Rapid uplift of southern Alaska caused by recent ice loss

Christopher F. Larsen,1 Roman J. Motyka,1 Jeffrey T. Freymueller,1 Keith A. Echelmeyer1 and Erik R. Ivins2

1Geophysical Institute, University of Alaska Fairbanks, 903 Koyukuk Dr., Fairbanks, AK 99775, USA. E-mail: chris@giseis.alaska.edu
2Jet Propulsion Laboratory, California Institute of Technology, 4800 Oak Grove Dr., Pasadena, CA 91109, USA

Accepted 2004 April 23. Received 2004 April 16; in original form 2003 September 6

SUMMARY

Extreme uplift rates and sea level changes in southern Alaska have been documented by Global Positioning System (GPS) surveys, tide gauge measurements and studies of raised shorelines. The movements detected in a network of 45 GPS survey points describe a broad pattern of rapid regional uplift. The majority of the study area is uplifting at a rate faster than 10 mm yr\(^{-1}\), with several sites uplifting more rapidly than 25 mm yr\(^{-1}\). New tide gauge data presented here consist of repeat occupations of 18 temporary gauge sites. Sea level rates at these sites agree with similar measurements in southern Alaska taken \(\sim\)50 yr earlier, and also with the pattern of uplift derived from the GPS data. Raised shoreline studies at 14 sites document total sea level change, with a maximum change in sea level of \(-5.7\) m found in upper Lynn Canal. The start of the ongoing uplift episode that raised these shorelines has been dated with dendrochronology and found to be coincident with the start of the collapse of the Glacier Bay Icefield, at \(ca.\) 1750 AD. The pattern of total sea level change is in general agreement with uplift-rate measurements, with greater sea level change found at the sites closest to the peak uplift rates in upper Glacier Bay. We use a viscoelastic earth model subjected to an ice load history built upon observations of glacial change to predict uplift rates at the tide gauge and GPS sites as well as the total uplift at the raised shoreline sites. Our modelling exercises are limited to an ice load model based on independent studies of the region’s glacial history over the past 1.7 kyr, to evaluate whether the uplift observations can be explained by simple earth models subjected to this load history. Two-layer earth models, consisting of an elastic crust and a low-viscosity upper mantle half-space, can be adjusted to fit either the raised shoreline data or the combined GPS and tide gauge uplift-rate data, but cannot fit all the data with a single set of earth model parameters. However, all three data sets are consistent with an approximated three-layer earth model. The combined model is constrained by a total of 77 uplift measurements, which at the 95 per cent confidence level require a low-viscosity asthenosphere \([\eta_A = (1.4 \pm 0.3) \times 10^{20} \text{ Pa s and thickness } 110^{+15}_{-20} \text{ km}]\) beneath a \(50^{+30}_{-20} \text{ km}\) thick elastic lithosphere and overlying an upper mantle half-space with a viscosity of \(4 \times 10^{20} \text{ Pa s}.\) This earth model achieves a low degree of misfit with the observations (reduced chi-square value of \(\chi^2 = 2.5\)), suggesting that glacial isostatic rebound associated with post-Little Ice Age melting can entirely account for the rapid uplift of southern Alaska over the last \(\sim 250\) yr.

Key words: Alaskan tectonics, asthenosphere viscosity, Glacier Bay, postglacial rebound, sea level change, uplift.

INTRODUCTION

Icefields and glaciers in the coastal mountains of southern Alaska and Canada have undergone rapid thinning over the last 100–200 yr (Fig. 1) (Goodwin 1988; Clague & Evans 1993; Motyka & Beget 1996; Wiles et al. 1999; Arendt et al. 2002). Associated unloading of the Earth’s surface has led to isostatic rebound in the region (Hicks & Shofnos 1965; Clark 1977; Sauber et al. 2000). We previously used observed and historical ice load changes as input to viscoelastic rebound models to show that rapid uplift rates in southern Alaska, measured by permanent tide gauges, could be entirely attributed to viscoelastic rebound (Larsen et al. 2003). In this study, we present uplift rates from Global Positioning System (GPS) measurements, sea level rates derived from temporary tide gauge observations, and total sea level change from raised shoreline studies. Each of the three data sets indicates solid earth uplift, but arises from very

© 2004 RAS
different measurements, and hence they are discussed separately. These data are then compared with uplift predictions from Maxwell viscoelastic-gravitational rebound models. The active tectonic deformation of the region is also considered as a possible source of the uplift. However, this effect is a minor contribution relative to viscoelastic rebound.

Rapid changes in sea level in southeast Alaska were noted from tide gauge observations in southeast Alaska as early as the 1920s, and in 1959–60 a region-wide survey was performed by National Ocean Service (NOS) specifically to characterize these changes. The results were summarized by Hicks & Shofnos (1965), and a regional pattern of sea level change surrounding Glacier Bay was found, with a peak rate of −40 mm yr$^{-1}$ near the mouth of the bay. To allow for the possibility of changes in the pattern and magnitude of the regional uplift, the sea level rates that we examine here all post-date the results of Hicks & Shofnos (1965). The GPS data presented here are from a network of 45 sites, mostly measured between 1998 and 2002, and provide a broader spatial distribution of measurements than the tide gauge data, as well as direct measurements of vertical crustal velocities uncontaminated by local geoid change and global sea surface rise. The raised shoreline data presented here are an extension of the method described by Motyka (2003), and constrain both the timing and total magnitude of the ongoing uplift. When the sea level observations from tide gauges and raised shorelines are considered together with the GPS data, the total combined data set provides an exceptional record of the regional uplift.

Each data set is examined separately within the context of a single two-layer viscoelastic earth model subjected to a model of ice load changes that are independently constrained by airborne laser altimetry (Arendt et al. 2002) and records of recent glacial change (Goodwin 1988; Clague & Evans 1993; Motyka & Beget 1996; Wiles et al. 1999). The Little Ice Age (LIA) glaciation was the largest Holocene glacial expansion in Alaska, beginning around 1200 AD and reaching its greatest extent at ∼1900 AD (Wiles et al. 1999; Calkin et al. 2001). The regional isostatic uplift of southern Alaska is on a scale large enough to allow determinations of upper mantle (asthenosphere) viscosity and lithosphere elastic thickness. The ultimate goal of the present study is to test various earth models against all of the uplift observations. Specifically, we restrict this effort to a single load model that is built upon observations of glacial change, rather than iteratively constrained by the uplift data themselves. Given the rheological simplicity of a linear Maxwell solid, the results provide robust constraints of lithospheric and asthenospheric structure, as well as the statistically significant conclusion that the regional uplift is dominated by isostatic rebound associated with the post-LIA deglaciation of southern Alaska.

**TIDE GAUGE DATA AND ERROR ANALYSIS**

We have augmented sea level rates found at four permanent tide gauges (Larsen et al. 2003) with temporary tide gauge observations at 18 sites throughout the northern part of southeast Alaska (Table 1 and Fig. 2). Temporary tide gauges typically record sea level over the course of one or more monthly tidal cycles, and the elevation of the gauge is then surveyed relative to a local network of benchmarks. Mean sea level at the site is calculated and referenced to the benchmarks. When this procedure is repeated some years later, sea level change can be found relative to the benchmarks. Temporary tide gauges are primarily installed to assist in the charting of waterways for navigational purposes by the NOS. Of the data presented here, all of the initial occupations and half of the repeat occupations were performed by NOS field crews. During fieldwork between 1999 and
Table 1. Tide gauge sea level rates. Occupation dates are listed for the temporary tide gauge sites. Agency refers to who performed the second occupation of the site; all initial occupations were by NOS. Duration refers to the length of time of the second occupation. Very short occupations were a result of equipment failures.

