Viscosity of the asthenosphere from glacial isostatic adjustment
and subduction dynamics at the northern Cascadia subduction zone,
British Columbia, Canada

Thomas S. James,1,2 Evan J. Gowan,3,4 Ikuko Wada,3,4 and Kelin Wang1,2

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[1] Late glacial sea level curves located in the Cascadia subduction zone (CSZ) fore arc in southwestern British Columbia show that glacial isostatic adjustment (GIA) was rapid when the Cordilleran Ice Sheet collapsed in the late Pleistocene. GIA modeling with a linear Maxwell rheology indicates that the observations can be equally well fit across a wide range of asthenospheric thicknesses, provided that the asthenospheric viscosity is varied from $3 \times 10^{18}$ Pa s for a thin (140 km) asthenosphere to $4 \times 10^{19}$ Pa s for a thick (380 km) asthenosphere. Present-day vertical crustal motion predicted by the GIA models shows rates of a few tenths of a millimeter per year, consistent with previous analyses. The model viscosities largely pertain to the viscosity of the oceanic mantle beneath the subducting Juan de Fuca slab but include a contribution from the mantle wedge above the slab. For comparison, effective viscosities for the upper mantle due to the tectonic regime (subduction) were computed using the strain rates and temperatures of an independent geodynamic model of the CSZ with a wet olivine power law rheology. The effective viscosities agree well with GIA model viscosities of $10^{19}$ Pa s or less, corresponding to an asthenosphere of 100 or 200 km thickness. The agreement suggests a significant role for power law flow in the GIA response. Regardless of the microphysical mechanisms responsible for the GIA response, the viscosity values inferred from GIA can be applied to studies of the megathrust earthquake cycle because both processes take place on comparable time scales.


1. Introduction

[2] During the last major phase of continental glaciation, the Cordilleran Ice Sheet (CIS) covered most of British Columbia and parts of northern Washington State (Figure 1) [e.g., Clague and James, 2002]. The weight of the ice sheet caused tens to hundreds of meters of crustal depression along coastal areas in southwestern British Columbia [Mathews et al., 1970; Clague et al., 1982]. During this time, relative sea level was higher than present despite global sea level being many tens of meters lower. As the ice sheet melted, the crust quickly rebounded and relative sea level fell owing to glacial isostatic adjustment (GIA).

[3] Southwestern British Columbia is located above the active Cascadia subduction zone (CSZ) where the North American plate overrides the young (6–9 Ma) and warm Juan de Fuca plate (Figure 1). Relative sea level and relic lake-shoreline tilt observations indicate that the viscosity of the shallow mantle in this tectonically active region is low compared to cratonic areas [Mathews et al., 1970; Thorson, 1989; James et al., 2000; Clague and James, 2002], but the uncertainty of the inferred mantle viscosity spans an order of magnitude or greater. Relative sea level observations recently obtained to improve the record of postglacial crustal movement in southwestern British Columbia indicate rapid sea level fall immediately following deglaciation [James et al., 2002; Hutchinson et al., 2004; James et al., 2005, 2009]. Preliminary analyses of the isostatic depression derived from these sea level observations confirm the low mantle viscosity but do not help to improve the uncertainties.

[4] Investigations are underway to observe and model the crustal motions in the earthquake-prone CSZ [e.g., Dragert et al., 1994; Rogers and Dragert, 2003; Mazzotti et al., 2003; Kao et al., 2005]. Models of subduction zone dynamics, including those of subduction zone thermal structure [e.g., Currie et al., 2004; Wada et al., 2008] and the great earthquake cycle of interseismic strain accumulation, coseismic displacement, and postseismic relaxation [e.g., Wang et al., 2003; Wang, 2007], incorporate mantle-viscosity parameters. These parameters can be tuned, on one hand, or compared, on
The other hand, to viscosity values inferred from GIA modeling of sea level observations. The goal of this paper is to determine the viscosity of the shallow mantle ("asthenosphere") at the northern Cascadia subduction zone. GIA modeling was carried out to explain the newly collected, precise relative sea level observations. The viscosity inferred from the GIA modeling was then compared to that predicted from a model of the tectonic regime that incorporates the dislocation creep of olivine [Wada et al., 2008].

2. Previous Regional GIA Modeling

Models constrained by observed tilts of relict proglacial lake shorelines from Puget Sound and available relative sea level data from the Strait of Georgia predicted the upper mantle viscosity in southwestern British Columbia to be $5 \times 10^{19}$ Pa s to $5 \times 10^{19}$ Pa s [James et al., 2000]. A lithospheric thickness of 35 km was assumed, which corresponds to the depth to the top of the subducting oceanic slab in Puget Sound. To obtain an upper bound on crustal uplift rates, James et al. [2000] employed a mantle viscosity of $10^{20}$ Pa s and estimated that present-day uplift rates due to ongoing GIA are less than 0.1 mm \(a^{-1}\).

Clague and James [2002] reviewed the history of the Cordilleran Ice Sheet (CIS), and then focused on the fit of the model to relative sea level observations from the Strait of Georgia. They used more realistic Earth models in which viscosity increased with depth, and assumed a thicker 60-km-thick lithosphere, which corresponds to the depth to the top of the subducting slab in the central Strait of Georgia. Their model indicated that a mantle viscosity of less than $10^{20}$ Pa s is needed to give a good fit to the data and that present-day uplift rates on Vancouver Island are less than 1 mm \(a^{-1}\).

