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# A REEVALUATION OF THE ISOSTATIC REBOUND OF LAKE BONNEVILLE

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Abstract. Recent data pertaining to the uplift of the Provo and Bonneville shorelines of the late Pleistocene Lake Bonneville are evaluated in a reanalysis using a rigorous model of deformation of a plane earth comprising an elastic layer overlying a viscoelastic channel. The new observations constrain the wavelength and amplitude of the uplift of two shorelines formed at two different epochs, and they partly resolve the ambiguities in the plausible mechanical parameters found in previous solutions. The results indicate that the effective thickness of the elastic lithosphere is 28 to 30 km, equal to the seismically determined crustal thickness in the region. The estimate of the upper mantle viscosity in the region varies from (1.5-3.4) $\times 10^{19}$ Pa s to (2.1-5.8)$\times 10^{18}$ Pa s, depending on whether the upper mantle is assumed to be of uniform viscosity or the flow is confined to a narrow channel in the asthenosphere. The predicted rate of present uplift at the center of the desiccated lake is 0.06 mm/yr and 0.03 mm/yr for the shallow channel and uniform mantle models, respectively. These values are beyond the first-order leveling accuracies, and the question of the existence of a narrow channel in the region cannot be resolved with the available observations.

## Introduction

The response of the solid earth to various surface loads is important in understanding the rheological properties of the earth. The characteristics of the earth's deformation are functions of, among other parameters, the duration and wavelength of the superposed surface load. The deformation is almost elastic under loads of short duration or small wavelength, whereas creep and relaxation are known to prevail under the influence of large loads with intermediate ($\sim 10^4$ years) and long ($\sim 10^6$ years) periods. The late Pleistocene Lake Bonneville in Utah is one of the best explored surface loading phenomena at intermediate wavelength and duration. The lake area at its maximum was $\sim 6\times 10^4$ km$^2$, which was large enough to stress the earth down to several hundred kilometers. Therefore, the uplift observations in the area once occupied by the lake are capable of yielding useful information on the mechanical properties of the lithosphere and the asthenosphere in the region.

In an earlier work (Nakiboglu and Lambeck, 1982), called paper 1 hereafter, various elastic-viscoelastic layered plane earth models were developed and investigated with respect to their applicability to the surface loading problems involving different geometrical and time scales. It was shown that the three principal parameters defining the rebound of Lake Bonneville are the effective thickness of the lithosphere and the effective thickness and viscosity of the underlying viscoelastic layer. These modeling studies indicate that the effects of the three principal parameters are coupled and that they can be separated only if the wavelength and amplitude of the deformation or the rate of deformation are available at more than one epoch. This ambiguity arises from the fact that the wavelength of deformation (used here to mean twice the distance between the points of maximum and minimum deformation) increases with the thicknesses of the elastic and viscoelastic layers, whereas the amplitude of deformation increases with increasing channel thickness and with decreasing elastic thickness and channel viscosity.

The observations of uplift are confined to the lake boundaries and former islands near the lake center. Consequently, the observed quantity is not the absolute deformation but the central deformation relative to the edge of the lake. The relative deformation increases not only with the increasing absolute deformation but also with decreasing wavelength of the deformation. Therefore, the resolving power of the relative deformation is less, although the resulting ambiguity in the mechanical parameters can be eliminated by the observations at several epochs.

Recently, Passey (1981) presented new field measurements which, when combined with the ones reported by Crittenden (1963a, b), yield the present warping of the two distinct shorelines formed at two different epochs, the Bonneville and Provo epochs. The new observations of the warping of the Bonneville shoreline agree reasonably well with the deformation map of Crittenden (1963a, b) which was adopted in paper 1. By introducing the relative differential uplift at a second epoch, Passey's compilation of the warping of the Provo shorelines provides a valuable constraint on the mechanical parameters.

Theoretically, the problem is still underconstrained, as there are many possible combinations of rheological parameters that fulfill the observations of the relative differential uplift at two epochs. There are, however, some exceptions to this where a decoupling occurs of the effects of one or more of the rheological parameters. These cases are discussed in the following section. Nevertheless, the observations of the warping of Bonneville and Provo shorelines may be combined to further constrain the mechanical parameters of the lithosphere and the asthenosphere in the region, once the chronology of the lake and the variations in the amplitude of the water load have been established.

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The chronology of the various cycles of transgression and regression of Lake Bonneville is based on the stratigraphic studies of the strandline deposits and a few radiocarbon dates from wood, charcoal and freshwater shells (see Crittenden (1963a, b) and Passey (1981) for a detailed discussion). The older and the higher of the two shorelines, called the Bonneville shoreline, was formed when the lake attained this height for some period between 22,000 and 13,000 years B.P. (before the present). However, the radiocarbon datings and the stratigraphic studies are limited, and there is no overall consensus on the chronology of the period. While Crittenden (1963b) assumes that the lake occupied the Bonneville stand twice between 22,000 and 16,000 years B.P. for a total period of ∿4,000 years, Passey (1981) reports a different history due to W. E. Scott in which the stand was occupied only once from ∿16,000 to ∿14,000 years B.P. Again there are marked differences of opinion on the chronology of the Provo epoch. Crittenden (1963b) assumes that the lake occupied the Provo level twice, once from ∿16,000 to ∿13,000 years B.P. and for a second time around 10,000 years B.P. The later studies by Bissell (1968, 1978) suggest a similar chronology for the Provo cycle. On the other hand, Currey (1980) and also Scott, as reported by Passey, place the formation of the Provo shoreline in the interval between ∿14,000 to ∿12,000 years B.P. The different interpretations of the late Pleistocene and Holocene history of the lake seem equally plausible and one needs to consider these differences in the evaluation of the mechanical model parameters. In paper 1 the chronology given by Crittenden (1963b) was adopted, and the errors in viscosity estimates arising from chronological uncertainties were assessed to be about 15%. However, the error estimate was based on the assumption that the Alpine cycle, the earliest part of the history, had the greatest uncertainty. But, if the upper mantle viscosity is as low as suggested by Passey, the memory of the upper mantle is very short. Consequently, even small errors in the loading history of the Holocene epoch may introduce considerable errors in the viscosity estimates.