<table>
<thead>
<tr>
<th>Site</th>
<th>Longitude</th>
<th>Latitude</th>
<th>S.L. Rate (mm yr(^{-1}))</th>
<th>Occupation Dates</th>
<th>Agency</th>
<th>Duration (months)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Muir Inlet</td>
<td>−136.11</td>
<td>58.91</td>
<td>−26 ± 3</td>
<td>1959 &amp; 2001</td>
<td>UAF</td>
<td>1</td>
</tr>
<tr>
<td>Willoughby Island</td>
<td>−136.12</td>
<td>58.61</td>
<td>−23 ± 5</td>
<td>1961 &amp; 2001</td>
<td>UAF</td>
<td>0.3</td>
</tr>
<tr>
<td>Composite Island</td>
<td>−136.57</td>
<td>58.89</td>
<td>−23 ± 3</td>
<td>1959 &amp; 2000</td>
<td>UAF</td>
<td>1.5</td>
</tr>
<tr>
<td>Excursion Inlet South</td>
<td>−135.44</td>
<td>58.42</td>
<td>−17 ± 3</td>
<td>1964 &amp; 1999</td>
<td>NOS</td>
<td>2</td>
</tr>
<tr>
<td>Lituya Bay</td>
<td>−137.62</td>
<td>58.61</td>
<td>−17 ± 3</td>
<td>1959 &amp; 1999</td>
<td>UAF</td>
<td>2</td>
</tr>
<tr>
<td>Inian Cove</td>
<td>−136.33</td>
<td>58.26</td>
<td>−16 ± 4</td>
<td>1959 &amp; 1991</td>
<td>NOS</td>
<td>2</td>
</tr>
<tr>
<td>William Henry Bay</td>
<td>−135.23</td>
<td>58.71</td>
<td>−14 ± 5</td>
<td>1959 &amp; 1984</td>
<td>NOS</td>
<td>3</td>
</tr>
<tr>
<td>Swanson Inlet</td>
<td>−135.11</td>
<td>58.20</td>
<td>−16 ± 6</td>
<td>1960 &amp; 1980</td>
<td>NOS</td>
<td>2</td>
</tr>
<tr>
<td>Salt Lake Bay</td>
<td>−135.66</td>
<td>57.96</td>
<td>−14 ± 7</td>
<td>1982 &amp; 2000</td>
<td>UAF</td>
<td>2</td>
</tr>
<tr>
<td>Elfin Cove</td>
<td>−136.34</td>
<td>58.19</td>
<td>−13 ± 4</td>
<td>1959 &amp; 1992</td>
<td>NOS</td>
<td>1</td>
</tr>
<tr>
<td>Auke Bay</td>
<td>−134.65</td>
<td>58.38</td>
<td>−12 ± 3</td>
<td>1959 &amp; 2000</td>
<td>UAF</td>
<td>2</td>
</tr>
<tr>
<td>Tenakee Springs</td>
<td>−135.21</td>
<td>57.78</td>
<td>−12 ± 6</td>
<td>1960 &amp; 1980</td>
<td>NOS</td>
<td>2</td>
</tr>
<tr>
<td>Miner Island</td>
<td>−136.34</td>
<td>58.01</td>
<td>−11 ± 3</td>
<td>1959 &amp; 2000</td>
<td>UAF</td>
<td>0.7</td>
</tr>
<tr>
<td>Annex Creek</td>
<td>−134.10</td>
<td>58.32</td>
<td>−10 ± 3</td>
<td>1959 &amp; 1997</td>
<td>NOS</td>
<td>1</td>
</tr>
<tr>
<td>Takai Harbor</td>
<td>−134.01</td>
<td>58.07</td>
<td>−10 ± 3</td>
<td>1959 &amp; 1997</td>
<td>NOS</td>
<td>2</td>
</tr>
<tr>
<td>Pt. Sinbad</td>
<td>−135.65</td>
<td>57.41</td>
<td>−8 ± 5</td>
<td>1976 &amp; 2001</td>
<td>UAF</td>
<td>2</td>
</tr>
<tr>
<td>Skagway</td>
<td>−135.33</td>
<td>59.45</td>
<td>−17.1 ± 1</td>
<td>Permanent Gauge</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Yakutat</td>
<td>−139.74</td>
<td>59.55</td>
<td>−13.7 ± 1</td>
<td>Permanent Gauge</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Juneau</td>
<td>−134.41</td>
<td>58.30</td>
<td>−13.6 ± 1</td>
<td>Permanent Gauge</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sitka</td>
<td>−135.34</td>
<td>57.05</td>
<td>−3.0 ± 1</td>
<td>Permanent Gauge</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

2001, we installed temporary tide gauges at nine sites that NOS had not recently reoccupied (Table 1).

The tide gauge data used here can be divided into three subsets: permanent (continuous recordings from 1937 to 2002); NOS temporary (one to two month occupations from 1959 to 2002); and UAF temporary (one to two month occupations from 1999 to 2002). Analysis and error estimation for the permanent gauge data used here follows Larsen et al. (2003). NOS temporary gauge data reduction is performed entirely by the NOS, and the rates used here are simply derived from the published tidal benchmark elevations for each occupation. The rate determinations at sites that were re-occupied with our gauges are derived by comparing previous NOS published tidal benchmark information with new benchmark heights determined through a combination of our gauge readings and level surveys.

The tide gauges we installed digitally recorded water depths every 15 min. We conducted level surveys from the sensor zero point to the local benchmark network when the gauge was installed and again when it was removed, both to ensure stability of the instrument and to provide a survey check. Survey errors, from circuit closure, were < ±10 mm. The historic tidal benchmark information for the sites we reoccupied is given in height above mid-tide level (MTL), which is a plane midway between all high- and low-tide readings. This datum is convenient for older analogue-style instruments, whereas mean sea level (MSL) is more convenient for digital instruments. To derive MTL from our digital records, a spline-fit was used to
approximate a continuous tidal record and the estimated high and low tides were taken from this spline-fit.

To reduce the effects of seasonal and oceanographic long-period fluctuations in mean sea level, tidal datum planes found at temporary tide gauges are typically adjusted by a concurrent offset at the nearest permanent gauge. This concurrent offset is found by taking the datum (e.g. MTL) found at the permanent gauge over the same time period and differencing it with a stable, long-term datum found over a continuous record of five or more years. All of the more recent occupations by NOS and our gauges were adjusted with this concurrent offset. NOS documentation is insuf- ficient to determine whether this adjustment was performed for many of the earlier occupations. In the error analysis that follows we may slightly overestimate the errors, as we assume that none of the initial occupations was adjusted in this manner.

The error budget for temporary gauge measurements includes level surveying errors, tide gauge instrument errors, and the variability of MSL sampled over intervals of two months or less. The first two error sources are <10 mm, while the third source is of the order of 40–50 mm (Swanson 1974). To estimate the effect of MSL fluctuations over the period of temporary gauge occupations, permanent gauge records were sampled to approximate temporary gauge records, with observation periods and time spans between observations similar to those in the temporary gauge data used here. The permanent gauge records used in this analysis are raw monthly MSL from Sitka and Juneau. The Juneau record was repeatedly sampled, providing two average sea levels, each over two consecutive months, with periods of 10, 20, 30 and 40 years between the two averages. To simulate the effect of adjusting the temporary gauge with a concurrent offset at the nearest permanent gauge, the later (‘reoccupation’) two-month samples from the Juneau MSL record were adjusted by concurrent MSL offsets at Sitka. The earlier (‘initial occupation’) samples were simply raw MSL that were not adjusted by this offset.

Several distributions of sea level rates thus found at Juneau are shown in Fig. 3. We take the standard deviation of the rates found in Fig. 3 to represent the sea level rate standard error at temporary tide gauges. A curve was fitted to these errors versus time span between occupations (Fig. 4). The errors listed in Table 1 for sea level rates were assigned according to this curve and the individual period between occupations at each site. If both the initial and repeat occupations are adjusted with the concurrent MSL offset, the errors thus derived are reduced by ~20 per cent. This gives an indication of the degree to which we may be overestimating these errors based on our assumption that NOS did not perform this adjustment on the earlier occupations.

