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OTHER AUTHORS
Carol A. Raymond
Thomas S. James

KEY WORDS FOR INDEXING (Separate terms with commas)
GPS geodesy, glacial rebound, sea-level change, earth structure

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The Influence of 5000 year-old and Younger Glacial Mass Variability on Present-day Crustal Rebound in the Antarctic Peninsula

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Proceedings of the GPS99 & APSG Meeting in Tsukuba Oct.18-22, 1999.

Abstract. Much of the late-Quaternary ice sheet history in the northern hemisphere is now relatively well-constrained, with the total contributions to eustatic sea level change from North America and Eurasia estimated at roughly 60 ($\pm$ 12) and 20 ($\pm$ 7) m, respectively, and with deglaciation bracketed at 22 to 8.5 kyr BP. The rate of rebound at the former ice sheet centers is roughly 11 ($\pm$ 3) mm/yr. Assessment of Antarctic rebound is, however, complicated by two issues: (1) The total ice volume at Last Glacial Maximum is contentious, with estimates ranging from just a few meters to several tens of meters of equivalent eustatic sea level rise. (2) The late Holocene mass budget is also uncertain. Space-based geodesy may provide important data in the coming years for estimating the recent ice mass balance state of Antarctica. Toward this end, GPS has an important role for isolating the solid earth movements that are associated with postglacial rebound. Here we provide numerical examples of vertical motions that are predicted by coupling realistic glacial load histories to 20th century ice mass imbalance estimates for the Antarctic Peninsula. The main complexity revealed by these examples is the striking difference among predictions that have an oscillatory mass change during the last 5000 to 50 years, as opposed to those having a continuous (non-oscillatory) mass drawdown of the grounded ice sheet.

1. Introduction

Global Positioning System (GPS) satellite geodesy has emerged during the last decade as a method of determining present-day deformation of the earth's crust at precisions of 1 - 3 mm in the horizontal and to within 10 mm in the vertical (Herring, 1999). Measurable changes in tilt and gravity accompany slow viscous rebound of bedrock once buried beneath the great ice sheets of the northern hemisphere that disintegrated from their Last Glacial Maximum (LGM) some 21 to 8 thousand years ago. GPS measurement of the pattern and rate of vertical crustal motion in Fennoscandia is now unfolding at a rapid pace (Scherneck et al., 1998). Efforts are now underway to make similar measurements on Antarctic bedrock (Tregoning et al., 1999; Raymond et al., 1999).

One of the main themes of current glaciological study in Antarctica is to unravel the mass history of the great ice sheet during the past 100,000 years, as this would provide a key parameter for the study of global paleoclimatology and paleoceanography. Significant progress has been made in

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the past decade, primarily fueled by new analyses of ice core data, datable volcanic-ice deposits and new terrestrial moraine and marine sediment chronologies (e.g., Bentley and Anderson 1998; Ingólfsson et al., 1998). In terms of measuring the present-day glacioisostatic motions of the crust, a major challenge for numerical modelers is to account for the changes that occur during the past several thousand years since it is likely that the ice sheet has a well-insulated internal dynamics, with a relatively slow response to external climatological forcing (Jóhannesson et al., 1989). Far-field relative sea level data may also provide evidence for a prolonged Antarctic ice sheet evolution over the past four thousand years (Okuno and Nakada, 1998). The situation, however, may be rather complicated in coastal regions where changes in moisture flux and oceanic thermal conditions occur on both decadal and centennial time scales, thus influencing shorter time-scale ice mass budgets.

Several facts motivate a study of the glacial isostasy of the Antarctic Peninsula. First, a geodetic project directed by the Scientific Committee on Antarctic Research (SCAR) has retrieved several years of epoch campaign data in the region (Dach and Dietrich, 1999). Secondly, the geographic setting of the Peninsula renders it more vulnerable to fluctuations in climate and precipitation than elsewhere in Antarctica. Finally, like the glacial geology that dominates West Antarctica, the region has clear evidence of a more expansive ice sheet during the last 35 kyr BP and, as such, several of the issues addressed in this study also bear upon our understanding of continent-wide Antarctic glacial isostasy. Using numerical experiments we show how some contrasting, yet realistic, scenarios for mass evolution over the past 5 kyr may affect future interpretation of secular trends in vertical GPS height measurements taken over the course of half a decade, or longer, in the Antarctic Peninsula. The most striking contrasts are to be discovered among models that exhibit mass oscillations into the present millennium as these may exhibit a forced viscoelastic wave-like structure. Models having a continuous mass drawdown produce a more predictable pattern of uplift at present-day.