Clague et al. [2005] modeled sea level for the early phases of the Fraser glaciation. Substantial ice was added to the existing ice sheet model in the Strait of Georgia in early times in order to match the sea level observations. The later stages of the ice sheet model were not altered.

Early global GIA models (ICE-3G [Tushingham and Peltier, 1991] and ICE-4G [Peltier, 1994]) assumed an upper mantle viscosity of $10^{21}$ Pa s. As a consequence of the relatively high viscosity, model-predicted relative sea level falls too slowly following deglaciation. The VM2 earth model used with ICE-5G [Peltier, 2004] employs an average upper mantle viscosity of about $4 \times 10^{20}$ Pa s. In the following, we will demonstrate that even this smaller value must be further reduced in the shallow parts of the upper mantle in order to explain sea level observations from the northern CSZ.

3. Earth Models

GIA models include a model of ice sheet growth and decay and a model of the Earth’s response to surface loading. The Earth models considered here are spherically symmetric and each layer has uniform elastic properties, density, and viscosity. The models have an elastic lithosphere, an incompressible Maxwell viscoelastic mantle divided into several layers, and a fluid core. Density and rigidity parameters are
average values from the Preliminary Reference Earth Model [Dziewonski and Anderson, 1981]. The Earth models do not capture the complexities of Earth structure in the CSZ but, instead, represent average quantities.

[11] The viscosity of the mantle is averaged from the VM2 model [Peltier, 2004], except that the asthenospheric thickness and viscosity are allowed to vary (Figure 2). Asthenospheric thicknesses range from 140 to 380 km. Below the asthenosphere and above the 670-km discontinuity, the mantle viscosity is $4 \times 10^{20}$ Pa s. The lower mantle has a viscosity of $1.6 \times 10^{21}$ Pa s above 1291 km depth and $3.2 \times 10^{21}$ Pa s below that.

[12] Surface heat flow in the CSZ fore arc is low [Hyndman and Lewis, 1995], suggestive of a cool, stagnant mantle wedge [Hyndman and Wang, 1995; Wada et al., 2008]. We assume that on the time scale of the GIA, the fore-arc material overlying the subducting oceanic lithosphere responds elastically without a viscous component and is decoupled from the subducting oceanic lithosphere, and we employ a 60-km lithospheric thickness. This value corresponds to the depth to the top of the subducting slab below the western Strait of Georgia and is the same thickness used in earlier modeling [Clague and James, 2002]. The response of the Earth models to surface loading was calculated using methods described by James and Ivins [1998].

4. Modeling Results

4.1. Ice Model

[13] Our ice sheet history model was modified from the sequence originated by James et al. [2000] and revised by Clague and James [2002] and Clague et al. [2005]. In the vicinity of the sea level observations, the original grid elements of the ice sheet model were divided into four equal areas to provide increased sensitivity to the retreat history (Figure 3). The finer grid also allows the ice thickness to reflect the high variability of the topography. The ice sheet model was first modified so that the predicted sea level curves fit the magnitude of the high stands. This increased the size of the ice load, especially in the Victoria area, a region that previously had not been modeled. The timing of deglaciation was then adjusted to provide rapid sea level fall at the same time as the observations.

[14] At the glacial maximum, the Cordilleran Ice Sheet flowed smoothly over topography [Booth, 1987], which indicates that there was significantly more ice in low-lying areas than in areas of high elevation. The isostatic depression was taken into account when adjusting the ice model to produce an ice surface that gently slopes toward the Pacific Ocean. At the CIS maximum, the model features more than 2500 m of ice in the Strait of Georgia, in contrast to less than 1500 m of ice on central Vancouver Island. This reflects the bathymetry of the Strait of Georgia, which reaches depths greater than 200 m, and the topography of Vancouver Island, which has substantial regions that are higher than 1000 m.

[15] Finally, the model was refined to fit available radiocarbon constraints on deglaciation of the southwestern sector of the Cordilleran Ice Sheet (CIS) [Gowan, 2007]. In general, the model indicates that the ice retreat began several hundred years before the oldest ages in an area and ice has completely melted 100–200 years before the oldest ages.

[16] The final ice model consists of twenty time periods starting at 35,000 calendar years before present (cal years B.P.). The deglaciation of Puget Sound, the Strait of Juan de Fuca, and the Strait of Georgia is divided into eight time periods between 15,200 and 13,900 cal years B.P. Between 13,900 and 12,900 cal years B.P., there is significant ice loss on all grid elements. Between 12,900 and 11,500 cal years B.P., there is an increase of ice thickness on grid elements in mainland British Columbia to simulate readvances that happened during this time. Since the sea level data from this time period are of low resolution, there was no attempt to simulate in detail the various readvances and retreats discussed by Clague et al. [1997], Kovanen and Easterbrook [2002], and Friele and Clague [2002]. After 10,700 cal years B.P., the ice model has no ice on any grid element, indicating that present glacial levels have been reached.