**GPS DATA AND ERROR ANALYSIS**

A contour map of GPS uplift rates from 45 sites is shown in Fig. 5. Collection and analysis of these data is similar to methods described in Freymueller *et al.* (2000), and will only be briefly discussed here. With the exception of the three continuous stations at Whitehorse (WHIT), Biorka Island (BIS1) and Gustavus (GUS2), all of the data are from campaign-style surveys, with each site typically having

---

**Figure 3.** Histograms of sea level rates. Rates are found by repeatedly (1000 times each) sampling the Juneau permanent tide gauge monthly mean sea level record. Prior to sampling, the record was partially corrected for concurrent mean sea level offsets observed at the Sitka permanent gauge (see text). Rates were calculated over record subsets that simulate a typical temporary tide gauge data set. This subset is formed by randomly selecting two consecutive months from within the record, then two more consecutive months are randomly selected from within (X ± 0.5 yr) later, where X = 10, 20, 30, 40 yr. The red curves overlying the histograms show Gaussian distributions over the range of rates found; the solid and dashed vertical lines show the mean and standard deviation, respectively. The standard deviations (σ) found from each time span are used to assign errors to the temporary tide gauge sea level rates (see Fig. 4).
Rapid uplift of southern Alaska

Figure 4. Sea level rate errors versus time. The x-axis is the time between occupations at temporary tide gauges. The points plotted as solid circles were taken from the analysis in Fig. 3, to which a logarithmic regression was fitted (dashed line). The errors assigned to the sea level rates in Table 1, shown by open circles, were found according to this regression and the time span between occupations.

2–3 occupations over 3–4 years. The GPS data were analysed using the GIPSY software with simultaneous data from global International GPS Services (IGS) stations (e.g. Freymueller et al. 2000). The daily free network solutions were transformed into the International Terrestrial Reference Frame, epoch 1997 (ITRF97). These daily solutions were used to estimate station velocities that were transformed into a North America fixed reference frame based on the REVEL model (Sella et al. 2002). The overall average 1σ vertical velocity error of our 45 sites is 5 mm yr$^{-1}$. The network of sites in Glacier Bay is more recent (ca. 1998), and certain sites have vertical velocities with 1σ errors as large as 6 mm yr$^{-1}$. The individual site formal 1σ errors were increased by a common factor of 1.5 to account for flicker noise inherent to vertical GPS measurements (Mao et al. 1999). Site-specific rates and errors are found in Table 2.

RAISED SHORELINE DATA AND ERROR ANALYSIS

The coastal areas of southeast Alaska clearly show the effects of recent land emergence, including newly raised shoals, raised shorelines, and wave-cut benches. Our observations at a number of coastal areas surrounding (and outside) Glacier Bay found coastal forests systematically increasing in age away from the present shoreline. These first-generation forests, consisting primarily of Sitka spruce (Picea sitchensis (Bong.) Carr) and Sitka alder (Alnus viridis (Vill.) Lam. & DC. ssp. sinuata (Regel) A. & D. Love) commonly end at a raised shoreline and a riser eroded from soft sediments, above which stands a terrace of old-growth forest dominated by western hemlock (Tsuga heterophylla (Raf.) Sarg.) (Motyka 2003). Raised shorelines were identified by the following criteria: a riser eroded into soft sediments, an abrupt change in the thickness of surficial organic-rich soil, termination of subsoil beach deposits contiguous with modern beach, and an abrupt change in nature and age of coastal forest. The combination of well-defined raised shorelines and robust stands of Sitka spruce colonizing the uplifted land allowed us to estimate both the onset date of the current episode of uplift and the total amount of emergence. Sites chosen for investigation had low-energy shorelines such as in protected bays, with moderate grades (5° to 7° slope).

Evidence from experiments and fieldwork indicates that moderately cohesive sediments, such as underlie most of our study sites, erode with relative ease (Shih et al. 1994). Furthermore, erosion...
of such sediments in low-energy environments occurs only at tides above the mean high spring tide (MHST) with some minimum wind speed to produce waves up to the riser base, and such episodes of erosion are relatively infrequent (<20 events yr⁻¹) (McGreal 1979). It is therefore reasonable to assume that the base of the riser represents a shoreline constructed at the level of the highest tide stages. To determine the total emergence since the formation of the riser, we need to determine the current equivalent shoreline stand. Although Sitka spruce is considered to be saltwater-intolerant, our field observations show that shoreline stands can withstand the brief inundations of saltwater that occur during highest tides. The elevations of the lowest spruce at individual study sites were found to be comparable with recorded or estimated highest tides in the individual districts. We make the assumption that these lowest beach-fringing spruce trees mark a shoreline stand equivalent to that which formed the riser, and that the total emergence is the difference in elevation between the lowest spruce and the base of the riser.

The elevation of the raised shoreline was measured relative to the current shoreline by standard level-line surveys, which were typically accurate to <1 cm. However, the root-mean-square scatter of a raised shoreline elevation measured repeatedly at the same site is ~20 cm, owing to a transitional zone at the uppermost marine extent. A correction for soil that has accumulated over the base of the riser since its erosion is also required, and natural variations in this soil thickness contribute additional uncertainties, estimated at ±5 cm. Uncertainties in establishing the elevation of the beach-fringing spruce are similar to those for the base of the riser. The total emergence and uncertainty are given in Table 3 for the 14 sites; site locations and total emergence are shown in Fig. 6.

Dendrochronology was used to determine the onset of emergence by dating Sitka spruce trees rooted at the base of the riser. Standard increment borers were used to extract cores, which were then dated according to the number of tree rings following standard procedures (Stokes & Smiley 1968). In addition to dating errors associated with these procedures, an uncertainty in establishing the onset of emergence lies in the estimation of ecesis, i.e. the time lag between emergence and spruce germination. Estimates of regional ecesis come from studies at retreating glaciers. At Mendenhall Glacier, where Sitka spruce is within a few kilometres or less of the terminus, ecesis averaged 5 ± 3 yr for most of the 20th century (Lawrence 1950; Lacher 1999). In contrast, ecesis in Glacier Bay averaged 15 ± 5 yr at sites that were 23 to 27 km from cone-bearing spruce (Fastie 1995). Thus, proximity of cone-bearing spruce appears to be a primary factor in reducing ecesis, although other factors may also be involved. Cone-bearing spruce trees are immediately adjacent to all our sites; thus we would expect ecesis to be fairly brief. This conclusion is supported by the fact that the recent rate of downward spruce colonization at Juneau and at our other coastal study sites has closely tracked the rate of sea level fall as documented by tide gauges (Motyka 2003; Motyka, unpublished data). We therefore estimate ecesis at 5 ± 3 yr for this study. The estimated ‘onset of emergence’ at our sites from dendrochronology is given in Table 3 along with uncertainties.

### RESULTS

The pattern of sea level changes found at the tide gauge sites indicates that the fastest sea level changes in southeast Alaska are found in Glacier Bay (Fig. 2). This finding is in general agreement with the results of Hicks & Shofnos (1965), although we find peak sea level rates in upper Glacier Bay rather than at Bartlett Cove near the mouth of the bay. During fieldwork in the summer of 2001, we installed a temporary tide gauge at Bartlett Cove to repeat prior NOS occupations of this site. When reducing the data for this occupation and comparing them with NOS occupations from 1937, 1959 and 1964, we found that none of the tidal benchmarks in the local reference network could be considered stable. All of these benchmarks are on boulders 3–5 m³ in size, resting on unconsolidated marine sediments and moraine deposits. Level surveying records indicate relative motion of two or more benchmarks between each historic occupation, and by the time of our survey the entire network had been deformed. As such, all data from this site should be considered suspect, and we rejected it from our analysis. The pattern of sea level rates found over the remaining sites agrees well with the pattern of uplift rates from GPS measurements within the Glacier Bay region (Fig. 5).
Hicks & Shofnos (1965) found a sea level rate of $-35$ mm yr$^{-1}$ at Muir Inlet between occupations in 1940 and 1959. Based on the time between occupations, we assign an error of $\pm 7$ mm yr$^{-1}$ to this rate, and as such it is not significantly different from the lower rate we find for this site ($-35 \pm 7$ mm yr$^{-1}$ versus $-26 \pm 3$ mm yr$^{-1}$). Overall, the newer sea level rates presented here are consistent with those found earlier (Hicks & Shofnos 1965) when the associated errors are considered. We conclude that both the pattern and magnitude of regional sea level rates have remained essentially constant at the level of measurement accuracy since the time of the earliest rate measurements. This finding is in agreement with the linear sea level rates found over the entire permanent gauge records at Sitka, Juneau and Skagway (Larsen et al. 2003).