2. The Role of Past and Present Ice Mass Changes.
2.1 Rebound from LGM.

Our knowledge of the past and present mass balance state of the Antarctic Peninsula is, unfortunately, rather limited. There is, however, clear evidence for periodic retreat and advance of glacier systems north of 65 °S (e.g., Björck et al., 1996) and a large body of evidence now confirms that a more massive ice sheet was grounded to the continental shelf during the Last Glacial Maximum (Payne et al., 1989; Bentley and Anderson, 1998). A detailed discussion of this evidence is beyond the scope of the present paper. However, following a recent summary by Ingólfsson et al. (1998), Figure 1 shows the locations of some of the dated terrestrial and marine carbon deposits used to infer the position of the retreating LGM ice sheet margin. Here we employ an ice load retreat history modified after models of Payne et al. (1989) and Denton et al. (1991). James and Ivins (1998) constructed a surface load for the entire Antarctic ice sheet, termed the 'D91 model', in which

deglaciation from LGM was assumed to begin at 12 kyr and terminate at 5 kyr BP. The rather uncertain timing was set by far-field analyses of global paleoshorelines (Peltier, 1994). The total Antarctic contribution to postglacial eustatic sea level rise (e.s.l.r) from the previous D91 model was 24.5 m.

Figure 2 shows a map of the predicted rate of present-day rebound for the Antarctic Peninsula using a disc-load modified from the D91. The modifications follow a regional retreat history developed in the numerical simulation of Payne et al. (1989). The relatively coarse disc-gridding is shown in Figure 2. Retreat from LGM is assumed to initiate at 14.0 kyr BP and terminate at 5.5 kyr BP, largely consistent with moraine data (Clapperton and Sugden, 1988). Dates older than 6 kyr BP shown in Figure 1 correspond to the recession from LGM, with younger dates corresponding to mid to late Holocene readvances and retreats. The D91 model was also reduced in total size to an e.s.l.r. of 20 meters. For computing the result shown in Figure 2 the deglaciating discs cover the entire Antarctic continent. The predictions for present-day uplift rates are affected by the assumed value of the mantle viscosity. For example, along the southernmost coastline of Alexander Is. (see Figure 1) the predicted uplift rate with a viscosity, $\eta = 4 \times 10^{20}$ Pa s, and lithospheric thickness, $h = 70$ km, is 1.5 to 2.5 mm/yr (see Figure 2). In contrast, computational results (not shown here) for a mantle viscosity of $\eta = 10^{21}$ Pa s and thicker lithosphere, $h = 120$ km, predict a rate of 6 - 9 mm/yr at this same location. (Also see James and Ivins (1998), Figure 13a, for computational results for the unmodified D91 load, upper mantle viscosity, $\eta = 10^{21}$ Pa s and $h = 120$ km). Additionally, we computed the uplift rates for the modified D91 load with $h = 50$ km and $\eta = 1.5 \times 10^{20}$ Pa s, and in this case the predicted uplift rates reduce to the level of 0.1 mm/yr. It is important to consider reduced upper mantle viscosity and thinner lithosphere due to the late Cenozoic tectonics of the Antarctic Peninsula. In particular, the mantle environment has absorbed a series of subducting ridge segments during the past 45 to 6 Myr BP (Barker, 1982; Hole et al., 1991; Scarrow et al., 1998). However, present-day rifting is limited to crust well north of Alexander Is. (Bell and King, 1998) and, consequently, solid earth structure might not be required to be as weak, for example, as constrained recently by James et al. (2000) for southern British Columbia ($\eta < 10^{20}$ Pa s, $h < 50$ km) where Neogene arc-related tectonics occur. We should note that Studinger and Miller (1999) have recently estimated an effective elastic flexural thickness of 35 km for the lithosphere at the inner margins of the Weddell Sea. Currently, estimates of the regional upper mantle viscosity do not exist.