4.2. Sea Level Modeling

[17] Using the refined ice sheet model described above, relative sea level was predicted for the three localities in southwestern British Columbia that have well constrained sea level curves. The root mean squared (RMS) error misfit of the predicted sea level curves to the observed curves is used to determine how well the predicted sea level matches the constraints:

$$\text{RMS} = \sqrt{\frac{1}{n} \sum_{i=1}^{n} (\text{true SL} - \text{pred SL})^2}.$$

Relative sea level observations have recently been collected in the Strait of Georgia and the Victoria area to augment the previous sparse record and facilitate more detailed GIA modeling [James et al., 2002; Hutchinson et al., 2004; James et al., 2005; Gowan, 2007; James et al., 2009] (Figures 1 and 4). The sea level high stand occurred between about 14,000 (Victoria) to 13,750 (northern Strait of Georgia) cal years B.P. Sea level in Victoria fell from a high stand position of 75 m above sea level to present-day sea level in about 800 years [James et al., 2009]. After falling to a probable low stand position around −30 m, sea level returned to within 4 m of present by about 6000 cal years B.P. Direct observations in the Victoria area only constrain the low stand to lie below −11 m and above −40 m, and this is the range shown in Figure 4. Data from the central Strait of Georgia indicate that
sea level fell rapidly from a high stand position of about 150 m to below present sea level in 1500–2000 years [Hutchinson et al., 2004]. After reaching a low stand of about –15 m, sea level returned to near present levels by about 8000 cal years B.P. Sea level in the northern Strait of Georgia fell from a high stand position of about 175 m to present levels in less than 3000 years [James et al., 2005].

[18] The sea level curves are well determined from 13,700 to 12,000 cal years B.P. for the northern Strait of Georgia, 13,900 to 12,300 cal years B.P. for the central Strait of Georgia and 14,100 to 13,100 cal years B.P. for Victoria. These time intervals are used for determining the RMS fit. An unweighted RMS fit is appropriate because the sea level curves are equally well constrained for these time intervals. The uncertainty on the timing of sea level fall at each locality is estimated to be about ±100 years.

[19] The magnitude and timing of the subsequent low stands and recovery to present levels in the mid-Holocene are less well determined. The mid and Late Holocene segments of the sea level curves may also include a significant tectonic component [e.g., Hutchinson, 1992]. For these reasons, RMS calculations were carried out only for the time intervals indicated above and do not include the low stand portions of the sea level curves.

[20] Predicted sea level curves are compared to observations in Figure 4 for Earth models with different asthenospheric thicknesses and the asthenospheric viscosity chosen to optimize the fit to the observations. The predictions agree well with the observations. They feature a very similar rapid fall in sea level after deglaciation between 14,000 and 12,000 cal years B.P. For the central Strait of Georgia (CSG), the Earth models with a thicker asthenosphere have a poorer fit than the thin asthenosphere models. A slight adjustment to an earlier time of deglaciation would improve the agreement for the thick asthenosphere models. Overall, the models successfully predict the rapid sea level fall observed in the first 1000 to 1500 years after deglaciation.

[21] From 12,000 to 10,000 cal years B.P., there are slight differences in the predicted responses that depend on the asthenospheric thickness and viscosity. In general, however, the features of the low stand calculation are consistent among the model predictions. The modeled level low stand position is about –35 m in Victoria, which is close to the maximum low stand scenario of –40 m and slightly deeper than the preferred low stand of –30 ± 5 m proposed by James et al. [2009]. The predicted low stand position of about –20 m in the central and northern Strait of Georgia is deeper than the available observations would indicate [Hutchinson et al.,
By 9000 cal years B.P., the modeled sea level at all locations approaches the same value, indicating that the crustal response has diminished to the same (small) level at all three locations. It is possible that changes to the assumed VM2 viscosity structure below the asthenosphere could produce a better fit to the low stand portions of the sea level curves.

Contour plots of the RMS misfit are given for a range of asthenospheric thicknesses and viscosities in Figure 5. There is no one thickness/viscosity pairing that produces the best fit. Instead, relatively uniform, good fits are obtained across the range of asthenospheric thicknesses that was considered. The optimal viscosity ranges from $3 \times 10^{18}$ (140-km-thick asthenosphere) to $4 \times 10^{19}$ Pa s (380-km-thick asthenosphere). For a given asthenospheric thickness, the RMS misfit rapidly increases away from the optimal viscosity value. An RMS misfit of 16 m provides fits to the observations that are visibly poorer than the minimum misfit values of ~8 m.

Overall, the optimal misfit range is tighter for the central and northern Strait of Georgia than for Victoria. This is likely the result of the Strait of Georgia curves being well constrained over a longer time period. The misfit of the Victoria sea level curve is slightly better with a thicker asthenosphere. This may indicate that an asthenospheric thickness greater than 220 km is to be preferred, although the fit for a thinner asthenosphere could be improved by slightly revising the timing of ice sheet retreat. An asthenospheric viscosity greater than $10^{20}$ Pa s or less than $10^{18}$ Pa s is ruled out for asthenospheric thicknesses ranging from 140 to 380 km.