The GPS data, not being limited to the coastline, provide a broader spatial description of the uplift pattern (Fig. 5). In addition to peak uplift rates of the order of 25 mm yr$^{-1}$ found in upper Glacier Bay, the GPS data identify an additional area of peak bedrock uplift to the east of Yakutat sustaining rates of up to 34 mm yr$^{-1}$.

The total sea level change found at the raised shoreline sites also describes a regional pattern surrounding Glacier Bay (Fig. 6). Quite notably, the greatest sea level change occurs at the sites closest to where the peak uplift and sea level rates are found. Dates for the initiation of emergence listed in Table 3 range from 1747 AD to 1818 AD, with an average age of 1782 AD. At the best-studied site, 9-mile (Juneau), Motyka (2003) found that the onset of uplift started between 1770 and 1790. However, based on the three sites that give the earliest and nearly synchronous dates, we hypothesize that emergence began at $\sim 1750$ AD. It is possible that the difference in values at the other sites is due to either that the oldest trees eluded sampling and/or that ecesis was considerably longer than our estimate. What is clear is that emergence must have begun within a 25–40 yr period during the mid- to late-18th century, the same time period as when the collapse of the Glacier Bay Icefield had just begun and was rapidly accelerating, based on observations by Capt. George Vancouver’s survey parties (Vancouver 1798) and subsequent historical observations. Between $\sim 1750$ and $\sim 1950$ AD, the glaciers draining this icefield retreated by more than 100 km. The simultaneous onset of unloading and sea level change is a direct observation of the causal relationship between glacial unloading and the region’s uplift.

### Table 3. Raised shoreline total sea level change. The dates listed for the start of emergence represent a minimum age of the onset of sea level change (see text).

<table>
<thead>
<tr>
<th>Site</th>
<th>Longitude</th>
<th>Latitude</th>
<th>Total Sea Level Change (m)</th>
<th>Start of emergence</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sullivan Cove</td>
<td>-135.36</td>
<td>58.92</td>
<td>$-5.7 \pm 0.3$</td>
<td>1747</td>
</tr>
<tr>
<td>Pleasant Island</td>
<td>-135.70</td>
<td>58.38</td>
<td>$-4.9 \pm 0.3$</td>
<td>1810</td>
</tr>
<tr>
<td>Horse Fly Cove</td>
<td>-136.41</td>
<td>58.37</td>
<td>$-4.2 \pm 0.5$</td>
<td>1766</td>
</tr>
<tr>
<td>Excursion Inlet</td>
<td>-135.47</td>
<td>58.46</td>
<td>$-4.2 \pm 0.5$</td>
<td></td>
</tr>
<tr>
<td>Graves Harbor</td>
<td>-136.69</td>
<td>58.27</td>
<td>$-4.0 \pm 0.3$</td>
<td>1747</td>
</tr>
<tr>
<td>Goose Island</td>
<td>-136.04</td>
<td>58.21</td>
<td>$-3.7 \pm 0.3$</td>
<td>1779</td>
</tr>
<tr>
<td>Boat Harbor</td>
<td>-135.16</td>
<td>58.65</td>
<td>$-3.6 \pm 0.3$</td>
<td>1750</td>
</tr>
<tr>
<td>Inian Cove</td>
<td>-136.33</td>
<td>58.26</td>
<td>$-3.5 \pm 0.5$</td>
<td>1818</td>
</tr>
<tr>
<td>Echo Cove</td>
<td>-134.96</td>
<td>58.67</td>
<td>$-3.3 \pm 0.5$</td>
<td>1810</td>
</tr>
<tr>
<td>9-mile, Juneau</td>
<td>-134.57</td>
<td>58.36</td>
<td>$-3.1 \pm 0.2$</td>
<td>1793</td>
</tr>
<tr>
<td>Bear Creek</td>
<td>-136.30</td>
<td>58.01</td>
<td>$-2.8 \pm 0.5$</td>
<td>1768</td>
</tr>
<tr>
<td>Lisianski Strait</td>
<td>-136.37</td>
<td>57.93</td>
<td>$-2.6 \pm 0.5$</td>
<td>1786</td>
</tr>
<tr>
<td>Kadashan Bay</td>
<td>-135.23</td>
<td>57.73</td>
<td>$-1.8 \pm 0.3$</td>
<td>1807</td>
</tr>
<tr>
<td>Sullivan Bay</td>
<td>-135.65</td>
<td>57.41</td>
<td>$-0.9 \pm 0.3$</td>
<td></td>
</tr>
</tbody>
</table>

**Figure 6.** Total sea level change found at raised shoreline sites. Sea level change is indicated both in metres and by the height of the bar plotted at each site.

**Figure 7.** An overhead view of the Kodiak region showing the GPS network. Points closer than 50 km are indicated. Black circles are the GPS sites and the solid earth uplift during the mid- to late 18th century, which is controlled by deglaciation and ecesis.

### Sea Level Rates Versus Uplift Rates

Sea level rates, such as those found at the temporary tide gauges here, result from sea surface (geoid) change as well as from solid earth uplift. Specifically, the rate of sea surface change, \( \dot{\Delta S} \), is that which would be observed from a tide gauge held fixed at a constant radial distance from the geocentre. At a site on the solid earth surface, it can be approximated by the difference between the tide gauge sea level rate and the solid earth uplift rate (Tamisiea et al. 2003). To find this difference at sites in southeast Alaska, we compared sea level rates with GPS uplift measurements. In general, the two types of measurements have not been co-located, so the GPS rates were interpolated to form a continuous, second-order differentiable surface that was sampled at the tide gauge locations. The results are shown in Fig. 7, and the average difference is found to be $-0.3 \pm 0.8$ mm yr$^{-1}$ ($\pm 2\sigma$), which indicates sea surface fall rather than rise as is found in global averages (Douglas 1997). Essentially the same
result is found if we limit this analysis to co-located GPS–tide gauge sites ($\Delta S = -0.4 \pm 1.0\ \text{mm yr}^{-1}$; eight sites).

Although it may be counter-intuitive that $\Delta S$ is found to fall local to melting glaciers, this result has been known for some time (e.g. Woodward 1888). While the melting adds water to the oceans globally, reduced gravitational attraction exerted by the shrinking ice mass causes a greater effect locally on sea surface rates. Tamisiea et al. (2003) have predicted $\Delta S$ at permanent tide gauge sites in southern Alaska related to the rapid melting of the region’s glaciers, based on the ice thinning data of Arendt et al. (2002). Tamisiea et al. (2003) results for Yakutat ($\Delta S \approx -3.2\ \text{mm yr}^{-1}$), Skagway ($\Delta S \approx -2.3\ \text{mm yr}^{-1}$), Juneau ($\Delta S \approx -2.2\ \text{mm yr}^{-1}$) and Sitka ($\Delta S \approx -1.5\ \text{mm yr}^{-1}$) give an average predicted $\Delta S \approx -2.3\ \text{mm yr}^{-1}$ for the region where we have both GPS and tide gauge data. Although the formal error associated with our average $\Delta S$ suggests that there is a significant difference between the measurements and Tamisiea et al.’s (2003) results, it needs to be stressed that the measurement accuracy required for this comparison is quite high. Furthermore, this effect is spatially variable (Mitrovica et al. 2001; Tamisiea et al. 2001, 2003), and our use of a regional average is clearly an approximation. The error bars on Fig. 7 show that the individual (site-specific) $\Delta S$ measurements are quite noisy.

In the following sections, when calculating the misfit between viscoelastic model predictions and the sea level data of Tables 1 and 2, we approximate solid earth uplift at tide gauge and raised shoreline sites by assuming a constant $\Delta S = -0.3\ \text{mm yr}^{-1}$. This adjustment allows both the tide gauge data and raised shoreline data sets to be modelled in an equivalent manner to the GPS data, and facilitates a direct comparison between each data set’s model predictions.