2.2 Rebound Caused by mid-Holocene to Present-day Mass Balance State.

Analyses of atmospheric energy and moisture transport indicate that the Antarctic Peninsula and environs is susceptible to relatively extreme precipitation conditions (Cullather et. al., 1998; Genthon and Krinner, 1998). It is estimated that the Peninsula receives roughly 25% of the total Antarctic mean annual surface accumulation, while accounting for only 6.8% of the total Antarctic ice sheet surface area (Drewry and Morris, 1992). The present mass flux of regional glacier systems into the

oceans may be substantial and non-steady, possibly as evidenced by the recent breakup of the Larsen Ice Shelf near Robertson Island along the eastern flanks of the Peninsula (Doake et al., 1998). Vaughan et al. (1999) recently determined that the Peninsula north of 75 °S sustains an accumulation input of roughly 2390 Gt/yr. Partitioning 1/3 of this input to grounded ice and taking the imbalance (capable of contributing to a secular 20th Century sea level rise) to be 6%, yields a value of −48 Gt/yr. We shall use this value for the mass imbalance of grounded ice in the Antarctic Peninsula in order to demonstrate how large present-day imbalances, and their associated isostatic disequilibrium, tradeoff with isostatic changes associated with earlier ice mass variability.

2.3 Oscillatory Load Examples.

Figure 3 shows the present-day vertical uplift rate for the Antarctic Peninsula region using the disc-load distribution indicated in mapview. This calculation assumes a purely elastic response to the load shown in Figure 4 with mass loss of -48 Gt/yr at present-day. The discs vary only in height and not in radius. Note that the maximum uplift is coincident with the disc coverage, a feature which diminishes when the viscous response is accounted for. In spite of the substantial mass loss that is assumed (equivalent to a 0.133 mm/yr contribution to secular sea level rise), this vertical response would be difficult to detect even under ideal conditions. For example, for a quasi-continuous time series having a scatter in the vertical component identical to the average of two southern hemisphere IGS stations, Perth (32 °S) and Yarragadee (29 °S), of 9.4 mm (Herring, 1999), then GPS observations at the SCAR site FOS1 (see Figure 1) would require 10 years of data in order for a linear trend to have an amplitude standout above the r.m.s. scatter in the vertical by more than 60%.

If the solid earth rheology, however, involves ductile flow then the prediction is altered substantially. For example, for a strength-depth profile similar to that of northern Europe, such that the mantle viscosity, η, is near 4×10^{20} Pa s and the lithospheric thickness, h, is 70 km (Lambeck et al., 1998), then the expected signal increases 4 to 5-fold. Assuming the same present-day mass loss as used in the purely elastic computation of Figure 3, Figure 5 shows the vertical rates for the same solid earth model as in the LGM-load case shown in Figure 2, but now with the saw-toothed load of Figure 4. Of interest in Figure 5 is the phase-lagged behavior of the solid earth response: note that the sign of the vertical motion is reversed from that of the elastic response (Figure 3). This is caused by the viscous memory of the 650-year load buildup. As these phase lags may be critical for correctly interpretating solid earth geodetic signals driven by late-Holocene ice loading-unloading sequences, we are obliged to provide an analysis of the pertinent viscoelastic-gravitational behavior. The basic physics is analogous to that of a dissipative mechanical system in forced oscillation.

2.4 Single Saw-tooth Load.

Consider a single disc of radius α and a single saw-tooth load history which would include only the final two linear segments of the history shown in Figure 4. For the example in Figure 4, such a

load would initiate at AD 1200, reach a maximum (M_{max}) at AD 1850, and then unload at a mass loss rate $\dot{M}_{pd}$ to the present-day. Analysis shows that the present-day vertical uplift rate is;

$$\dot{w} = -\frac{g}{\pi \mu_2^e \alpha} \cdot < \Gamma'(k') \left\{ \dot{M}_{pd} \left[a'_2(k') + \frac{\mu_2^e}{4 k' \mu_1^e} \right] \right. $$

$$+ v'_p(k') \left\{ \dot{M}_{pd} \left[1 - e^{-\gamma_p(k')(t - \Delta t_1)} \right] \right.$$

$$\left. \left. - \frac{M_{max}}{\Delta t_1} (1 - e^{\gamma_p(k') \Delta t_1}) e^{-\gamma_p(k') t} \right\} \right\} > , \tag{1}$$

with an implied sum over the two decay modes (p). The convolution from wavenumber (k') to radial position r' away from the disc center is

