacio-isostatic deformation around Vatnajökull has been modeled by *Thoma and Wolf* [2001] and *Sjöberg et al.* [2004] who argue for viscosities below Iceland as low as 10^{17} Pa s. Our model assumes a uniform viscoelastic material under an elastic plate. For this model, the inferred upper limit on viscosity is 10×10^{18} Pa s. Considering uncertainties in the temporal and spatial distribution of the ice load a lower limit on the viscosity is 4×10^{18} Pa s. Therefore viscosity values as low as 10^{17} Pa s are not consistent with any of our modeling results. Moreover, such very low viscosities would allow rapid adjustments to ongoing load changes, with current vertical velocities expected to show increasing rates, which are not observed at the GPS sites measured annually (Figure 4). Thus viscosities below Iceland as low as 10^{17} Pa s

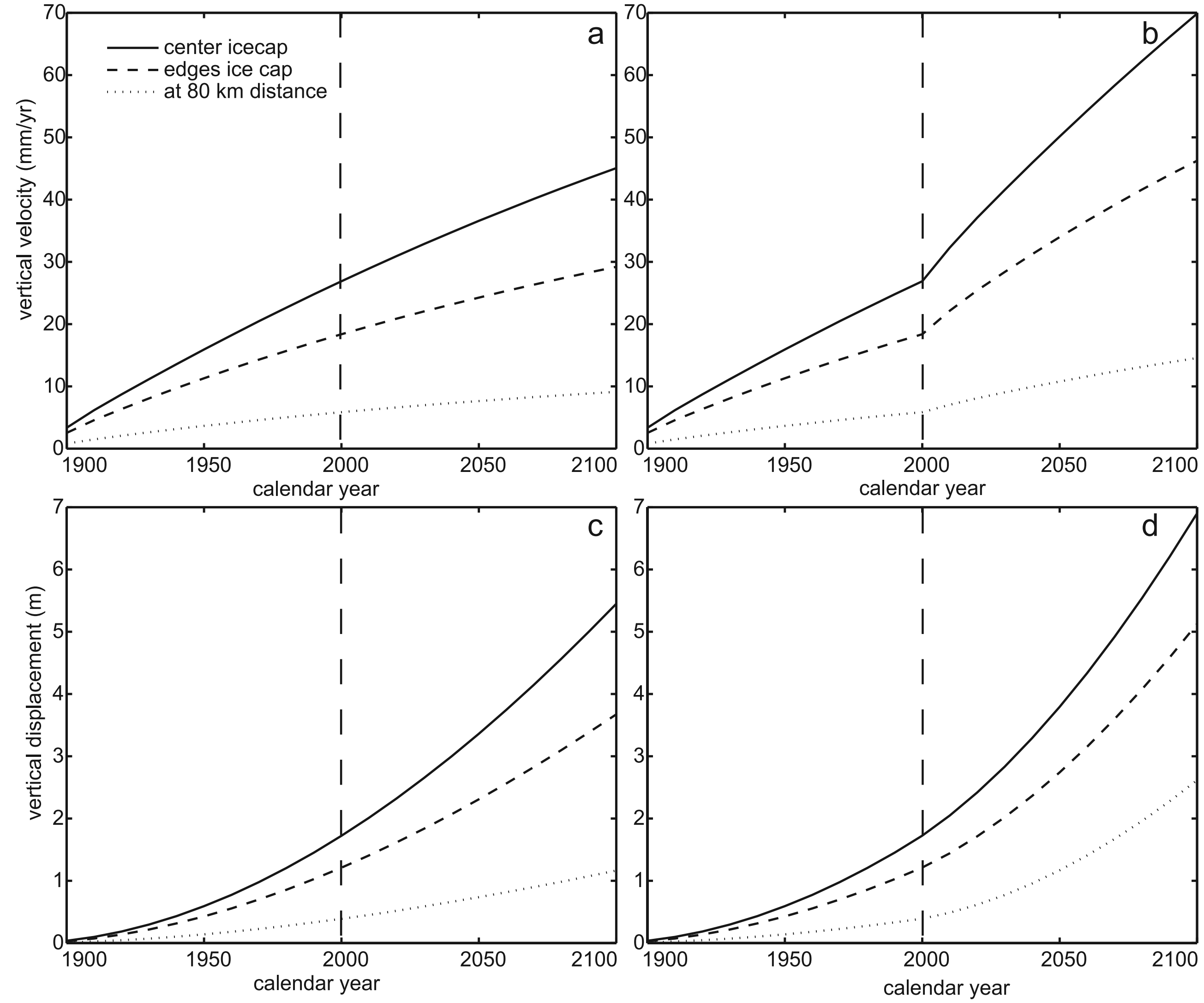

Figure 8. Predicted vertical velocities and displacements at Vatnajökull, assuming an elastic thickness of 10 km and an underlying viscosity of 8×10^{18} Pa s. Lines show the (a) predicted vertical velocities and (c) displacements assuming a thinning rate of 4 km^3 until 2100, Figures 8b and 8d show the response if thinning rates were to double in 2000 and remain constant until 2100.