**ISOSTATIC REBOUND MODELLING**

**Earth model and ice load model**

The isostatic modelling technique used here is that of Ivins & James (1999). It assumes a gravitating, density-stratified, incompressible two-layer earth model consisting of an elastic lithosphere and Maxwellian viscoelastic mantle half-space. This flat earth approximation is sufficient to evaluate the effects of regional ice load variations over 20° of the Earth’s surface (Amelung & Wolf 1994; Wu & Johnston 1998; Ivins & James 1999), and thus is appropriate for uplift modelling over the spatial scale of uplift observations in southern Alaska. This earth model is subjected to a composite ice load model built upon published glacial histories both specific to Glacier Bay and of the entire region of southern Alaska. Because Glacier Bay followed a tidewater glacier cycle, the unloading there was asynchronous with regional ice thickness changes, and the ice thickness changes were locally much greater than regional post-LIA thickness changes. Our approach is to test whether glacial rebound could cause the deformation observed while using the simplest possible earth model.

As noted earlier, the LIA glaciation was the largest Holocene glacial expansion in Alaska, and reached its greatest extent at ~1900 AD (Wiles et al. 1999; Calkin et al. 2001). To build a load model of the LIA, several simplifying assumptions were made. The total ice coverage of southern Alaska is composed of thousands of individual systems of partially interconnected valley glaciers and icefields, and some degree of generalization will be involved in any modelling effort. The regional ice load history shown in Fig. 8(a) is an estimate of the change in ice volume through the LIA (differential ice volume), and assumes that all of the ice accumulated through the LIA has now melted. This first assumption is based on observations of broken-off stumps of trees that were glacially overrun at the onset of the LIA and are presently being exposed in situ by glacial retreat in many areas of coastal Alaska (Wiles et al. 1999; Calkin et al. 2001).

To construct the regional ice load history of the LIA (Fig. 8a), we approximated both the timing and magnitude of volume changes based on glacier length changes, which in turn are obtained from

![Figure 8](image_url)

**Figure 8.** Volume change history of the (a) regional and (b) Glacier Bay ice load models.
geomorphic studies and dating of terminal moraines (Motyka & Beget 1996; Wiles et al. 1999; Calkin et al. 2001). In doing so, we essentially assumed an area response timescale of zero in the relationship of glacial volume change to change in area. However, this area response time is only 8 yr for a temperate glacier with a high mass-exchange rate similar to the glaciers of southern Alaska (Harrison et al. 2003), and thus negligible relative to the duration of the LIA. An additional ice model simplification arises from the assumption that all glaciers advanced and retreated simultaneously through the LIA, following a regionally averaged thickness change versus elevation relationship \(dz(z)\). We adopted this simplification based on observations of generally synchronous terminus behaviour of glaciers on the Gulf of Alaska coast throughout the LIA (Wiles et al. 1999), as well as on the similarity of \(dz(z)\) curves for various glacial subregions of Alaska (Arendt et al. 2002).

The regional-load model (Fig. 8a) is based on the measured \(\sim 1955\) to present thinning rates (Arendt et al. 2002), which are extrapolated back to the end of the LIA in Alaska at 1900 AD. We calculate ice changes continuing into the future (2005), so that our model is experiencing ongoing unloading at the evaluation date (2000). Thus, using the ‘recent’ rate of 96 km\(^3\) yr\(^{-1}\) for the period 1995–2005 AD, and the ‘early’ rate of 52 km\(^3\) yr\(^{-1}\) for the period 1900−1995 AD (Arendt et al. 2002), the maximum differential ice volume for the regional ice load model is as follows:

\[
(2005−1995\text{ yr}) \times 96 \text{ km}^3 \text{ yr}^{-1} + (1995−1900\text{ yr}) \times 52 \text{ km}^3 \text{ yr}^{-1} = 5900 \text{ km}^3.
\]

Prior to 1900 AD, we assigned the region’s advance and retreat stages each a percentage of this peak differential ice volume. The LIA occurred in three phases between \(\sim 1200\) and 1900 AD (Wiles et al. 1999). The timing and relative strength of each advance, retreat and standstill stage of the LIA are from independent studies of terminal moraine positions and dendrochronology of glacially overrun trees (Motyka & Beget 1996; Wiles et al. 1999; Calkin et al. 2001). Details of an earlier regional glaciation between 300 and 900 AD are less well characterized (D. Barclay, private communication, 2002), and our model consists of a rather crude approximation of advance, standstill and retreat over this time frame (Fig. 8a). This simplification in the load model history prior to the LIA is necessary because the larger LIA advance generally overrides and erased most of the geomorphic evidence of earlier Holocene glaciations, and we omit glacial history prior to 300 AD because of this lack of evidence.

Furthermore, a detailed ice history specific to southern Alaska during the Last Glacial Maximum (LGM) is not presently available in the literature and would be beyond the scope of the present work to develop. Thus, our modelling is limited to the last \(\sim 1.7\) kyr of glacial load changes.

The regional-load changes are distributed on an equally spaced grid of 20-km diameter discs, a spatial resolution that is approximately half the crustal thickness, with 531 disc loads used to represent all ice in Alaska and adjoining Canada from 154°W to 129°W, and from 67°N to 52°N. The ice thickness change assigned to each disc is calculated according to a southern Alaska average \(dz(z)\) (Arendt et al. 2002), and the average elevation of the ice cover within that particular grid cell. The average thickness change versus elevation relationships for the periods 1900–1995 and 1995–2003 are from the ‘early’ and ‘recent’ periods, respectively (Arendt et al. 2002). We assume that the relationship from the early period can be scaled to describe all changes prior to 1900 throughout the load history. In an exception to the southern-Alaska-wide average \(dz(z)\), the glaciers of the Yakutat Icefield (Fig. 1) are represented with a \(dz(z)\) curve specific to those glaciers (K. Echelmeyer, unpublished laser altimetry data), because thinning rates on the Yakutat Icefield are roughly three times the regional average.

A separate load model specific to Glacier Bay, which experienced an extreme ice volume loss locally, augments the regional-load model. Glacier Bay began a rapid tidewater retreat between 1750 and 1790 AD that was mostly completed by 1950 AD (Fig. 8b). Lateral moraines and trimlines (geomorphic markers formed by glacial erosion along valley walls) provide indications of post-1750 AD ice thickness changes within Glacier Bay (Field 1947; Clague & Evans 1993). Using trimlines associated with the LIA maximum in the upper west arm of Glacier Bay (Clague & Evans 1993) as a guide, we identified additional trimlines and lateral moraines throughout Glacier Bay using a combination of vertical air photo analysis and field investigations. With these points representing the LIA maximum ice extent and extending ice coverage out to the 1750 AD terminus position, we used present-day tidewater glacier analogues to estimate ice elevation profiles within Glacier Bay. Comparing these ice profiles with a digital elevation model of present-day topography and ice cover generates a map of ice thickness change (Fig. 9), and we find that Glacier Bay lost a minimum of 2500 km\(^3\) of ice from 1750 to \(\sim 1950\) AD. This load model was assigned a history based on Goodwin (1988) as shown in Fig. 8(b), and is distributed over five discs (Larsen et al. 2003). Throughout the modelling presented here the density of ice is constant at 917 kg m\(^{-3}\).

Model results: tide gauge and GPS data

The modelling approach used here for the tide gauge and GPS data is identical to that used in Larsen et al. (2003). Holding the load model fixed, we varied the lithospheric thickness and mantle viscosity of the two-layer, half-space earth model. The misfit of the rate predictions was evaluated using the chi-square merit function (Press et al. 1992). The results are shown in Figs 10 and 11. The best-fitting models are relatively insensitive to lithospheric thickness but are critically sensitive to mantle viscosity. The misfit becomes extremely large outside a narrow range of viscosities, and the 95 per cent confidence region spans less than \(2 \times 10^{19}\) Pa s (Figs 10b and 11b). Overall, the tide gauge data set requires a best-fitting model with a slightly lower viscosity than the GPS data (\(5.0 \times 10^{19}\) versus \(5.5 \times 10^{19}\) Pa s), but the best-fitting models are not significantly different at any reasonable confidence level.