$$< \cdots > \equiv \int_0^\infty \cdots J_0(k'r') J_1(k'\alpha') dk' ,$$

with J_n the Bessel functions of order n and the prime indicating the scaling by h as discussed by Ivins and James (1999). Here the final phase (post-1850 AD, see the youngest saw-tooth portion of Figure 4) is of duration Δt_2 and the time at present-day is $t \equiv \Delta t_1 + \Delta t_2$, with Δt_1 representing the duration of the assumed single growth phase. Equation (1) is a time-derivative of expression (36) of Ivins and James (1999) for the vertical displacement at the surface of a hydrostatically pre-stressed, two-layered gravitational half-space with the deepest layer of an incompressible Maxwell rheology with elastic shear modulus μ_2^e, density ρ_2 and viscosity η. The model top layer (lithosphere) has thickness h, shear modulus μ_1^e and density of ρ_1. All of these layered earth parameters are implicitly retained in the amplitude factors $\Gamma'(k')$ and $v'_p(k')$ and the inverse decay times $\gamma_p(k')$. The explicit expressions are given in Ivins and James (1999). Equation (1) has a simple physical interpretation:

$$\dot{w} = \text{Present-day Mass Balance Rate} \times$$

(Elastic Deformation + Viscous Memory of Current Evolution) +

(Most Recently Terminated Mass Balance Rate ×

Viscous Memory of Most Recently Terminated Change) .

Note the existence of two competing terms in parentheses. If the present-day rate of surface displacement is to be "in-phase" with the present-day mass balance, then there must be sufficient time (Δt_2) to generate a viscous memory of the "current" (i.e., interdecadal) linear evolution in ice mass. Both the ability to "remember" (or "forget") the earlier phase and to establish a sufficiently robust present-day

To illustrate this fundamental difference in the vertical motion prediction between monotonic and oscillatory cases, we now explore a continuous drawdown of a late-Holocene load having the same disc structure as in Figures 3, 5 and 6. The final linear segment in each of the oscillatory cases (i.e., since AD 1850) are identical in mass change. The drawdown case also contains an identical final segment. Although the load is unrealistic in the sense that it utilizes the same 146 discs throughout a simulated 108,000 year evolution, it serves the purposes of a systematic comparison to the oscillatory load cases. The assumed total volume change over the complete 100 kyr glacial cycle is quite small; a mere 0.72 meters of e.s.l.r. and only an equivalent 0.26 meters since 11 kyr BP. The present-day response is predicted in Figures 7a-c for three different mantle viscosity values, all other parameters being identical. Note the sensitivity to mantle viscosity, with the $\eta = 4 \times 10^{19}$ Pa s case (Figure 7a) predicting more than double the uplift rates of the case of viscosity that is increased by one order of magnitude (Figure 7c). However, even in the case of Fennoscandian-like viscosity $\eta = 4 \times 10^{20}$ Pa s (Figure 7c), the relatively small deglaciation ($<$ 1 m of e.s.l.r.) predicts a surprisingly large uplift rate at the present-day. While this has been noted in previous calculations of Antarctic deglaciation, this is the first systematic study of alternative styles of late-Holocene ice mass change. Here the last 150 years of evolution for the computation of Figure 7 is identical to that assumed in Figures 3, 5 and 6. It would appear that a continuous drawdown mode of deglaciation has important implications for geodetic observational strategies on solid bedrock due to the relatively large predicted signatures. Possibly as important is the fact that the continuous drawdown response is relatively uncomplicated by the strong sensitivity to mantle viscosity, load sequencing and wavenumber-dependence that the oscillatory load cases exhibit.

3. Conclusions

In this paper we have examined three different types of ice load changes for the Antarctic Peninsula in order to predict present-day vertical rebound that could be measured using GPS. The three different load types are: (1) no evolution (constant ice mass) since 5.5 kyr BP, but having a model for LGM ice mass that is consistent with the reconstructions by Payne et al. (1989) and Denton et al. (1991); (2) oscillatory evolution since 4.0 kyr BP which includes a realistic (albeit large) mass change rate since the year AD 1850; and, finally; (3) a model having a small LGM buildup, but a continuous drawdown of ice mass with the final 150 years identical to case (2). The first case (1) is similar to the classical study of rebound wherein the time elapsed since deglaciation, size of the ice load, mantle viscosity and lithospheric thickness are the main parameters that influence the prediction of present-day uplift rate and pattern. A single mapview of one prediction is given in Figure 2. In the second case, one must add the details of the last millennial to centennial-scale oscillation(s), including the duration of loading and unloading sequences and net volume exchange with the ocean to this list. If the mantle viscosity is in the range 10^{19} to 4×10^{20} Pa s, then much of the wavenumber-dependent

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Figure Captions

Figure 1. Map of the Antarctic Peninsula. Locations of dated glacial retreat from LGM, as summarized by Ingólfsson et al. (1998), are indicated. The southernmost six SCAR GPS sites (Dach and Dietrich, 2000) and the continuously operating IGS station (O'Higgins) are shown as large solid dots and a solid diamond, respectively. The complete SCAR network is shown at the website http://www.tu-dresden.de/ipg/tpgsc98.html.