Another input in the mechanical modeling is the geometry and the amplitude of the water load at various stages of the lake. The areal extents of the lake at the Provo and Bonneville stages are determined from the shorelines with considerable certainty, except perhaps in the southern extremity of the lake where the lake was shallow and hence the shorelines are poorly defined. The water depths have to be determined from the elevation differences between the undeformed shorelines and the lake floor, both of which are unknown. The subsidence and uplift extend well beyond the boundaries of the lake, depending on the thicknesses of the elastic and viscoelastic layers. Therefore none of the present shoreline points can be assumed to be free from uplift or subsidence with any certainty. The problem is solved iteratively by first assuming a deformation for the shorelines and the lake floor and then determining the water load and computing the deformation due to this estimated load. This procedure is then repeated until an agreement is attained between the load and the deformation. The earlier calculations in paper 1 showed that the load given by Crittenden (1963a, b) was essentially correct with a slight error not exceeding ∿10 m, or about 3%.

Passey (1981) estimated the flexural rigidity of the lithosphere and the effective viscosity of the upper mantle using the theoretical models of Brotchie and Silvester (1969) and Vening-Meinesz (1937). Passey's approach is based on the following assumptions: (1) The elastic layer and the underlying mantle are totally uncoupled with respect to their influence on the observables, with the former determining the wavelength of the uplift and the latter being responsible for the rebound which occurred between the Bonneville and Provo epochs. (2) The elastic effects of the mantle are negligible, and the mantle can be represented as a viscous half space. The validity of these assumptions and the applicability of the simplified mechanical model have been discussed in detail in paper 1. Passey justifies the aforementioned simplifications by stating that the Provo and the Bonneville shorelines are compensated to the same degree at the present time and therefore both shorelines found enough time to attain regional isostasy. Hence the relaxation time and the viscosity of the mantle must be comparatively small, the mantle behaving essentially as an inviscid fluid in its present response to the removal of the water load in two stages at the end of the Provo and Bonneville epochs. The difficulties with this argument are (1) that the water load and the absolute uplift are not known accurately enough prior to deformation calculations to enable a rigorous determination of the degree of compensation and (2) that approximately equal compensation of the two shorelines implies not only the possibility of very small viscosity and decoupling but also a relatively high viscosity and a total coupling of the wavelength and amplitude information. This nonuniqueness aspect is discussed in paper 1.

The aim of the present work is to establish the plausible range of upper mantle viscosity and the elastic lithosphere thickness in the Basin and Range Province by utilizing the rigorous theory developed in paper 1 and the new uplift data reported by Passey (1981).

## Computations and Results

The theory of deformation of a layered elastic-viscoelastic plane earth is given in paper 1. Only the relevant results are presented here. The mechanical model comprises an elastic layer overlying a viscoelastic channel which is bounded from below by a high viscosity medium taken to be rigid. The model represents a mantle with uniform viscosity if the thickness of the viscoelastic layer is taken to be sufficiently large in comparison to the areal extent of the surface load. On the other hand, if the flow is confined into a narrow channel in the asthenosphere, the strain and the strain rate will diminish more rapidly with depth, with the deeper parts not participating in the flow behaving like a rigid base (see Figure 15 of paper 1). Therefore the model is able to represent equally the channel flow with the appropriate choice of the thickness of the viscoelastic layer. The mechanical parameters involved in the modeling of

the surface deformation are the thicknesses, the rigidities, and the Poisson's ratio of the top two layers and the viscosity of the second layer.

Taking the elastic layer as a thin plate, assuming that the viscoelastic layer is incompressible and that the boundary conditions are axially symmetrical, one can obtain the Laplace transform of the vertical deformation at the surface due to a disc load of radius A and unit height as

$$\hat{\omega}(r,s) = 6(1-\nu_1)\frac{\rho_\ell g A}{\mu_1} \cdot \int_{u=0}^{\infty} \frac{J_1(au)\, J_0(ux)\, \hat{T}\,(s+\tau^{-1})}{Us + (u^4+h^{*4})/\tau}\, du \qquad (1)$$

where

- $\omega$ vertical deformation;
- $(\hat{\ })$ Laplace transform of a function of time;
- $s$ Laplace transform variable with the unit of frequency;
- $J_0$, $J_1$ Bessel functions of the first kind;
- $U = u^4+h^{*4}-12(1-\nu_1)Fu\,\mu_2/\mu_1$;
- $F = [ud+\sinh(ud)\cosh(ud)]/[(ud)^2-\sinh^2(ud)]$;
- $h^* = h[6g\rho_s(1-\nu_1)/\mu_1 h^3]^{1/4}$
- $\rho_\ell$, $\rho_s$ densities of the load and the viscoelastic layer;
- $r$ distance from the center of the load;
- $\mu_1$, $\nu_1$, $h$ rigidity, Poisson's ratio, and thickness of the elastic layer, respectively;
- $\mu_2$, $\eta$, $\tau=\eta/\mu_2$, $H$ rigidity, viscosity, Maxwell time constant, and thickness of the viscoelastic layer, respectively;
- $T$ history of the disc load giving its height as a function of time.