Model results: raised shoreline data

For the purposes of modelling the raised shoreline data, we made the simplifying assumption that the uplift at all sites started simultaneously at 1750. To model the total uplift since 1750, we calculated surface deformation at both 1750 and 2000, and then found the difference between the two. Again, the ice load model used was held fixed and the misfit was evaluated over a range of lithospheric thicknesses and viscosities (Fig. 12). This data set requires a significantly lower viscosity than either the GPS or tide gauge uplift-rate measurements (note that Figs 12b and c are drawn over a lower range of viscosities than Figs 10b, c, 11b, c). The confidence regions placed on viscosity by the raised shoreline data do not overlap at any meaningful value with those of the GPS and tide gauge rate data.

Because total uplift predictions are driven more by cumulative ice thickness changes than by rate of thickness change, the raised shoreline predictions are most sensitive to the largest thickness changes in the ice load model. The Glacier Bay load model has much greater
Figure 9. Ice thickness change in Glacier Bay since 1750 AD. Total volume of the ice loss is 2500 km$^3$. Trimlines, lateral moraines, and terminal moraines used to define the Little Ice Age maximum ice surface are shown with black circles. The present-day shoreline is indicated by the black outline.

Figure 10. Misfit of the tide gauge uplift rates. Misfit is shown as a function of earth model parameters, evaluated using the reduced chi-square merit function; the reduced chi-square is defined as $\chi^2_r = \chi^2 / \nu$, where $\nu$ = degrees of freedom (in this case, 20). (a) Misfit versus lithospheric elastic thickness for the best-fit mantle viscosity. (b) Misfit, shown by the colour scale, over the range of earth model parameters. The best-fitting earth model is shown by the red cross, and has $\chi^2_r = 1.3$. The 95 per cent confidence region is shown by the red contour line. (c) Misfit versus mantle viscosity for the best-fit lithospheric thickness.

Figure 11. Misfit of the GPS data as a function of earth model parameters. Misfit is evaluated using the reduced chi-square merit function ($\nu = 43$). (a) Misfit versus lithospheric elastic thickness for the best-fit mantle viscosity. (b) Misfit, shown by the colour scale, over the range of earth model parameters. The best-fitting earth model is shown by the red cross. The 95 per cent confidence region is shown by the red contour line. (c) Misfit versus mantle viscosity for the best-fit lithospheric thickness.
thickness changes than the regional-load model, and the raised shore predictions were found to be predominantly related to the Glacier Bay load model. Indeed, the raised shoreline predictions are reduced by less than 15 per cent if the regional-load model is omitted altogether and only Glacier Bay is considered. Sensitivity tests in which the Glacier Bay load model was varied in magnitude by ±25 per cent changed the best-fit viscosity by only ±0.1 × 10^{19} \text{ Pa s}. This result illustrates that an in-phase viscous response to a loading cycle on a centennial timescale requires a decay time (and therefore viscosity) determined largely by the relative timing of the load history and response (Ivins & James 1999; Ivins et al. 2002).

**Model comparison**

Because the model attains values of $\chi^2 > 2$ for each data set (see left and right panels of Figs 10 to 12), we conclude that overall the model assumed can explain each data set individually, despite the significant difference in required parameter values. All three data sets are consistent with a range of lithospheric thicknesses between ~30 and 60 km, although the uplift-rate data sets (tide gauge and GPS) are more permissive of higher values as well as of a broader range of thickness. The viscosity required by the uplift-rate measurements disagrees with that required by the total uplift measurements. As noted above, model predictions of total uplift are almost exclusively sensitive to the large ice thickness changes in the Glacier Bay load model. In contrast, sensitivity tests in which we varied the magnitude of the load models show that model predictions of uplift rates are sensitive to both the regional and Glacier Bay load models (Larsen et al. 2003).

**LOW-VISCOSITY ASTHENOSPHERE**

The modelling technique employed above assumes a simple, two-layer structure with an elastic lithosphere overlying a viscoelastic mantle half-space, and as such the predictions for mantle viscosity represent an effective viscosity over the range of mantle depths in which the majority of flow is induced. Clearly, no single two-layer model can explain all of the uplift observations. We explore here the possibility of a low-viscosity asthenospheric layer beneath southeast Alaska, using the differences in sensitivity of the uplift-rate predictions and the total uplift predictions to the two load models. The influence of a low-viscosity asthenospheric layer has been noted in other regional-scale isostasy studies (e.g. Sigmundsson 1991; Bills et al. 1994; Kaufmann & Amelung 2000). We can construct an approximate three-layer model that satisfies all data acceptably well.

**Figure 12.** Misfit of total uplift at raised shoreline sites. Misfit is shown as a function of earth model parameters, evaluated using $\chi^2$ (in this case degrees of freedom $\nu = 12$). (a) Misfit versus lithospheric elastic thickness for the best-fit mantle viscosity. (b) Misfit, shown by the colour scale, over the range of earth model parameters. The best-fitting earth model is shown by the red cross, and has $\chi^2 = 1.8$. The 95 per cent confidence region is shown by the red contour line. (c) Misfit versus mantle viscosity for the best-fit lithospheric thickness.

**Figure 13.** Spatial power spectrum of (a) the regional-load model and (b) the Glacier Bay load model. Units of the y-axis (power) are normalized to allow spectrum comparison. The peak power of the regional-load model is found at harmonic load wavelength $\lambda = 730$ km, and the peak power of the Glacier Bay load model is found at $\lambda = 180$ km.

A more thorough exploration of depth-dependent viscosity will be the subject of a future paper.

The essential concept of the following analysis is that the Glacier Bay load induces mantle flow over a shallower range of mantle depths than does the regional load, due to differences in the spatial extent of the two loads. Spatial filtering of the ice load models indicates peak harmonic load wavelengths ($\lambda$) of 730 km for the regional model and 180 km for the Glacier Bay model (Fig. 13), and therefore 80 per cent of the mantle flow induced by these load models occurs above mantle depths of ~350 km and ~85 km, respectively (Cathles 1975). These mantle depths are relative to the base of the lithosphere. We will assume that the half-space viscosity required by two-layer models of the raised shoreline data is primarily a constraint on the asthenospheric viscosity beneath southeast Alaska; this assumption is correct only if the asthenospheric thickness is greater than ~85 km (Cathles 1975).

To approximate a three-layer model consisting of an elastic lithosphere, a low-viscosity asthenosphere and a viscous mantle half-space, we can consider the two-layer-model viscosity required by the uplift rates to be an effective viscosity. This effective viscosity is a function of the thickness and viscosity of the asthenosphere, as well as of the viscosity of the underlying mantle. The relationship
outlined below allows the misfit associated with the effective viscosities of the two-layer models to be mapped to the misfit related to the thickness and viscosity of the asthenosphere, and allows for an assessment of three-layer models by the combined data set of all GPS, tide gauge and raised shoreline measurements.

The presence of a low-viscosity layer reduces the characteristic relaxation time of rebound, such that half-space models predict an effective relaxation time \( \tau_{\text{HS}} = \tau_{\text{effective}} \). Cathles (1975) showed that the effective relaxation time is related to the upper mantle relaxation time through a factor \( \Re \) that is a function of load wavenumber \( \kappa = 2\pi / \lambda \), asthenospheric thickness \( (D) \), and the ratio of asthenospheric viscosity to upper mantle viscosity \( (\eta_A / \eta_{\text{UM}}) \):

\[
\tau_{\text{HS}} = \tau_{\text{UM}} \Re \left( \kappa, D, \frac{\eta_A}{\eta_{\text{UM}}} \right).
\]

The function \( \Re \) is given by Cathles (1975, eq. III-21):

\[
\Re = \frac{2C'\bar{\eta} + (1 - \bar{\eta})\kappa^2 D^2 + (\bar{\eta} S^2 + C^2) - (\bar{\eta} - 1)S' C' + \kappa D(\bar{\eta} - 1) + (S^2 + C^2)}{\eta C' - \eta S'},
\]

where \( S' = \sinh(\kappa D) \), \( C' = \cosh(\kappa D) \), and \( \bar{\eta} = \eta_A / \eta_{\text{UM}} \).