The viscosity of the asthenospheric channel can be estimated for a thinner asthenosphere. For the case where viscous flow is confined to a thin layer, a diffusion constant ($D$) for channel flow is given by

$$D = \frac{\rho g H^3}{3\eta}$$
where \( \rho \) is density, \( g \) is the acceleration due to gravity, \( H \) is the thickness of the channel, and \( \eta \) is the channel viscosity [Cathles, 1975]. The rate of crustal response is determined by the diffusion constant. Solving for viscosity \( \eta \), we have

\[
\eta = \frac{\rho g H^3}{3D}
\]

Relation (3) suggests a linear relationship between the logarithms of asthenospheric viscosity and thickness. Figure 6 shows that the expected linear relationship starts to break down above 300-km thickness, suggesting that the assumption of channel flow is less valid at larger thicknesses. A linear regression through the other points provides an extrapolation to a thinner asthenosphere. A 50-km-thick asthenosphere would require a viscosity of about \( 2 \times 10^{17} \) Pa s and a 10-km-thick asthenosphere would need a viscosity substantially less than \( 10^{16} \) Pa s in order to explain the sea level observations.

4.3. Current Crustal Uplift Rates

[25] The model-predicted crustal uplift rates due to GIA for the ice load history are given in Figure 7a for an asthenospheric thickness and viscosity of 300 km and \( 2.5 \times 10^{19} \) Pa s, respectively. The present-day uplift rate on Vancouver Island is predicted to be about 0.1 mm a\(^{-1}\). The uplift predicted by a variant of the ICE-3G deglaciation model [Tushingham and Peltier, 1991], without the southwestern British Columbia ice history included, generates uplift rates of 0.3 to 0.35 mm a\(^{-1}\) on Vancouver Island for the same viscosity structure (Figure 7b). The sum of the two predictions, which includes the effect of distant ice sheets as well as the local Cascadia model, is given in Figure 7c and features uplift rates on Vancouver Island of 0.4 to 0.45 mm a\(^{-1}\).

[26] These predictions are consistent with recent analyses of the postglacial isostatic depression at Victoria [James et al., 2009] and the northern Strait of Georgia [James et al., 2005], in which the present-day rate of crustal uplift was estimated to be 0.12 and 0.25 mm a\(^{-1}\), respectively. Although the model predictions (Figure 7c) given here are larger by factors of two to four, the magnitude of all the predictions is only a few tenths of a millimeter per year and is below the resolution of GPS observations. This is in contrast to other regions where current postglacial uplift rates reach magnitudes of many millimeters per year (e.g., Churchill, Manitoba, \( \sim 10 \) mm a\(^{-1}\) [Lambert et al., 2006]) and, in circumstances that feature recent ice mass wastage and low mantle viscosity, a few centimeters per year (e.g., Glacier Bay, Alaska, \( 30 \) mm a\(^{-1}\) [Larsen et al., 2005]).
Measured present-day crustal uplift rates from Global Positioning System (GPS) observations along the west coast of North America (from northern California to the southern Alaska panhandle) range from about $-2$ to $4$ mm a$^{-1}$ [Mazzotti et al., 2008]. The GPS observations have substantial scatter that may reflect inherent uncertainties in the GPS time series or local site effects, but spatially coherent patterns of vertical crustal motion suggest that there are real regional variations at the level of a few millimeters per year. The small GIA rates predicted here from the late Pleistocene deglaciation of the CIS indicate that other processes, such as active tectonics, must be largely responsible for the variations in current vertical crustal motion.

5. Implications for Upper Mantle Viscosity in the Cascadia Subduction Zone

Are the asthenospheric viscosities inferred from GIA modeling with linear Maxwell viscoelastic Earth models consistent with laboratory-constrained, nonlinear, temperature-dependent flow properties of mantle minerals? The sea level observations are located in the CSZ fore arc where crustal heat flow is low [Hyndman and Lewis, 1995] and the subducting oceanic lithosphere is 40 to 70 km deep (Figure 1). Geodynamic modeling of the CSZ that explains features of the observed heat flow, depth of crustal seismicity, and the relative paucity of arc volcanism shows that the sea level sites are underlain by a stagnant mantle wedge that reaches about 600°C at 70 km depth [Wada et al., 2008] (Figure 8a). Beneath the fore-arc mantle wedge lies the subducting oceanic lithosphere, which reaches temperatures of 1000°C at about 100 km depth. A few tens of kilometers further landward of the sea level observations, temperatures of 1000°C are reached at shallower depths of about 60 km in the arc and back-arc regions of the subduction zone.

The sea level sites are located in the boundary zone between the oceanic and continental plates. It is likely that the observed rapid sea level response is primarily a consequence of the viscosity of the oceanic mantle that lies directly beneath the sites, but the ice sheet extended east over continental mantle, and the observed response must also have a contribution from the continental mantle wedge. Thus, in the following we consider the viscosity of the subducting oceanic mantle, as well as the hot, circulating arc and back-arc continental mantle wedge.