are unlikely. These earlier values were based on a previous estimate of ice volume change, where *Thoma and Wolf* [2001] and *Sjöberg et al.* [2004] assumed an average thinning rate after 1890 of about 2 km^3/yr. As we discussed earlier, a recent revision of the ice volume loss for Vatnajökull from 1890 to 2003 gives an average thinning rate of about 4 km^3/yr, leading to higher viscosity estimate.

[38] Our preferred viscosity estimate is in agreement with results obtained by different studies of the rheological structure beneath Iceland [e.g., *Pollitz and Sacks*, 1996; *Árnadóttir et al.*, 2005]. *Pollitz and Sacks* [1996] modeled GPS measurements in northern Iceland in terms of post-rifting relaxation, following the 1975–1984 Krafla rifting episode. They favor a lower crustal viscosity of 3×10^{19} Pa s and an upper mantle viscosity of 3×10^{18} Pa s. Similar results are obtained by *Árnadóttir et al.* [2005], modeling GPS measurements of the postseismic deformation following two large earthquakes in the South Iceland seismic Zone (SISZ) in 2000 assuming viscoelastic relaxation. They found that a lower crustal viscosity of 1×10^{19} Pa s and an upper mantle viscosity of 3×10^{18} Pa s best fit the data. Our results suggest viscosity within the range of the above studies, consistent with glacio-isostatic response taking place both in the lower crust and upper mantle. A somewhat higher viscosity estimate for the lower crust and upper mantle, in the range 4×10^{19} Pa s, is suggested by *LaFemina et al.* [2005] based on measurements and modeling of inter-rifting deformation in South Iceland. They conclude, however, that their GPS velocity field reflects low strain rate processes, and that a higher viscosity estimate is expected in that case (late stage of a rifting cycle) compared to studies based on data acquired during higher strain rate processes (postseismic, postrifting).

[39] The viscosity under Iceland is lower than the global average. *Sigmundsson* [1991, 2006] suggested that low viscosity values are not surprising, as Iceland is located on a hot spot on the mid-Atlantic Ridge with high heat flow and recent volcanism. Similar conditions may be found beneath the Basin and Range province of western

North America, where a viscosity on the order of 10^{18} Pa s in the uppermost mantle (30–200 km) is attributed to high heat flow, recent volcanism and extensional tectonics [*Kaufmann and Amelung*, 2000]. Current uplift rates up to about 30 mm/yr, similar to those observed in Iceland, have also been documented in southern Alaska, where the uplift began since the collapse of the Glacier Bay Icefield around the year 1770 [*Larsen et al.*, 2005]. The viscosity regime in Iceland is consistent with a "weak mechanical regime" as presented by *Ivins and James* [1999]. They suggest that low viscosities are typical of regions experiencing tectonic extension and volcanism, such as the Basin and Range province and Iceland. The authors speculate that low viscosities may exist in southern Patagonia due to the subduction process and the release of H_2O at depth, creating a low mantle viscosity there [*Ivins and James*, 1999].

8. Conclusions

[40] Extensive geodetic measurements indicate that glacio-isostatic adjustment is an ongoing process at Vatnajökull, causing uplift rates up to 25 mm/yr at sites near the edge of the ice cap. Finite element modeling of this process, using an Earth model with uniform Maxwell viscoelastic material underlying an elastic layer, favor viscosities in the range of $4-10 \times 10^{18}$ Pa s, assuming an elastic thickness between 10 and 20 km. A uniform viscoelastic material is considered, thus our viscosity estimate corresponds to the effective value for the lower crust and upper mantle. Our viscosity estimate is also consistent with results obtained by modeling post rifting relaxation in northern Iceland [*Pollitz and Sacks*, 1996] as well as post seismic relaxation study in southern Iceland [*Árnadóttir et al.*, 2005]. On the basis of our current model of the rheological structure beneath Iceland, we can predict future uplift rates at the edges of Vatnajökull. If ice volume decrease due to global warming proceeds at similar rates as at present, a minimum uplift of 2.5 meters between 2000 to 2100 is expected at the edges of the ice cap. An uplift of 3.7 m is, however, anticipated if the average thinning rates this century were to double from 2000 and remains unchanged until 2100.

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