Noting that the relaxation time is linearly proportional to viscosity, \( \tau = 2\eta / \rho g \), we can use the factor \( \Re \) to calculate the effect of a low-viscosity asthenosphere on our results for the half-space mantle viscosity:

\[
\eta_{\text{HS}} = \eta_{\text{UM}} \Re \left( \kappa, D, \frac{\eta_A}{\eta_{\text{UM}}} \right).
\]

In the following analysis, we have assumed an upper mantle (half-space) viscosity in the range of \( \eta_{\text{UM}} = (2-5) \times 10^{20} \, \text{Pa s} \), in accord with glacial isostatic adjustment (GIA) estimates (Mitrovica & Forte 1997; Lambeck et al. 1998), and a load wavenumber \( \lambda \) equivalent to \( \lambda = 730 \, \text{km} \) representing the peak of the regional-load model spatial power spectrum (Fig. 13).

**Combined Model**

In the combined model, a total of 77 measurements (43 GPS, 20 tide gauge and 14 raised shoreline data) are used to constrain three earth model parameters: lithospheric elastic thickness, asthenospheric thickness, and asthenospheric viscosity. The upper mantle in this model is represented by a viscoealstic half-space with \( \eta_{\text{UM}} = (2-5) \times 10^{20} \, \text{Pa s} \); the model results we present are for a fixed value of \( \eta_{\text{UM}} = 4 \times 10^{20} \, \text{Pa s} \). Again, we performed a grid search over a range of reasonable values and calculated the misfit for each combination of parameters. When calculating the misfit, all data were given equal weighting, subject to their errors, regardless of measurement type (GPS, tide gauge or raised shoreline).

The best-fitting model is found to have a reduced chi-square value of \( \chi^2 = 2.5 \) at asthenospheric viscosity \( \eta_A = 1.4 \times 10^{19} \, \text{Pa s} \), asthenospheric thickness \( D = 110 \, \text{km} \), and lithospheric elastic thickness \( = 50 \, \text{km} \). The residuals to this model are shown in Fig. 14. To display the 3-D \( \chi^2 \) distribution, Fig. 15 shows contour plots drawn on the three orthogonal planes passing through this global minimum. These distributions are strongly non-Gaussian, and the minima have broad ‘floors’ and steep ‘sides’ so that the 99.73 per cent confidence regions are only slightly larger than the 68.3 per cent confidence regions. Sensitivity tests in which the value assumed for \( \eta_{\text{UM}} \) was varied between \( (2-5) \times 10^{20} \, \text{Pa s} \) found that these changes had little overall effect on either the model parameters or this best-fitting value of \( \chi^2 \), with the exception that slightly thinner asthenospheric thicknesses \( (\sim 100 \, \text{km}) \) were required when lower values of \( \eta_{\text{UM}} \) were assumed.

**Figure 14.** Residuals from the best-fitting three-layer model. This model has asthenospheric viscosity \( \eta_A = 1.4 \times 10^{19} \, \text{Pa s} \), asthenospheric thickness \( D = 110 \, \text{km} \), and lithospheric elastic thickness \( = 50 \, \text{km} \). The upper panel shows contours of uplift-rate residuals found over the combined GPS and tide gauge data sets (2 mm yr\(^{-1}\) contour interval). Tide gauge sites are shown by open blue circles and GPS sites by red diamonds. The lower panel shows the total uplift residuals found at the raised shoreline sites, indicated both in metres and by the height of the bar plotted at each site. Blue bars indicate that observations are greater than model predictions, and red bars indicate that observations are less than model predictions.

The parameter constraints shown in Fig. 15 allow a range of lithospheric thicknesses similar to the individual two-layer model constraints shown in Figs 10–12. The asthenospheric viscosity prediction of the best-fitting three-layer model is the same as for the best-fitting two-layer model of the total uplift. However, the three-layer model allows for a wider range of asthenospheric viscosity, as there is some trade-off between asthenospheric thickness and viscosity permitted by the misfit distribution. As one might intuitively expect, the data require a thinner asthenosphere if the viscosity of this layer is lower than the best-fit value, and a thicker asthenosphere if it is higher. The best-fit values of the three-layer model for asthenospheric thickness and viscosity correspond to a two-layer mantle half-space effective viscosity of \( 5.6 \times 10^{19} \, \text{Pa s} \), essentially
Figure 15. Contour maps of $\Delta \chi^2$ confidence regions for three-layer earth models that satisfy the combined tide gauge, GPS and raised shoreline data set of 77 uplift observations. The plots are on three slices of the $\Delta \chi^2$ volume distribution through the global minimum located at asthenospheric viscosity $\eta_A = 1.4 \times 10^{19}$ Pa s, asthenospheric thickness $D = 110$ km, and lithospheric elastic thickness = 50 km. Each plot has three contours that correspond to 68.3, 95.4 and 99.73 per cent confidence regions of $\Delta \chi^2$. The top panel shows the misfit as a function of asthenospheric viscosity and thickness at a constant value of lithospheric thickness = 50 km. The middle panel shows misfit as a function of asthenospheric viscosity and lithospheric thickness at a constant value of asthenospheric thickness = 110 km. The bottom panel shows misfit as a function of lithospheric thickness and asthenospheric thickness at a constant value of asthenospheric viscosity = $1.4 \times 10^{19}$ Pa s.

Rapid uplift of southern Alaska

equivalent to the best-fit value required by the uplift-rate data in a two-layer model. That is, the inclusion of the total uplift measurements in the combined model does not force the uplift-rate viscosity predictions away from their preferred minimum misfit value.

TECHNICAL AND GIA UPLIFT CONTRIBUTIONS

The long-term effect of tectonically driven uplift is dramatically evident in the high mountains of the northeastern portion of our study area, particularly along the northern and northeastern boundaries of the Yakutat Block. The Yakutat Block is a microplate, travelling at nearly full Pacific Plate velocity (Fletcher & Freymueller 2003), and actively colliding with North America (Plafker et al. 1994). The resultant Fairweather Range and St Elias Mountains exhibit large relief, steep topography and peak elevations of 5–6 km. Bedrock exhumation rates from apatite fission-track data have been estimated at $\sim 1.5$ mm yr$^{-1}$ for the St Elias Mountains to the northwest of our study area (O’Sullivan et al. 1997), and 0.2 mm yr$^{-1}$ within the Coast Mountains to the southeast (Farley et al. 2001). Tectonic uplift rates in these mountains have also been inferred by a number of studies, and range between 1 and 10 mm yr$^{-1}$. The fastest of these inferred uplift rates come from localized areas within the St Elias Mountains. Based on sedimentation rates over the last 10 kyr, Sheaf et al. (2003) estimate uplift rates averaged over both the St Elias Mountains and Fairweather Range to be $\sim 5$ mm yr$^{-1}$. Meigs & Sauber (2000) suggest locally higher uplift rates, $\sim 10$ mm yr$^{-1}$, along major thrust faults within the St Elias Mountains. Finite-element modelling of Alaska-wide neotectonics suggests crustal thickening rates of 7 mm yr$^{-1}$ at the leading edge of the Yakutat Block collision (i.e. the St Elias Mountains), with resultant uplift rates of the order of 1 mm yr$^{-1}$ (Bird 1996).

In the rebound modelling presented above, we chose not to account for any tectonic contribution to the regional uplift. We based this decision on uncertainties in both the magnitude and distribution of these rates, as well as on the fact that the highest tectonic uplift rates are expected to be within the St Elias Mountains. The majority of our measurements are in a tectonic setting that is fundamentally different from the thrust faulting and crustal shortening of the leading edge of the Yakutat Block collision in the St Elias Mountains. To the southeast of Yakutat Bay, motion along the Fairweather Fault is almost entirely fault-parallel strike-slip, and the fault-normal component of relative motion between the Pacific and North American plates is accommodated by offshore structures such as the Transition Zone fault (Fletcher & Freymueller 2003). Sensitivity tests in which we assumed a constant $5.0 \pm 5.0$ mm yr$^{-1}$ tectonic contribution at all GPS and tide gauge sites within 50 km of the Fairweather Fault had little effect on the best-fitting three-layer earth model parameters.