5.1. Effective Viscosity of the Mantle Derived From Nonlinear Flow Laws

Karato and Wu [1993] synthesized laboratory studies and geological and geophysical observations and found that the upper mantle is likely to deform by power law (dislocation) creep. Nonlinear, temperature-dependent plastic flow arising from dislocation creep is given by

$$\dot{\varepsilon} = A_0^\sigma \exp \left( -\frac{Q + PV}{RT} \right)$$

Figure 7. Uplift rates (mm a$^{-1}$) calculated for (a) the ice sheet model developed here loading an Earth model with a lithosphere thickness of 60 km, asthenosphere thickness of 300 km, and asthenosphere viscosity of $2.5 \times 10^{20}$ Pa s, (b) the ICE-3G model with southwestern British Columbia masked, using the same Earth model as Figures 7a, and (c) the sum of Figures 7a and 7b.
Figure 8. (a) Temperatures (°C), (b) strain rates (s⁻¹), and (c) effective viscosities (Pa s) for the preferred Cascadia subduction model of Wada et al. [2008, Figure 6c]. For the oceanic mantle, effective viscosities were obtained assuming a tectonic strain rate of $5 \times 10^{-15}$ s⁻¹, as discussed in section 5.2. Labeled sites (arrows) 1, 2, and 3 correspond to the approximate location of the Victoria, central Strait of Georgia, and northern Strait of Georgia sea level curves, respectively. The location of arc volcanism is given by the triangle. The gray shading in the overriding and subducting mantle corresponds to the stability zone of the serpentine mineral antigorite.

where $\dot{\varepsilon}$ is the flow strain rate, $A$ is a material parameter, $\sigma$ is the stress, $n$ is the power law exponent ($n = 3–3.5$ for the materials considered here), $Q$ is the activation energy, $R$ is the gas constant, and $T$ is temperature [e.g., Evans and Kohlstedt, 1995]. The dependence on the pressure $P$ and activation volume $V$ is relatively small and is neglected here. We also ignore the tensorial nature of stress and strain, as the focus of the discussion is on the magnitudes. Defining an effective viscosity $\eta_{\text{eff}} = \sigma/2\dot{\varepsilon}$, we have

$$\eta_{\text{eff}} = \frac{1}{2} \dot{\varepsilon}^{(1-n)/n} A^{-1/n} \exp\left(\frac{Q}{nRT}\right)$$  \hspace{1cm} (5)$$

With values of $n$ typically in the range of 3 to 3.5, the exponent of the strain rate $\dot{\varepsilon}$ takes on values of $-0.67$ to $-0.71$. Thus, a larger strain rate delivers a smaller effective viscosity, if other factors are held constant.

[32] If two processes, such as GIA and subduction, are both occurring, then the determination of the effective viscosity requires additional attention [e.g., Schmeling, 1987; Ranalli, 1995]. Assuming that the two processes are occurring with a subduction strain rate $\dot{\varepsilon}_S$ and GIA strain rate $\dot{\varepsilon}_G$, then the effective viscosity depends on the total strain rate $\dot{\varepsilon} = \dot{\varepsilon}_S + \dot{\varepsilon}_G$. With reference to equation (5), if one strain rate is much larger than the other, then the effective viscosity is determined by the larger strain rate. In this case, the process occurring at slower strain rates does not influence the effective viscosity greatly, and the slower process will proceed as though operating under a linear flow law. On the other hand, if the two strain rates are similar in magnitude, then the effective viscosity is scaled by a factor of $\dot{\varepsilon}_S^{(1-n)/n} (= 0.6$ for $n = 3.5$) relative to the effective viscosity that would be determined if only one of the processes were active.

[33] In the following, we first determine the effective viscosities for subduction and GIA separately, and then we discuss the effective viscosity when GIA and subduction are considered jointly.

5.2. Effective Viscosity of the Upper Mantle From Tectonic Processes

[34] For the preferred Cascadia model of Wada et al. [2008, Figure 6c] (and their associated discussion), strain rates in the convecting mantle wedge due to subduction reach $10^{-14}$ s⁻¹ at depths of 70 to 90 km and are above $10^{-15}$ s⁻¹ below about 60 km depth (Figure 8b). Above the subducting slab, at depths greater than 70 km, effective viscosities are mostly in the range of $10^{18}$ Pa s to $10^{19}$ Pa s (Figure 8c). These values are attained to the east (landward) of the sea level observations.

[35] The model of Wada et al. [2008] predicts the thermal structure of the subducting oceanic lithosphere but does not consider its deformation. To calculate the effective viscosity of the oceanic mantle from equation (5) with the predicted temperature requires an estimate of the strain rate at which the oceanic mantle undergoes strain. We derive the estimate in two different ways:

[36] 1. Assuming the mantle beneath the subducting Juan de Fuca plate is moving with the oceanic plate, its strain rate can be derived by noting that beneath the study region the subducting slab sinks 40 km as it travels 100 km eastward in about 2.4 Ma (assuming a convergence rate of 4.5 cm a⁻¹). This gives a strain rate of $40 \text{ km}/100 \text{ km}/2.4 \text{ Ma} = 5 \times 10^{-15}$ s⁻¹ for the oceanic mantle in the vicinity of the CSZ.