Figure 2. Uplift rate, $\dot{w}$, in mm/yr predicted from a continent-wide model of Antarctic deglaciation. The load model terminates evolution at 5.5 kyr BP. The load is modified from 'D91' of James and Ivins (1998) as discussed in the text.

Figure 3. Vertical motion due to present-day mass loss at rate $\dot{M}_{pd}$ = −48 Gt/yr with elastic rheology. Unloading occurs on all 146 circular discs. The rate is equivalent to a regional imbalance of 6%, corresponding to a 0.133 mm/yr contribution to present-day sea level rise. The assumption of a purely elastic rheology means that the crustal motion is sensitive only to $\dot{M}_{pd}$ (see Eq. 1).

Figure 4. The load history for generating Figures 3 and 5. All phases are included in the model for Figure 5, but only phase "IV" with a 4500 year-long growth period is included in Figure 6.

Figure 5. Uplift rate predicted for multiple late-Holocene oscillations, the last having a growth phase duration of 650 years (see Figure 4). The rate of mass loss for the combined 146 discs since AD 1850 is identical in Figures 3, 5, 6 and 7.

Figure 6. Uplift rate predicted for a single oscillation having growth phase duration of 4500 yr. The combination of longer growth phase and lower viscosity (versus the oscillatory case of Figure 5) allows isostatic equilibrium to be approached by 1850 AD. Post-1850 ice loss is identical to the cases shown in Fig. 2, 3 and 5.

Figure 7. Uplift rate prediction maps from non-oscillatory (continuous drawdown) deglaciation. The LGM load has a volume equivalent to 0.72 meters of e.s.l.r., with approximately 0.26 meters eustatic equivalent since 11 kyr BP. Disc positions are identical to those shown in Figure 3. Note the stronger signature (by about 3 ×) predicted by a mantle viscosity that is reduced by one order of magnitude and note the consistent uplift pattern, a feature which does not appear in the oscillatory load cases (contrast Figures 5 and 6). The following volumetric evolution is