The dimensionless quantities in (1) are defined as $a=A/h$ and $x=r/h$. The flexural rigidity and flexural distance of the elastic lithosphere are related to the present parameters as

$$D = (\mu_1 h^3)/[6(1-\nu_1)] \qquad L = (D/g\rho_s)^{1/4} \qquad (2)$$

The surface deformation follows from the inverse Laplace transform of (1), once the loading history T is prescribed.

The loading and unloading of Lake Bonneville is represented by 108 to 192 discs situated at 12 different locations in the lake area and each with a radius A=38.25 km (see paper 1 for details). Each individual disc has a different history which is represented by one of the three following simple functions of time:

$$T_1 = h_1[H(t-t_0) - H(t-t_1)] \qquad (3a)$$

$$T_2 = \dot{h}_1\, t[H(t-t_0) - H(t-t_1)] \qquad (3b)$$

$$T_3 = (h_1+\dot{h}_1 t)\,[H(t-t_0) - H(t-t_1)] \qquad (3c)$$

Here H is the Heaviside unit step function, and $h_1$ and $\dot{h}_1$ are the water depth and its rate of change, respectively. As described by relation (3a), a load of height $h_1$ is applied instantaneously at time $t=t_0$, remains constant in the interval $t_0<t<t_1$, and is then removed suddenly at $t=t_1$. The second history $T_2$ implies that the load increases linearly with a rate $\dot{h}_1$ in the interval $t_0<t<t_1$ and is removed suddenly at $t=t_1$. The third history is obtained by the superposition of the first two histories. The time origin for the loading-unloading of Lake Bonneville is chosen to be the beginning of the Alpine cycle at 70,000 years B.P. Therefore the lake retreated from the Bonneville and Provo stands at t=54,000∿56,000 years and t=57,000 ∿ 58,000 years, respectively. The ranges here denote the epochs deduced from the lake chronologies of Crittenden (1963b) and Scott as given by Passey (1981) (Figure 1).

Substituting $T_i$ (i=1,2,3) into (1) and taking the inverse Laplace transform yields the deformation and its rate due to a single disc of history $T_i$ as

$$\left.\begin{matrix}\omega(r,t)\\ \dfrac{d\omega(r,t)}{dt}\end{matrix}\right\} = 6(1-\nu_1)\frac{Ag\rho_\ell}{\mu_1}\int_{u=0}^{\infty} J_1(au)J_0(ux)\frac{1}{\alpha_1}\left\{\begin{matrix}W_i\\ \dfrac{dW_i}{dt}\end{matrix}\right. du \qquad (4)$$

where

$$\alpha_1 = u^4 + h^{*4}$$

$$W_1 = h_1 \sum_{j=0}^{1} (-1)^j [1 - \frac{\alpha_2}{\alpha_1+\alpha_2}\, e^{-\beta(t-t_j)}]H(t-t_j)$$

$$W_2 = \dot{h}_1\tau \sum_{j=0}^{1} (-1)^j [t/\tau-\alpha_2/\alpha_1+(\alpha_2/\alpha_1 - \frac{\alpha_1}{\alpha_1+\alpha_2}\frac{t_j}{\tau}) \cdot e^{-\beta(t-t_j)}]\, H(t-t_j)$$

$$W_3 = W_1 + W_2$$

$$\alpha_2 = -12\, u\, F(1-\nu_1)\frac{\mu_2}{\mu_1}$$

$$\beta = \alpha_1/(\alpha_1+\alpha_2)\tau$$

The deformation at a point resulting from more than one disc load can be obtained by linear superposition of the deformations induced by each disc load. The surface deformation given in (4)