[37] 2. Regionally, the Juan de Fuca plate is a few hundred kilometers wide (ranging from about 350 to 650 km from the Juan de Fuca ridge to where it reaches about 100 km depth in the CSZ). The Juan de Fuca ridge is spreading at about 6 cm a⁻¹, while the Juan de Fuca plate is subducting at about 4.5 cm a⁻¹. Assuming an average width of 500 km and assuming similar upwelling and downwelling rates in the upper mantle of 5 cm a⁻¹ gives a strain rate of $10 \text{ cm} a^{-1}/500 \text{ km} = 6 \times 10^{-15}$ s⁻¹. The regional oceanic mantle strain rate is essentially identical to the strain rate estimated for the oceanic mantle at the CSZ.
Table 1. Flow Law Parameters

<table>
<thead>
<tr>
<th>Material</th>
<th>A (s⁻¹ Pa⁻²)</th>
<th>n</th>
<th>E (kJ mol⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dry Olivine</td>
<td>4.85 x 10⁻¹⁷</td>
<td>3.5</td>
<td>535</td>
</tr>
<tr>
<td>Wet Olivine</td>
<td>4.89 x 10⁻¹⁵</td>
<td>3.5</td>
<td>515</td>
</tr>
<tr>
<td>Serpentinite</td>
<td>4.47 x 10⁻¹⁸</td>
<td>3.8</td>
<td>8.9</td>
</tr>
</tbody>
</table>

*Hirth and Kohlstedt [1996].
*Hilairet et al. [2007].

[38] In discussing the generation of mid-ocean ridge basalts and the evolution of the oceanic lithosphere and mantle, Hirth and Kohlstedt [1996] considered the oceanic mantle to be composed of dry olivine (dunite) above about 60 km depth and wet olivine below. Using their flow law parameters for the two materials (Table 1) and temperatures given in Figure 8a and assuming a uniform strain rate 5 x 10⁻¹⁵ s⁻¹ gives effective viscosities for the subducting oceanic mantle (Figure 8c). If oceanic mantle deformation is localized in a thin, low-viscosity layer, perhaps arising from the slab sliding over a layer containing partial melt, then strain rates in this part of the oceanic mantle could be larger than assumed here and the effective viscosities would be smaller. Thus, the oceanic mantle effective viscosities shown in Figure 8c are a probable upper bound.

5.3. Effective Viscosity of the Upper Mantle From GIA

[39] A representative GIA strain rate during the time of the fastest response can be derived by noting that following deglaciation, the central Strait of Georgia rose about 35 m relative to Victoria in 1000 years [James et al., 2009]. The sites are 140 km distant so the strain rate is about 10⁻¹⁴ s⁻¹. This is the peak rate, and it applies to the mantle near the changing surface load during the time of fastest rebound. At greater distances (or depths) from the changing load, or at later times, GIA strain rates were lower.

[40] Effective viscosities were computed for uniform strain rates of 10⁻¹⁴ and 10⁻¹⁵ s⁻¹ (Figures 9a and 9b) using equation (5), the temperatures given in Figure 8a, and the parameters appropriate to wet and dry olivine (dunite) [Hirth and Kohlstedt, 1996]. For the hot, actively convecting mantle wedge, a wet dunite rheology was assumed. For the oceanic mantle we have adopted the structure that was assumed in section 5.2.

[41] For a strain rate of 10⁻¹⁴ s⁻¹, the effective viscosity of the continental mantle wedge is less than 10²⁰ Pa s below about 60 km depth and less than 10¹⁹ Pa s below about 80 km depth (Figure 9a). In the oceanic mantle, the dry dunite rheology leads to viscosities larger than about 10¹⁹ Pa s for distances less than about 40 km from the slab top. At 50 km, the increasing temperatures cause the viscosity to drop to about 5 x 10¹⁹ Pa s. The assumed transition to a wet dunite rheology at greater depths then leads to a sharp reduction in viscosity to values of 5 x 10¹⁸ Pa s and lower. Overall, the viscosity of the oceanic mantle is about a factor of 2 less than that predicted for the arc and back-arc mantle wedge. The presence of volatiles and melt in the arc and back-arc mantle wedge would be expected to reduce effective viscosity and the values shown in Figure 9 are a probable upper bound.

[42] The effective viscosity is a factor of 5 larger when the strain rate is decreased by a factor of 10 to 10⁻¹⁵ s⁻¹ (Figure 9b). A smaller strain rate may be more representative of conditions further away from the imposed surface load, and would also relate to the reduced rate of response some time after the load was removed. Deep oceanic mantle viscosities are less than about 3 x 10¹⁹ Pa s, and the mantle wedge viscosity is less than about 5 x 10¹⁹ Pa s at depths greater than about 80 km.

5.4. Stagnant Mantle Wedge

[43] We also considered whether the cold, stagnant fore-arc mantle wedge could have a viscous response on GIA time scales. Tectonic strain rates here are much smaller and GIA strain rates dominate. If the stagnant mantle wedge were to contribute significantly to the sea level response, viscosities appropriate to a channel with a thickness in the range of 10 to 40 km would be required. As discussed in section 4.2, this would imply effective viscosities in the range of 10¹⁵ to 10¹⁷ Pa s (Figure 6).

[44] Petrological considerations and seismic observations suggest the fore-arc mantle is heavily serpentinized [Bostock et al., 2002; Hyndman and Peacock, 2003; Brocher et al., 2003; Ramachandran et al., 2005]. Using flow law parameters for antigorite serpentine [Hilairet et al., 2007] (Table 1), which is at least 1 order of magnitude weaker than any other major mantle mineral, the effective viscosity of the stagnant mantle wedge ranges from 10²⁰ Pa s to 6 x 10²⁰ Pa s for
At greater distances and depths from the shrinking ice mass, GIA strain rates were lower and GIA effective viscosities were higher. For this case, the tectonic effective viscosities would dominate, and the sea level observations would sample effective viscosities of $10^{18}$ Pa s and lower.