Global-scale GIA contributions to the measured uplift rates were also not included because of uncertainties in the magnitude of this signal, stemming from limited knowledge of last glacial maximum (LGM) ice sheet history here. GIA predictions throughout our study area using the ICE3G ice load model (Tushingham & Peltier 1991) and a variety of mantle viscosity profiles are consistently $< 1$ mm yr$^{-1}$ (Bölling 2001). This negligible effect should not be particularly surprising, as ICE3G has little ice in southern Alaska. To accurately assess the impact of LGM loading in southern Alaska, a detailed LGM load history specific to the region would need to be developed, as has been done for the region just to the south of our study area in British Columbia (Clague & James 2002). Finally, GIA ocean-loading-induced relative sea level effects are also likely to be limited to less than 1 mm yr$^{-1}$ (Kendall et al. 2003).
IMPLICATIONS

Studies that constrain the rheological structure of the Earth roughly fall into the three categories of (1) regional-scale deformation studies caused by surface load changes related to transient lakes, ‘small glaciers’ (i.e. other than the polar ice sheets of Greenland and Antarctica) and mining, (2) GIA studies of deformation related to the rise and fall of Pleistocene ice sheets, and (3) studies of the transient deformation following large earthquakes (postseismic deformation). The uplift pattern and load changes in southern Alaska are large compared with many of the studies that fall in the first category (i.e. Klein et al. 1997; Kaufmann & Amelung 2000; Thoma & Wolf 2001). Lake Bonneville is perhaps the only well-studied analogue that is close both in the spatial scale and the magnitude of the load changes ($\lambda \approx 520$ km and $\Delta V \approx 9.5 \times 10^4$ km$^3$). Bills et al. (1994) modelled an extensive shoreline data set and found that the best-fitting earth models generally consist of a 40-km thick lithosphere, a 110-km thick asthenosphere with viscosity of $1 \times 10^{18}$ Pa s, and a mantle lid of 150-km thickness and $3 \times 10^{20}$ Pa s. The thicknesses of the lithosphere and asthenosphere in these models are in excellent agreement with our three-layer model results. However, their result for asthenospheric viscosity is an order of magnitude lower than ours.

Other regional-scale studies of deformation caused by surface load changes agree with asthenospheric viscosities of the order of $10^{18}$ Pa s (Kaufmann & Amelung 2000; Thoma & Wolf 2001), although the loads considered in these studies are too small to allow for a determination of asthenospheric thickness. The higher viscosity required by the data presented here may be related to differences in tectonic setting. High heat flow associated with mantle upwelling beneath the Basin and Range province (Bills et al. 1994; Kaufmann & Amelung 2000) and the mid-Atlantic ridge (Thoma & Wolf 2001) may lead to the lower asthenospheric viscosities found there. Our results for lithospheric thicknesses and asthenospheric viscosities are essentially median values in the ‘weak mechanical regime’ modelled by Ivins & James (1999), who suggest that such values are characteristic of tectonically active regions.

It is difficult to compare directly the earth model results presented here with GIA studies owing to the larger spatial scale and coarser temporal resolutions of the load models used in such studies. In general, GIA studies provide robust constraints on bulk upper and lower mantle viscosity but provide little detailed information on uppermost mantle and crustal structure (Mitrovica & Peltier 1991). We assumed an upper mantle viscosity based on GIA results. In a study of isostatic response to the Late Wisconsinan Cordilleran ice sheet, $\sim 1000$ km to the south of our study, Clague & James (2002) find that asthenospheric viscosities in the range $0.5 \times 10^{19}$ Pa s are required by relative sea level and tilted lake shoreline observations. In addition, these data require a thin elastic lithosphere ($35–60$ km thickness), in excellent agreement with our results. However, they note that the effects on uplift predictions caused by reasonable earth model variations are small compared with those brought about by plausible variations in the ice sheet history assumed.

Comparisons with earth models based on postseismic studies can also be difficult, although here the problem is one of ambiguity. Postseismic deformation may be caused by a combination of (1) aseismic slip on the main fault or neighboring ones, (2) poroelastic relaxation within the crust, and/or (3) viscoelastic relaxation of the lithosphere and asthenosphere (Bürgmann et al. 1997). With that said, our results for asthenospheric viscosity are strongly consistent with those of Pollitz et al. (2001) based on deformation following the 1992 Landers, California earthquake. Moreover, the models presented here should provide a useful starting point for future studies of viscoelastic postseismic mechanisms in southern Alaska.

CONCLUSIONS

Large sea level changes have been measured in the northern part of southeastern Alaska with tide gauges and studies of raised shorelines. The tide gauge data set depicts a regional pattern of sea level rates from $-3$ to $-26$ mm yr$^{-1}$, centred over upper Glacier Bay. We find that the overall magnitude and pattern has not significantly changed since previous similar measurements (Hicks & Shofnos 1965). GPS uplift measurements from 45 sites also indicate peak uplift rates of the order of 25 mm yr$^{-1}$ in upper Glacier Bay. A second area of peak uplift rates of up to 40 mm yr$^{-1}$ is centred over the rapidly thinning Yakutat Icefield. Raised shorelines at 14 sites show that total sea level changes in the range $-0.9$ to $-5.7$ m have occurred since $\sim 1750$ AD, with a regional pattern of higher uplift closest to upper Glacier Bay. The onset of uplift measured at the raised shoreline sites occurred at the same time that the Glacier Bay Icefield began its dramatic collapse.

When the GPS uplift rates and tide gauge uplift rates are modelled independently, the two data sets require similar constraints on the earth model parameters (lithospheric elastic thickness and mantle half-space viscosity). The uplift data from the raised shorelines require a significantly lower mantle half-space viscosity, which we propose is due to the almost exclusive sensitivity of this data to the large ice thickness changes local to Glacier Bay. When modelled together as a combined data set, the GPS, tide gauge and raised shoreline measurements require (at the 95 per cent confidence level) a three-layer earth model consisting of a $50^{+20}_{-25}$ km thick elastic lithosphere, an asthenosphere with viscosity $\eta_A = (1.4 \pm 0.3) \times 10^{19}$ Pa s and thickness $110^{+20}_{-15}$ km, overlaying a viscous upper mantle half-space ($\eta_m = 4 \times 10^{20}$ Pa s). The best-fitting combined model results in an overall misfit of $\chi^2 = 2.5$, and is thus able to account for a total of 77 measurements of uplift from three distinctly different techniques with a low overall degree of misfit. The low misfit achieved clearly suggests the presence of a low-viscosity asthenosphere beneath southern Alaska; conversely, further model complexity (i.e. additional layering) is not strongly warranted by this already low misfit and the current level of measurement errors. Throughout the modelling presented here, we have assumed load models based solely on independent observations of glacial change, rather than iteratively modifying load models to minimize this misfit. As such, our results show that (1) uplift observations in southeast Alaska can be entirely explained by post-glacial isostatic rebound, and (2) these observations seem to offer robust constraints on lithospheric elastic thickness, asthenospheric thickness and asthenospheric viscosity.

ACKNOWLEDGMENTS

The research was funded in part by the National Science Foundation’s Active Tectonics program, grant #EAR9870144. CFL was supported by a NASA Earth System Science Graduate Fellowship, and a graduate student award from the Center for Global Change and Arctic System Research. Anthony Arendt and By Valentine assisted with airborne laser altimetry data of glacial thinning which formed the basis for the ice load models. Extensive field measurements were made easier and more enjoyable with the assistance and company of Joel Johnstone, Brian Hitchcock, Sandy Zirnheld, By Valentine and Hilary Fletcher. Mike Schmidt and Paul Flick at the Pacific Geoscience Center provided GPS data from several points in Canada.
Many of the earlier GPS data and tide gauge records in southeast Alaska are the result of hard working and yet anonymous personnel of the National Geodetic Survey, the National Ocean Service, and, before them, the Coast and Geodetic Survey. We thank Steve Lyles at the National Ocean Service for assistance with retrieving archived tide gauge records. Glacier Bay National Park and the crew of the R/V Nunatak provided much-appreciated field logistical assistance. The map figures and most of the ice load model calculations were generated with the Generic Mapping Tools software (Wessel & Smith 1998; http://gmt.soest.hawaii.edu/).

REFERENCES


© 2004 RAS, GJI

A macroscopic approach to glacier dynamics, J. Glaciology, 49, 164, 13–21(9).