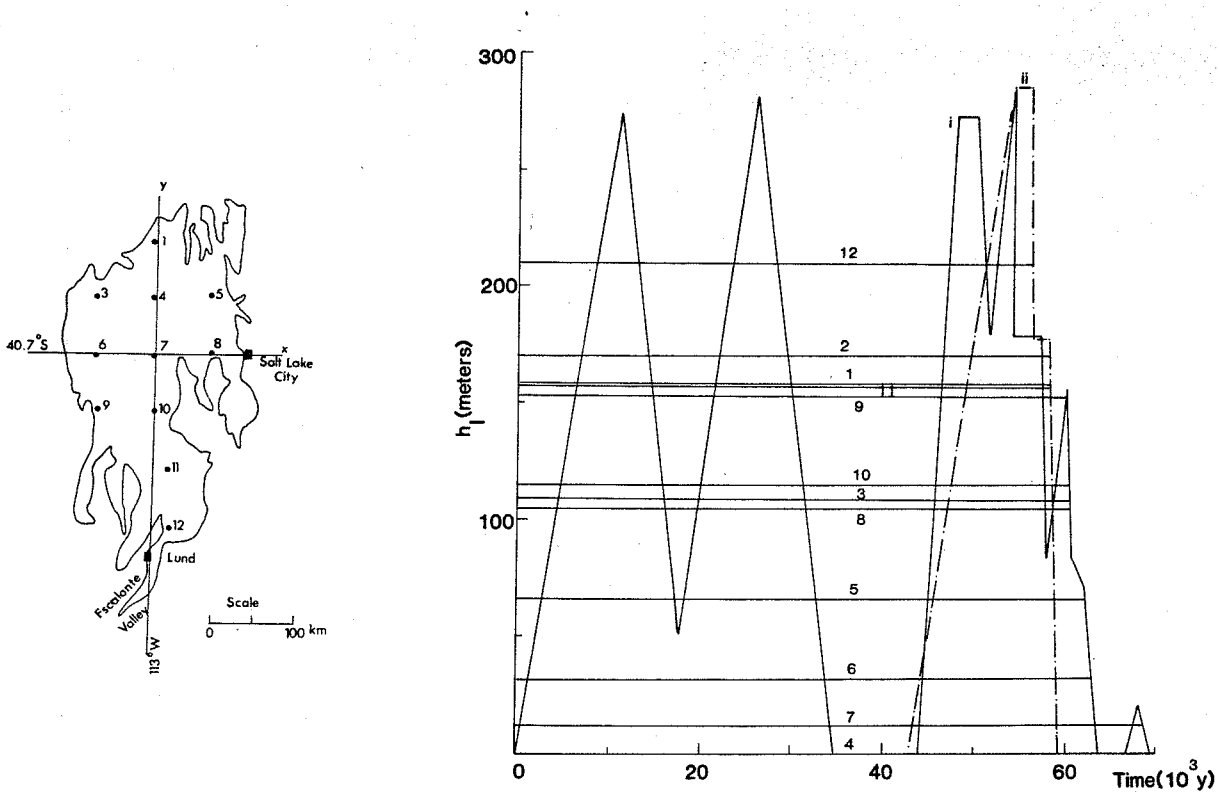



Fig. 1. (Left) The boundary of Lake Bonneville at its maximum stand. The centroids of the disc loads are shown by solid circles. (Right) The loading histories I and II. The horizontal lines numbered from 1 to 12 illustrate the datum elevation at each centroid. For example, water load at disc location 4 loads the earth for nearly the entire history of the lake, while the ones at location 12 load the earth for only four short periods.

is measured with reference to the undeformed surface before the lake water loaded the solid earth. On the other hand, the observations of the warping of the old shorelines yield the amount of relative uplift within the lake area that occurred since the time of unloading. A quantity more useful for the present purposes is therefore the differential deformation which occurred since an arbitrary epoch $t_k$, i.e.,

$$\Delta\omega = \omega(r,t_{present}) - \omega(r,t_k)$$

The $\Delta\omega$ is still not directly comparable to the observations, as it refers to an unknown elevation of the undeformed lake floor. Taking the elevation of a shoreline point sufficiently away from the lake center as the datum, the observations yield the relative differential uplift at an arbitrary location as

$$\delta\omega = \Delta\omega(r,t_k) - \Delta\omega(r_0,t_k)$$

where $r_0$ is the distance from the disc center to the reference point at the flanks of the now desiccated lake. The relative differential uplift as given in (5) is the theoretical estimate which is to be compared with the observations. In the ensuing calculations, the effective rigidity and the Poisson's ratio of the lithosphere are taken to be $\mu_1$=0.323x10$^{11}$ Pa and $\nu_1$=0.299, respectively. The corresponding values for the underlying viscoelastic layer are assumed to be $\mu_2$=0.734x10$^{11}$ Pa and $\nu_2$=0.5.

The estimates of water depths given in paper 1 are modified for the present study by incorporating the calculated deformations in paper 1 and the new uplift data by Passey (1981) (Figure 1). The new load configuration differs only little from the old one, with the maximum difference being 7 m.

Two sets of calculations are performed for the relative differential uplift of the Bonneville and the Provo shorelines. The first set utilizes the lake chronology as given in paper 1. The second set is based on the chronology of Scott as given by Passey (1981) (Figure 1). The two chronologies are termed here as the load histories I and II, respectively. The coordinate system is chosen such that the x and y axes coincide with the 40.7° parallel and the 113° meridian, respectively. The origin is roughly coincident with the location of the center of the lake.

The extent of the lake in the y direction (north-south) is about 450 km, and the uplift appears to be better constrained by the observations in this direction than in the east-west direction. Also, the differential uplift in this latter direction may have been influenced by the presence of the north-south trending Wasatch Fault (see paper 1). Consequently, the theoretical modeling is restricted to fitting the observed differential uplift along the 113° meridian. Similar modeling, using the

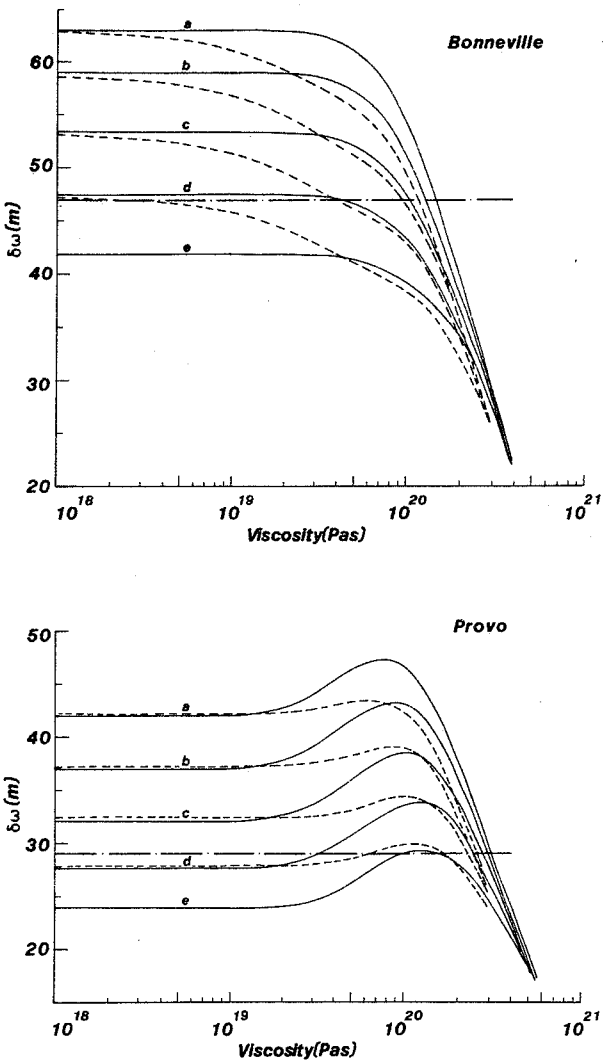



Fig. 2a. Relative differential uplift of the Bonneville and Provo shorelines at the center of the load given as a function of the thickness of the elastic layer and upper mantle viscosity for loading histories I (dashed lines) and II (solid lines). The mantle is assumed to be of uniform viscosity ($H\to\infty$). The thickness of the elastic layer is 15 km (line a), 20 km (line b), 25 km (line c), 30 km (line d), and 35 km (line e). The dot-dashed horizontal line shows the observed relative differential uplift.