assumed: 0 (108), 4.5 (25), 5.0 (18), 4.9 (17), 4.1 (15), 3.8 (13), 3.19 (11), 2.65 (9), 2.15 (8), 1.6 (7), 1.3 (6), 1.0 (5.5), 1.1 (5.0), 1.0 (0.15), where the first value represents a factor $\times (V_{max} - V_{min})$ and the second (in parentheses) is the corresponding time in kyr BP. The value of the volume difference, $V_{max} - V_{min}$ ($= 5.65 \times 10^4$ km^3), is identical to Figures 5 and 6.
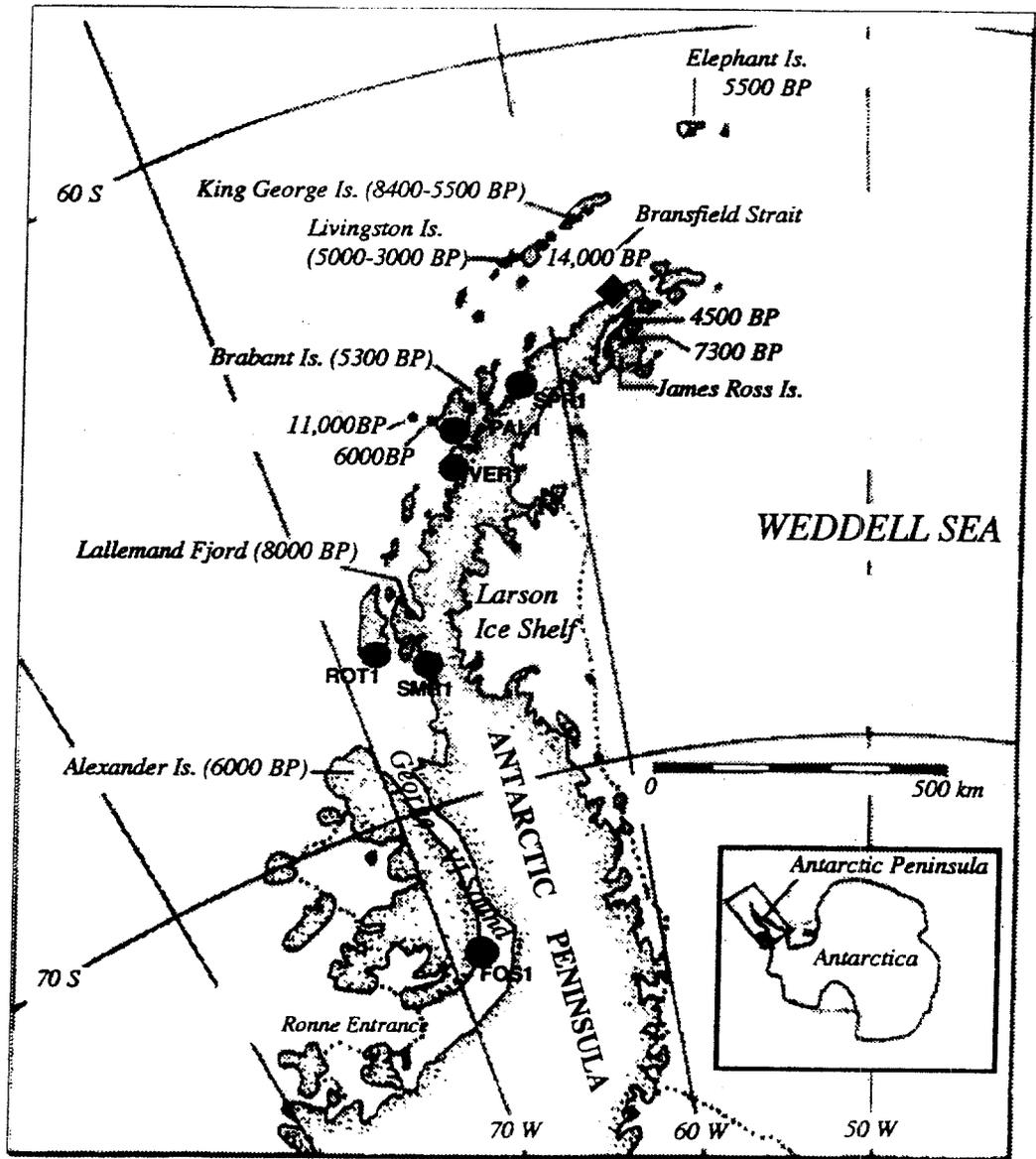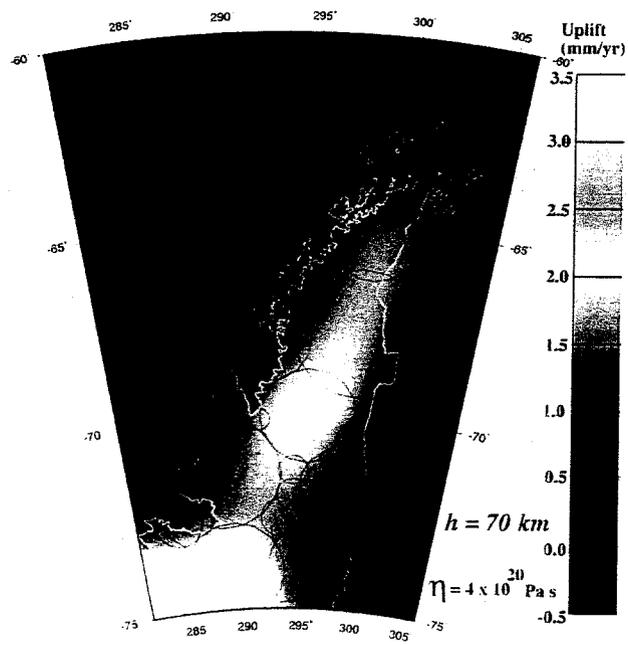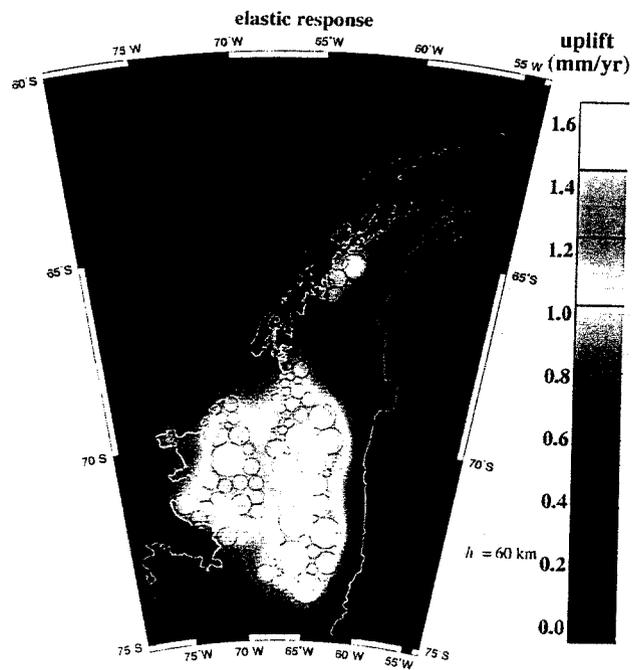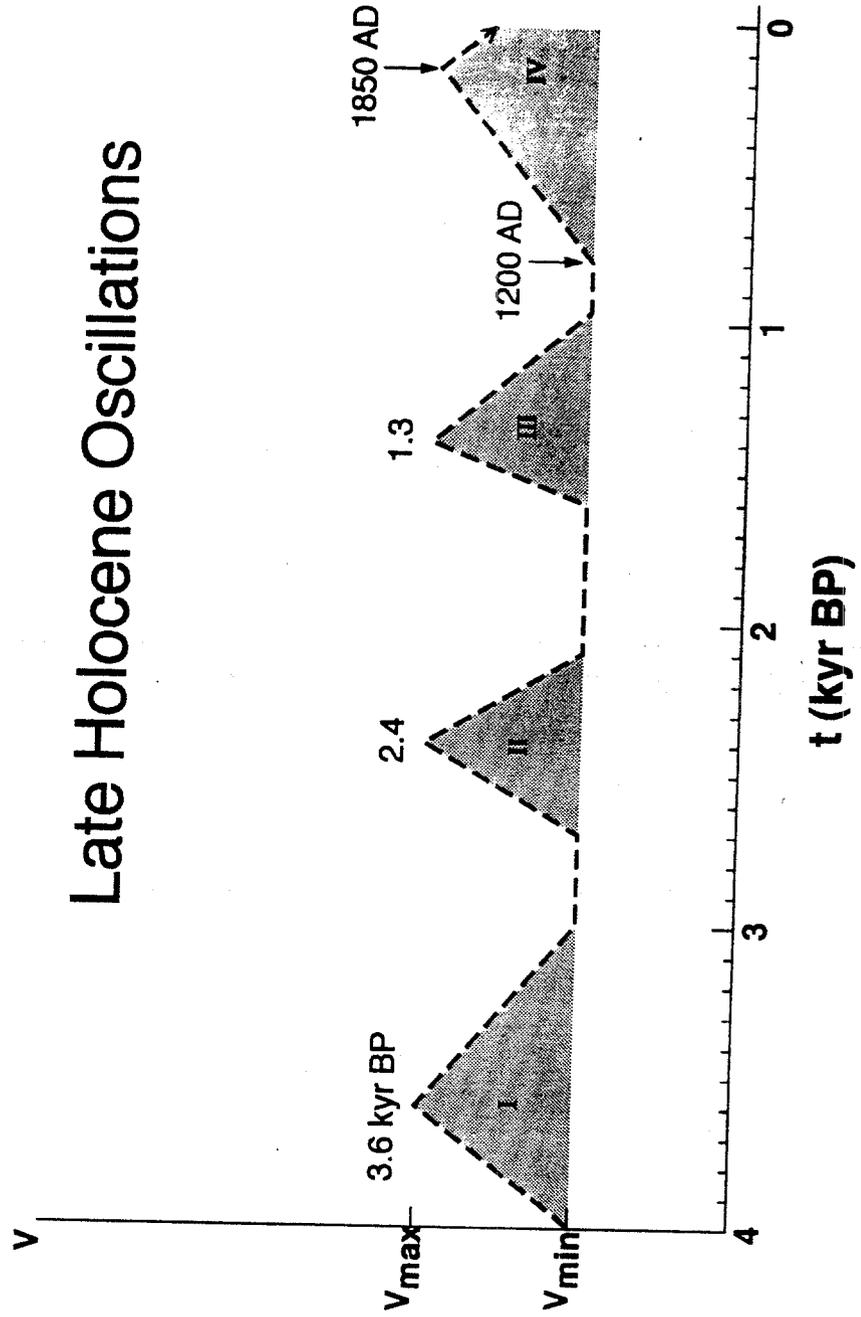
Figure 1