The effective tectonic viscosities are the same as the lower end of the viscosity range determined from the GIA modeling. The sea level observations could be fit well with an asthenosphere having a viscosity of $3 \times 10^{18}$ Pa s and thickness of 140 km, $10^{19}$ Pa s at 220 km, and $4 \times 10^{19}$ Pa s at 380 km. Thus, an asthenosphere of 100- or 200-km thickness and a viscosity between $10^{18}$ and $10^{19}$ Pa s fits the sea level observations well and is consistent with experimentally determined nonlinear dislocation flow laws for wet olivine deforming at rates consistent with the subduction of the Juan de Fuca oceanic plate. Below this layer lies higher-viscosity upper mantle.

5.6. GIA Model Viscosity: Oceanic or Continental Mantle?

Both the effective viscosity of the oceanic mantle and the hot, circulating arc and back-arc mantle wedge agree with the viscosity range obtained from the GIA modeling. This is consistent with both regions contributing to the observed response, although the oceanic mantle directly underlying the observations would respond directly to changes in surface load brought about by deglaciation, and thus would be expected to contribute the majority of the response.

The Strait of Georgia sea level sites are located about 70 km from where surface heat flow rises sharply near the volcanic arc [Hyndman and Lewis, 1995] and low-viscosity continental mantle is expected. Spatial wavelengths of the Earth’s response to the ice load that are longer than 50–100 km would be sensitive to the viscosity of the back-arc (continental) mantle. As well, relative to the sea level sites, the ice sheet extended a much greater distance eastward past the subduction arc into the back-arc region than it extended westward. Thus, long spatial wavelengths would sample continental mantle more than oceanic mantle, although they also sample deeper into the mantle and the effect of a shallow, thin, low-viscosity layer is thus reduced. These considerations suggest that the GIA model viscosities are a blend of oceanic and continental mantle viscosity, but that the oceanic mantle contributed most of the response.

5.7. GIA and a Transient Mantle Rheology

Instead of GIA sampling a power law medium whose viscosity is conditioned or determined by subduction-related processes, it is possible that GIA is sampling a transient rheology that is independent of the long-term deformation. Karato [1998] has suggested that because dislocation density is proportional to stress, a power law medium undergoing deformation from two processes will not simply respond as the sum of the deformation rates as described in section 5.1 (equation (5)). Instead, deformation by the more ephemeral process (GIA) will proceed by a transient creep mechanism. Ranalli [1998] also put forward arguments suggesting that dislocation creep (power law) and diffusion creep (linear) almost balance. He introduced the concept of a transition stress marking the boundary between predominantly linear creep (low stress) and power law creep (high stress).
[51] We do not rule out the possibility that GIA occurs primarily through a transient creep mechanism at the CSZ. It is possible that the correspondence of the GIA model viscosities, GIA effective viscosity, and tectonic effective viscosity is accidental. Nevertheless, the agreement suggests to us that power law creep plays an important role in the deformation induced by GIA. Karato’s [1998] argument depends on a uniform population of dislocations, all of which are participating in the deformation related to the background stress. But the mantle is heterogeneous on many spatial scales. It is possible that the imposition of an additional stress from GIA could lead to the activation of additional dislocations and a power law response as described in section 5.1. We also note that a power law rheology explains features of short-term transient deformation following large earthquakes [e.g., Freed and Bürgmann, 2004; Freed et al., 2006], although postseismic observations can be explained using a transient rheology [e.g., Pollitz, 2003; Pollitz et al., 2006] and other processes (poroelasticity, after slip) may also be at play.

[52] Although transient creep may be at work for monthly to decadal-scale postseismic deformation following large earthquakes, its importance in the postglacial response at the timescale of hundreds to thousands of years is presently unclear. Given the agreement between effective viscosities inferred from GIA and long-term subduction processes, we think there is good reason to believe that a power law rheology is important on the GIA time scale. A complete resolution of this issue is beyond the scope of the paper. Explicit GIA modeling at the CSZ with a power law rheology and including subduction zone processes would be a useful next step.

5.8. Implications for Models of the Megathrust Earthquake Cycle

[53] As reviewed by Wang [2007], models are being actively developed to explain deformation features of the megathrust earthquake cycle. One observed feature is that the location of the largest coseismic subsidence, which tends to occur approximately above the landward termination of the locked part of the subduction interface, is also where the largest postseismic uplift occurs [e.g., Thatcher, 1984; Cohen and Freymueller, 2001]. This feature can be replicated in megathrust earthquake cycle models by assuming, below the strong oceanic lithospheric lid a few tens of kilometers thick, an oceanic mantle with long-term effective viscosity around $10^{19}$ Pa s. Immediately after the earthquake, loading of the locked zone recommences and for this choice of oceanic mantle viscosity, postseismic uplift occurs. For a smaller oceanic mantle viscosity (e.g., $10^{19}$ Pa s), however, subsidence occurs because the weaker oceanic mantle yields quickly to the increased load of the overriding continental lithosphere caused by coseismic slip along the subduction interface. In this case, to generate uplift, it is necessary to also assume that significant after slip occurs on deeper portions of the subduction fault.