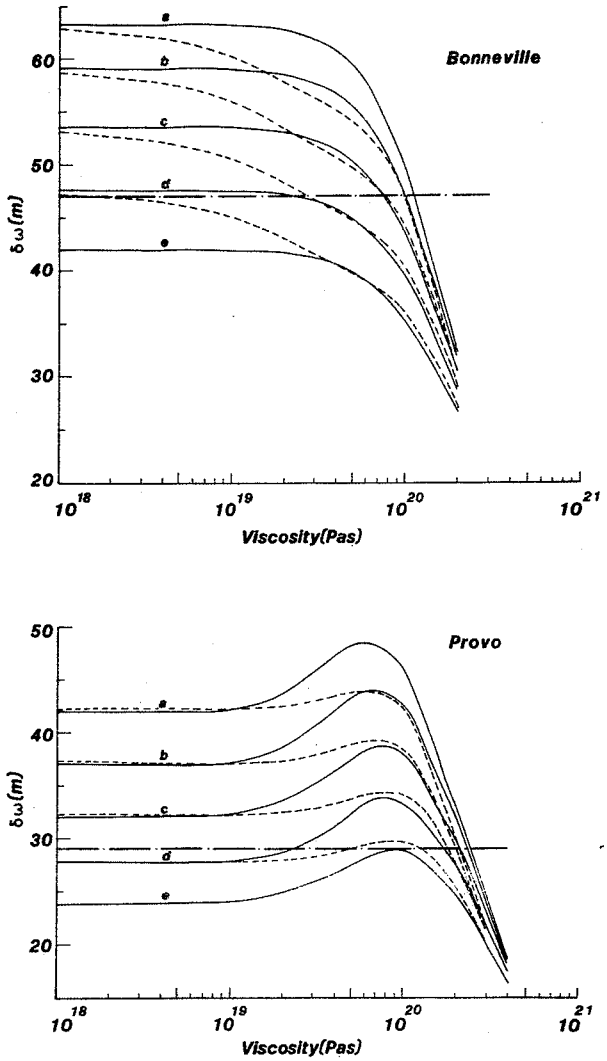
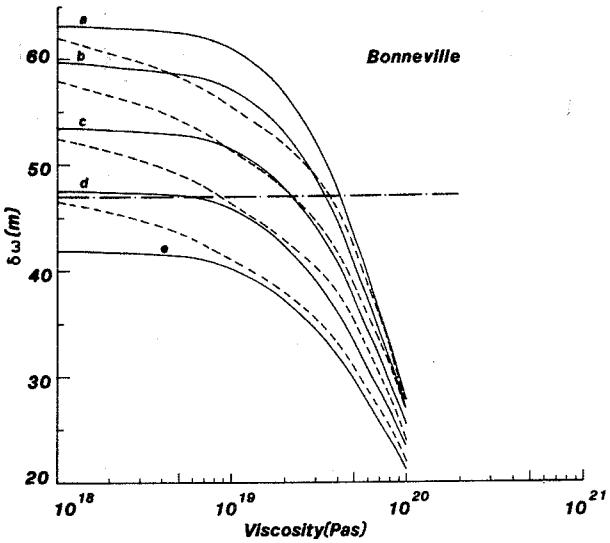



Fig. 2b. Same as Figure 2a but for a channel flow with H=200 km.

east-west profile, yields consistently lower estimates for viscosity. Hence the present estimates, based on the north-south profile, will constitute the upper limit (see paper 1 for the discussion of this anisotropy).

The shoreline deformation is a smooth function of position compared with the water load. This is because the wavelength of the former is longer than that of the load except in the asymptotic limit of exact local isostasy. Therefore one can interpolate between the observations over short distances to predict the relative differential uplift at intermediate points. The old and the new observations are combined, and the relatively differential uplift of the Bonneville and Provo shorelines along the 113° meridian is estimated by linear interpolation. The reference point for the relative differential uplift is chosen at the southern flank of the lake at y=-200 km. The elevation of this flank point is determined from the elevations of the profile points situated at $y<-100$ km by linear regression.

