Geodynamics and Temporal Variations in the Gravity Field

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N90-20549

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ABSTRACT

Just as the Earth's surface deforms tectonically, so too does the gravity field evolve with time. Now that precise geodesy is yielding observations of these deformations it is important that concomitant, temporal changes in the gravity field be monitored. Although these temporal changes are minute they are observable: changes in the J2 component of the gravity field have been inferred from satellite (LAGEOS) tracking data; changes in other components of the gravity field would likely be detected by Geopotential Research Mission (GRM), a proposed but unapproved NASA gravity field mission. Satellite gradiometers have also been proposed for high-precision gravity field mapping. Using simple models of geodynamic processes such as viscous postglacial rebound of the solid Earth, great subduction zone earthquakes and seasonal glacial mass fluctuations, we predict temporal changes in gravity gradients at spacecraft altitudes. We find that these proposed gravity gradient satellite missions should have sensitivities equal to or better than 10^{-4} E in order to reliably detect these changes. We also find that satellite altimetry yields little promise of useful detection of time variations in gravity.

1. INTRODUCTION

Because the solid Earth is dynamic, its internal density structure evolves with time. This evolution in density structure produces temporal changes in the gravity field. And, although these changes in gravity are quite small ($dg/g < 10^{-6}$, local) over time spans of years or decades, they could tell us much about dynamics of the Earth's interior. Observations of these time variations will be particularly useful when combined with precise observations of surface deformation.

This paper examines the prospects for detecting these time variations from satellites. Changes in the second zonal harmonic, J2, of the geopotential have been detected [Yoder et al., 1983; Rubincam, 1984] using 6 years of LAGEOS tracking data. These changes have been attributed to ongoing viscous rebound of the Earth's crust and mantle [Wu and Peltier, 1983]. Other high altitude satellites such as Starlette, LAGEOS, II, III, and Stella will help refine estimates of temporal changes in the long-wavelength (λ>4000 km) components of the gravity field. A proposed system of two, low-altitude satellites, the GRM [Taylor et al., 1983] would, were it flown successfully, have very likely detected other temporal changes in the gravity field such as those in the nonzonal and intermediate-wavelength components [Wagner and McAdoo, 1986, hereafter WM]. Gradiometer missions which have been proposed [Paik, 1981; Balmino, 1986] for mapping the gravity field could make useful observations of these time variations if these missions are sufficiently high in sensitivity and low in altitude.

2. GEOPHYSICAL RATIONALE

Gravity fields derived from satellite data possess an ever increasing accuracy at longer spatial wavelengths. The importance of this accuracy to geophysicists is not obvious. For most geophysical applications, high spatial resolution (inherently lacking in satellite gravity fields) is more important than accuracy at lower resolution. However, one application does require this very high accuracy: the geodynamical study of active regional deformation in the crust and mantle. Mass motion in these deformation zones produces slow, subtle changes in the gravity field. If we can detect these changes, we can place significant constraints on models of deformation. Figure 1 shows a simple example of how observations of changes in gravity might discriminate between two possible modes of compressional, crustal deformation: swelling and folding, modes which might well be indistinguishable in observations of surface strain.

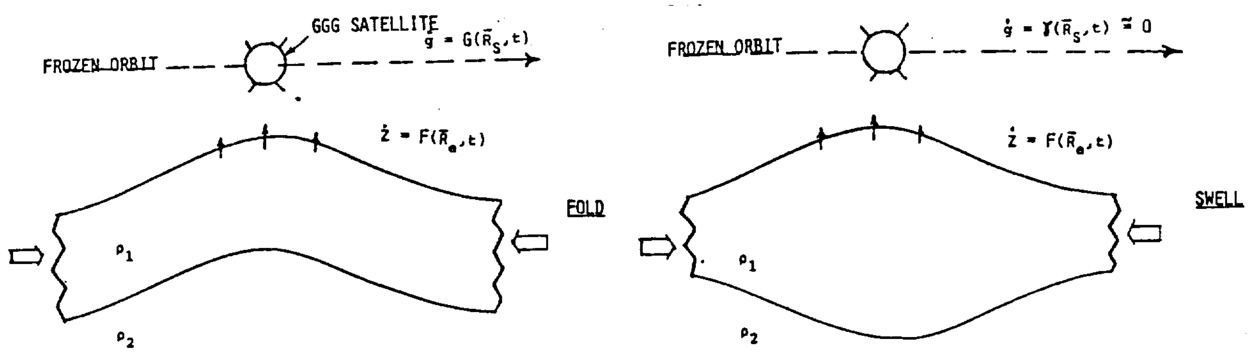

Figure 1. Folding and swelling: a similar vertical motion signal but folding produces greater changes in gravity.