Figure 2

Figure 3

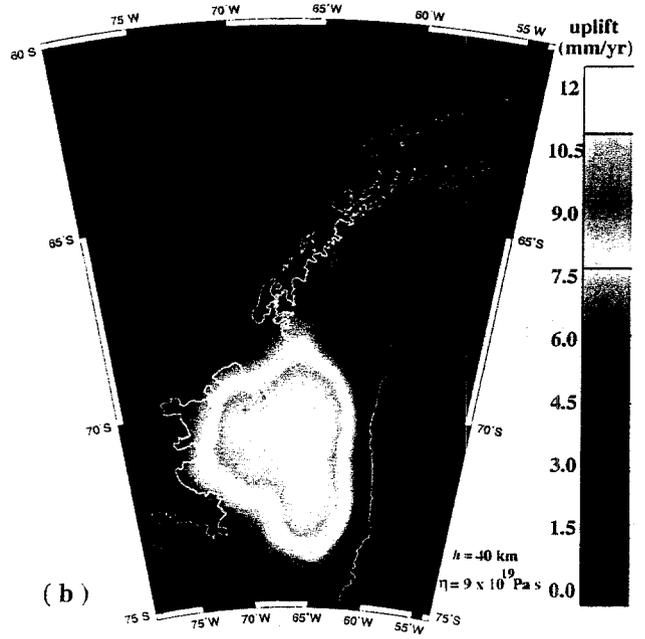# Late Holocene Oscillations

Figure 4
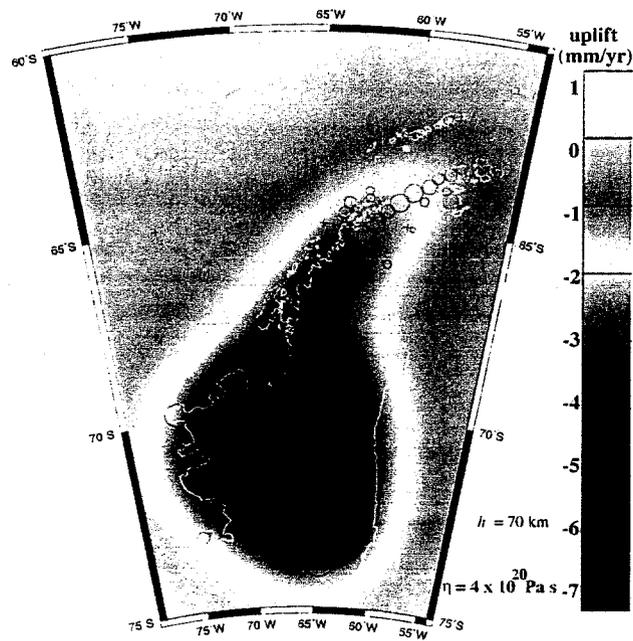
Figure 5
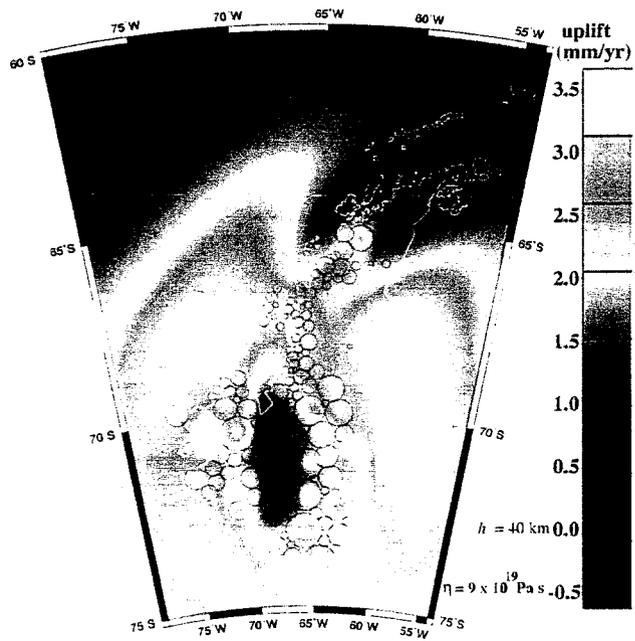
Figure 6

Figure 7a & 7b

Figure 7c