According to the new combined data set, the relative differential uplift at the origin since the end of the Bonneville and Provo epochs is $\delta\omega$=47 m and 29 m, respectively. Obviously, the observed $\delta\omega$ of any of the shorelines can be matched by many different combinations of the unknown mechanical parameters of elastic thickness, channel thickness, and viscosity. The observed $\delta\omega$ incorporates information on both the wavelength and the amplitude of the deformation. The same $\delta\omega$ can fit equally well a broad deformation with large amplitude or a smaller absolute deformation confined into a more restricted area. The former model of deformation implies a relatively thick elastic layer overlying a low-viscosity mantle, whereas the second model implies just the opposite, with a further possibility of a narrow channel flow. Figures 2a-c illustrate the dependence of the



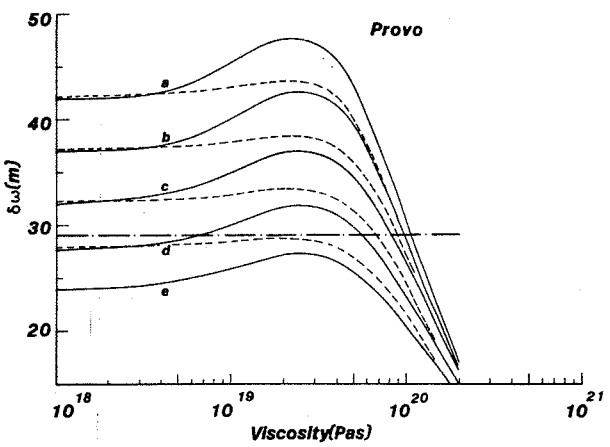




Fig. 2c. Same as Figure 2a but for a channel flow with H=100 km.

relative differential uplift on the rheological parameters involved.

For a uniform mantle (Figure 2a) the effects of viscosity and the elastic thickness are practically decoupled for viscosities below $\sim 2\times10^{19}$ Pa s, and the relative differential uplift is governed by the thickness of the elastic layer. This indicates that the present Provo and Bonneville shorelines should be in a state of regional isostasy if the mantle is uniform and its viscosity is less than $\sim 2\times10^{19}$ Pa s. The corresponding threshold viscosity for the channel flow is lower, with values of $\sim 1\times10^{19}$ Pa s and $\sim 5\times10^{18}$ Pa s for the channel thicknesses of 200 km and 100 km, respectively. At the other extreme, the relative differential uplift is almost independent of the thickness of the elastic layer if the viscosity of the upper mantle exceeds $\sim 4\times10^{20}$ Pa s for the channel thicknesses considered here. Again, a comparison of Figures 2a-c shows that the effect of the elastic layer is marginally more pronounced if the mantle flow is confined into a narrow channel. Figures 2a-c reveal that the observed relative differential uplift is influenced by both the elastic thickness and the viscosity, if the viscosity is within the range $\sim 2\times10^{19} < \eta < \sim 4\times10^{20}$ Pa s. That is, both the amplitude and the wavelength of the absolute uplift are functions of these parameters, resulting in the nonuniqueness mentioned above. An apparently curious feature of the curves for the Provo shoreline in Figures 2a-c is that they imply that two possible viscosity values can yield the same relative differential uplift for given channel and elastic thickness. The Provo shoreline was formed several thousand years after the removal of the Bonneville water load. The rebound is an exponentially decreasing function of time, and a major part of the rebound occurred in this intermediate interval between the two still-stands. If the viscosity is too low, most of the rebound would have been completed during the period, and there would be little rebound after

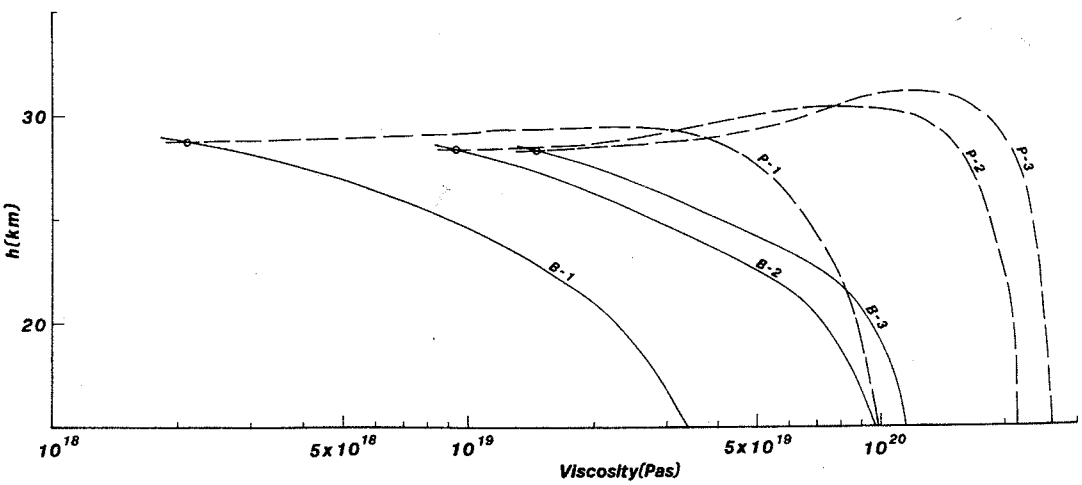



Fig. 3a. Combination of the rheological parameters h and η yielding the observed relative differential uplift for the Bonneville (B) and Provo (P) shorelines for load history I. The curves are for H=100 km (B-1 and P-1), H=200 km (B-2 and P-2), and H→∞ (B-3 and P-3). The circles indicate the viscosity and the elastic thickness which satisfy the Bonneville and Provo observations simultaneously.

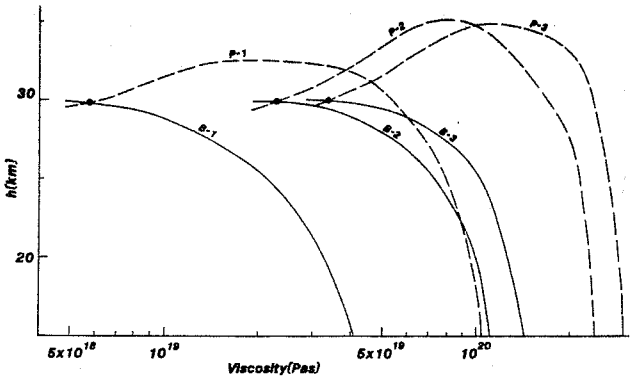



Fig. 3b. Same as Figure 3a but for load history II.

the Provo epoch. On the other hand, an excessively high viscosity leads to only little rebound, again resulting in a small rebound since Provo time. Therefore one is faced with an ambiguity in resolving a viscosity if the observations are confined to the epochs at the tail end of the rebound history. According to Figures 2a-c the only restrictions on the plausible range of parameters are that the elastic layer and the effective viscosity of the upper mantle are smaller than 35 km and $10^{20}$ Pa s, respectively.