[54] Our GIA modeling of sea level observations from the northern CSZ indicates that the (shallow) oceanic mantle has a viscosity less than $4 \times 10^{19}$ Pa s. Smaller values below about $10^{19}$ Pa s are required if the deformation is proceeding through power law creep. Models developed to explain the great earthquake cycle at subduction zones with young oceanic lithosphere will need to incorporate a low-viscosity oceanic mantle. This may require significant deep after slip in the megathrust earthquake cycle to explain the nature of postseismic deformation.

6. Discussion and Summary

[55] GIA modeling of new relative sea level data has determined a close relationship between assumed asthenospheric thickness and inferred viscosity. The optimal value of the asthenospheric viscosity ranges from $3 \times 10^{18}$ Pa s to $4 \times 10^{18}$ Pa s for asthenospheric thicknesses ranging from 140 km to 380 km. The viscosity values are similar to the $5 \times 10^{18}$ to $5 \times 10^{19}$ Pa s range originally estimated by James et al. [2000] and Clague and James [2002], but the earlier estimates were not based on a confined asthenospheric thicknesses.

[56] We have also calculated effective viscosities of both the subducting oceanic mantle below the assumed dry ‘lid’ and the continental arc and back-arc mantle wedge assuming power law creep and using temperatures from a geodynamic model of the Cascadia subduction zone. Values thus determined are consistent with the lower range of asthenospheric viscosities derived here from GIA modeling. A 100- or 200-km-thick layer with a viscosity of $10^{18}$ Pa s to $10^{19}$ Pa s reproduces the sea level observations and is consistent with the effective viscosities expected at this plate boundary, which features subduction of a young oceanic plate. An alternative interpretation of the GIA modeling results in terms of transient deformation is possible but the agreement between the GIA model viscosities and the GIA and tectonic effective viscosities is consistent with a substantial role for power law deformation in the GIA response.

[57] Regardless of the microphysical mechanisms at work to generate the GIA response, the viscosity values inferred from the GIA modeling can be applied to studies of the megathrust earthquake cycle because there is substantial overlap between the timescales of both processes. The GIA response in northern Cascadia was rapid, featuring an exponential decay time of about 500 years [James et al., 2005], and sea level fell rapidly in 1000 to 2000 years (Figure 4). This time range is comparable to the 500–600 year average repeat time of great megathrust earthquakes at the CSZ [Atwater and Hemphill-Haley, 1997; Goldfinger et al., 2003].

[58] If strain is localized in the oceanic mantle just below the moving lithosphere, perhaps because of the nonlinear rheology, then the effective viscosities for the oceanic mantle based on thermal modeling results may require careful interpretation, because viscous deformation takes place in both the ‘layer’ of higher strain rate and thus lower viscosity and the underlying mantle of lower strain rate and thus higher viscosity. However, the GIA modeling still gives constraints on the viscosity of the asthenosphere. Extrapolation of the viscosities inferred from the GIA modeling (Figure 6) indicates that viscosities around $10^{18}$ Pa s are needed for a 100-km-thick asthenosphere and $10^{19}$ Pa s for a 40-km-thick asthenosphere.

[59] Good fits to GIA observations using linear rheology Earth models have been obtained for many regions. This could indicate the prevalence of a transient linear rheology on GIA time scales, which for many parts of the world is a few
thousand years. Two related studies of the Laurentide post-glacial uplift which compared a composite linear and nonlinear rheology to a purely linear rheology found a slight preference for the composite rheology to explain relative sea level [Gasperini et al., 2004; Dal Forno et al., 2005]. For the composite rheology, the majority of the GIA deformation took place through the nonlinear mechanism.

It could also be that GIA samples linear effective viscosities that are determined by tectonic and mantle convection processes acting on power law materials, as we suggest for the CSZ. This is also suggested by the results of Wu [2001], who found a good fit to relative sea level data in North America with a power law rheology, provided an ambient tectonic stress of 10 MPa was assumed in the mantle. Other studies have reported variable success in explaining GIA observables with a purely nonlinear rheology and no viscosity conditioning by tectonic processes [e.g., Wu, 1998, 2002; Wu and Wang, 2008]. These studies illustrate some of the approaches for the treatment of Earth deformation that have been taken to explain GIA observations.

It would be valuable to determine if linear viscosity models developed to explain GIA observations are also consistent with constraints on heat flow, experimentally determined flow laws, and inferred tectonic and mantle convection strain rates, as described here for the northern CSZ. For example, GIA modeling of the response to Little Ice Age retreat derived an optimal viscosity around 3–4 × 10^{19} \text{Pa s} for a 110-km-thick asthenosphere at Glacier Bay (GB), southeast Alaska [Larsen et al., 2005]. The GB viscosity is only a factor of 3–4 larger than the value extrapolated from our modeling for a 110-km-thick asthenosphere [Brocher et al., 2003]. Sisewed evidence for widespread serpentinized forearc mantle underlies the southern margin of the Cordilleran ice sheet, margin [Geology, 31, 267–270, doi:10.1130/0091-7613(2003)031<0267:SEFWSF>2.0.CO;2].

References