One can infer the viscosity for given thicknesses of the elastic and viscoelastic layers from Figures 2a-c such that the theoretical $\delta\omega$ is equal to the observed value at a single epoch. The results of such calculations are summarized in Figures 3a and 3b. Apparently, the elastic thickness and the viscosity are determined uniquely from the observations of the relative differential uplift at two epochs once the depth of the low viscosity channel is given. The results are summarized in Table 1 for three different channel thicknesses. An interesting conclusion is that the thickness of the elastic layer is fixed at 28 to 30 km irrespective of whether or not a channel flow prevails in the mantle. The corresponding range of the plausible flexural rigidity for the lithosphere is $\sim 1.8\times10^{23}$ N m to $\sim 2.1\times10^{23}$ N m. Obviously, the viscosity decreases with decreasing channel thickness, and the present results set the effective viscosity of the upper mantle in the range $1.5\times10^{19} \sim 3.4\times10^{19}$ Pa s, depending on the adopted load history.

The comparison of the observed uplift profile with the theoretical ones is shown in Figures 4a and 4b. The theoretical profiles are almost identical, and they all fit the observations equally well with discrepancies that are generally below about 5 m. The good fit of the central and the flank points indicates the essential correctness of the model in recovering both the wavelength and the amplitude of the uplift.

The available observations cannot resolve whether the mantle flow is confined to a narrow channel or not. A potentially sensitive indicator of the channel thickness is the present rate of uplift and its variation with distance from the center of the lake (see paper 1). The present uplift rate at the center should be larger for a shallow channel flow because the confinement of the flow into a narrow zone retards the attainment of isostasy. Also, the extent of the positive uplift region increases with decreasing channel depth. Therefore one would observe ongoing subsidence in places like the Escalante Desert in the southern extremity of the lake if the flow is restricted to a shallow channel. Observations of the deformation in these regions are, however, ambiguous, in part because strandlines are poorly developed there. However, the present results indicate that the elastic layer is quite thick and the compatible viscosities are extremely small for the channel flow. The calculated uplift rate at the center of the lake is $\sim$0.06 mm/yr for the uniform mantle and $\sim$0.03 mm/yr for a 100 km thick channel. Although these values differ by a factor of 2, they are well beyond the accuracy of the first-order leveling operations repeated over several decades. Such low values for the present uplift rate indicate that the present topography of the lake area is in a state of regional isostasy. This confirms Passey's earlier deduction.

## Comments

The present estimates of viscosity and flexural rigidity using the load history II exceed by a factor of 2 the values obtained by Passey (1981) using the same chronology. The discrepancy may be attributed partly to the

TABLE 1. Estimates of Mantle Viscosity and Lithospheric Flexural Rigidity on the Basis of Deep and Channel Flow Models

| | Viscosity, $10^{19}$ Pa s | | Flexural rigidity, $10^{23}$ N m | |
|---|---|---|---|---|
| Mantle Model | Load History I | Load History II | Load History I | Load History II |
| Uniform mantle ($H\to\infty$) | 1.5 | 3.4 | 1.8 | 2.1 |
| Channel flow (H=200 km) | 0.93 | 2.3 | 1.8 | 2.1 |
| Channel flow (H=100 km) | 0.21 | 0.58 | 2.0 | 2.1 |

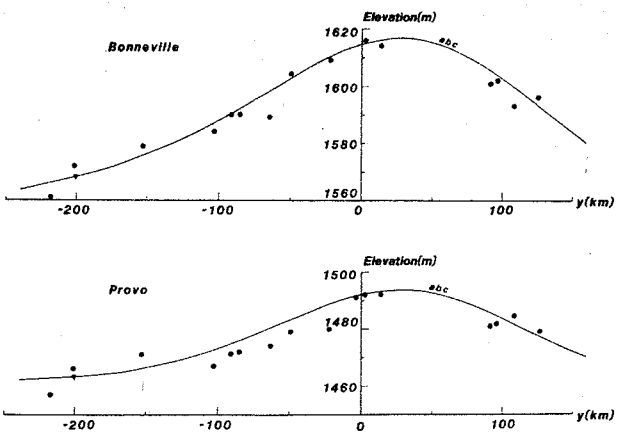



Fig. 4a. Comparison of the observations of relative differential uplift of the Bonneville and Provo shorelines (solid circles) with the computed ones using the flexural rigidity and viscosity estimates for load history I (see Table 1). The uplift profiles are for a uniform mantle (curve a), H=200 km (curve b), and H=100 km (curve c). The solid triangle is the reference flank point.

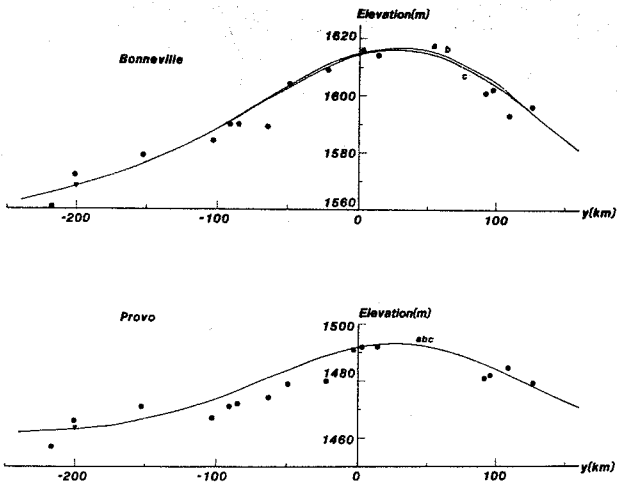



Fig. 4b. Same as Figure 4a but for load history II.

approximations involved in the theoretical model of Vening-Meinesz (1937) as applied by Passey to a much smaller load than the initial model was intended for. However, the viscosity estimates based on the two equally plausible load histories differ again by a factor of 2. Therefore the observational uncertainties related to the reconstruction of the load partly justify the approximate theoretical model. The flexural rigidity of the lithosphere is constrained much better, with less than 10% errors introduced by the uncertainty in the load history. The remaining observational uncertainties such as the ones pertaining to the water depths and the areal extent of the lake are unlikely to perturb the viscosity estimates by more than 10∿15% (see paper 1 for a detailed error analysis).

The present theoretical modeling indicates that the maximum uplift of the Bonneville and Provo shorelines is in the range of 55 ∿ 59 m and 31 ∿ 32 m, respectively, depending on the lake history and the channel thickness. These values are about 10 m smaller than those adopted by Passey (1981) on the basis of arbitrary datums for zero uplift.

The present estimate of the effective elastic thickness of the lithosphere, 28 to 30 km, is in a remarkable coincidence with the seismically determined crustal thickness of 30 km in the Basin and Range Province (Braile et al., 1974). If not fortuitous, the agreement suggests that the crust in this region behaves more or less elastically in response to loads of small amplitude for time periods of $7 \times 10^4$ years. In paper 1 we obtained a much lower estimate for the effective elastic thickness. This was a consequence of attempting to fit the model to shorelines in the southern extremity of Lake Bonneville investigated by Anderson and Bucknam (1979). The effective viscosity of the upper mantle in the Basin and Range Province is $1.5 \times 10^{19}$ Pa s to $3.4 \times 10^{19}$ Pa s. However, the viscosity may be as low as $2 \times 10^{18}$ Pa s if there exists a thin layer of low viscosity in the province. The low-velocity layer in the region extending from 50 km to 200 km depth (Burdick and Helmberger, 1978) may be suggestive of such a low-viscosity channel. The effective viscosity of the whole mantle is $10^{20}$ Pa s to $10^{21}$ Pa s as deduced from the nontidal acceleration and polar wander of the earth as well as from the long-wavelength response of the earth to changing sea levels due to late Pleistocene deglaciation (Nakiboglu and Lambeck, 1980; Nakiboglu et al., 1983). These results combined with the present estimates of the upper mantle viscosity suggest that the earth's mantle has a fairly uniform viscosity with perhaps a small increase with depth.

## References

Anderson, R.E., and R.L. Bucknam, Two areas of probable Holocene deformation in southwestern Utah, Tectonophysics, 52, 417-430, 1979.

Bissell, H. J., Bonneville - An ice age lake, Brigham Young Univ. Geol. Stud., 15(4), 66 pp., 1968.

Bissell, H. J., Biography of Bonneville: An ice age lake, Geol. Soc. Am. Abstr., 10, 211, 1978.

Braile, L. W., R. B. Smith, G. R. Keller, R. M. Welch, and R. P. Meyer, Crustal structure across the Wasatch Front from detailed seismic refraction studies, J. Geophys. Res., 79, 2669-2677, 1974.

Brotchie, J. F., and R. Silvester, On crustal flexure, J. Geophys. Res., 74, 5240-5252, 1969.

Burdick, L. J., and D. V. Helmberger, The upper mantle P velocity of the western United States, J. Geophys. Res., 83, 1699-1712, 1978.

Crittenden, M. D., New data on the isostatic deformation of Lake Bonneville, Prof. Pap. U.S. Geol. Surv., 454-E, 1963a.

Crittenden, M. D., Effective viscosity of the earth derived from isostatic loading of Pleistocene Lake Bonneville, J. Geophys. Res., 68, 5517-5530, 1963b.

Currey, D. R., Coastal geomorphology of Great Salt Lake and vicinity, in Great Salt Lake: A Scientific, Historial and Economic Overview,

Bull. Utah Geol. Miner. Surv., vol. 116, edited by J. W. Gwynn, Utah Geological and Mineralogical Survey, Salt Lake City, 1980.

Nakiboglu, S. M., and K. Lambeck, Deglaciation effects on the rotation of the earth, Geophys. J. R. Astron. Soc., 61, 49-58, 1980. (Corrigendum, Geophys. J. R. Astron. Soc., 64, 559, 1981.)

Nakiboglu, S. M., and K. Lambeck, A study of the earth's reponse to surface loading with application to Lake Bonneville, Geophys. J. R. Astron. Soc., 70, 577-620, 1982.

Nakiboglu, S. M., K. Lambeck, and P. Aharon, Sea level changes in the Pacific: Implications with respect to deglaciation history and local tectonics, Tectonophysics, 91, 335-358, 1983.

Passey, Q. R., Upper mantle viscosity derived from the difference in rebound of the Provo and Bonneville shorelines: Lake Bonneville Basin, Utah, J. Geophys. Res., 86, 11701-11708, 1981.

Vening-Meinesz, F. A., The determination of the earth's plasticity from the post-glacial uplift of Scandinavia: Isostatic adjustment, Proc. K. Ned. Akad. Wet., Ser. B, 40, 654-665, 1937.

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