3a. DIP-SLIP EARTHQUAKES

The most energetic earthquakes are the great, thrust faulting events which occur at subduction zones. These, among all earthquakes, should produce the largest changes in the gravity field at satellite elevations. Assume that a gradiometer satellite is operating in a repeating orbit when such an earthquake occurs, and that this orbit lies directly above the epicenter and is orthogonal to the strike of the subduction zone. Characteristics of the great dip-slip event are taken to be identical to those of the 1964 Prince William Sound, Alaska event ($M_o = 8 \times 10^{29}$ dyn cm). Resultant coseismic changes in the gravity gradient field at the spacecraft (Fig. 2) as well as geoid changes are computed using an extension [WM] of a model due to Walsh and Rice, [1979]. This model represents the earthquake as dislocations in a uniform elastic half-space and predicts that the surface free-air gravity anomalies produced by dip-slip events are proportional to the coseismic surface height changes [WM]. We have upward-continued and transformed predicted surface gravity to obtain the horizontal (xz) component of the gradient (z is vertical and x is along-track). The corresponding coseismic geoid change has an amplitude of <2 cm which would be difficult to detect with a satellite altimeter due to omnipresent oceanographic 'noise'. Note also that gradient signals at a spacecraft altitude of 160 km are about 4 x 10^{-3} Eotvos units or E in amplitude. The 1964 Alaska earthquake was, however, extraordinarily large, perhaps the second largest in seismological history. A more typical great earthquake is the 1985 Michoacan, Mexico event [$M_o = 0.2 \times 10^{29}$ dyn cm; Eissler et al., 1986] which was also a thrusting event but was of smaller spatial extent. Predicted changes in horizontal gravity gradients aloft are shown in Figure 3. Note diminished signal at 160 km as well as the more pronounced attenuation at 230 km altitude. Detection of this event requires that a gradiometer with a sensitivity at least 10^{-4} E be flown at the lowest possible altitude.

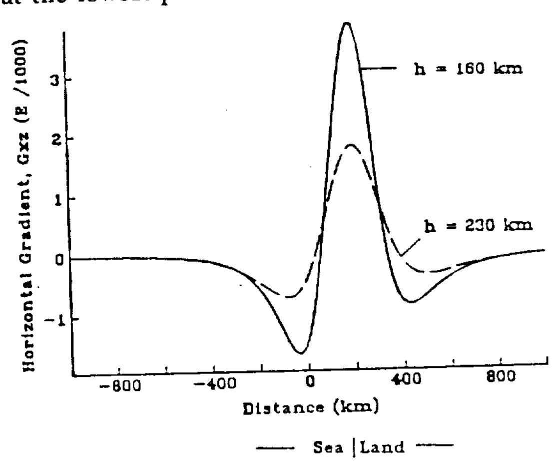

Figure 2. 1964 Prince William Sound, Alaska earthquake. Change in xz component of gravity gradient. Satellite elevation is h.

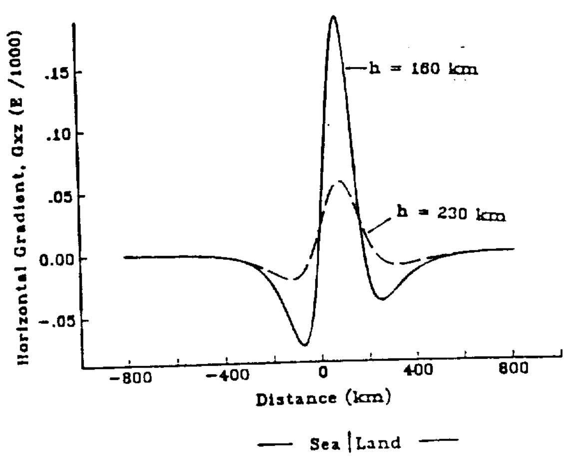

Figure 3. Predicted change in xz component of gravity gradient for 1985 Michoacan, Mexico earthquake.

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3b. POSTGLACIAL REBOUND

The Earth's mantle and crust continues to rebound today in response to the last global deglaciation which began 18000 years ago and largely ceased 6000 years ago. Detailed models [Wu and Peltier, 1983] of this process have been developed and extended [Yoder et al., 1983; Rubincam, 1984] to compute concomitant dynamic changes in the gravity field. For this study, we use a simpler model of [WM], which takes the Earth to be a uniform, Maxwell viscoelastic sphere and the ice masses (Laurentide, Fennoscandian and Antarctica) to be circular, spherical caps. This simple model predicts vertical velocities of the solid Earth's surface (Fig. 4) which agree rather well with both observations and predictions of more complex models. Using Hotine (1969) we compute, from geopotential changes, the change in the XZ gradient component (Figure 5: Z is polar and X is equatorial) which will accrue in one year's time at an elevation of 230 km. Note that the amplitude of this change is about 10^{-5} E. So, to detect the long wavelength gravitational effects of postglacial rebound with a satellite gradiometer, one must have an instrument (assumed lifetime of one year) with a sensitivity of 10^{-5} E. Satellite gradiometry (measuring second derivatives of potential) is better suited to detecting shorter wavelength signals.

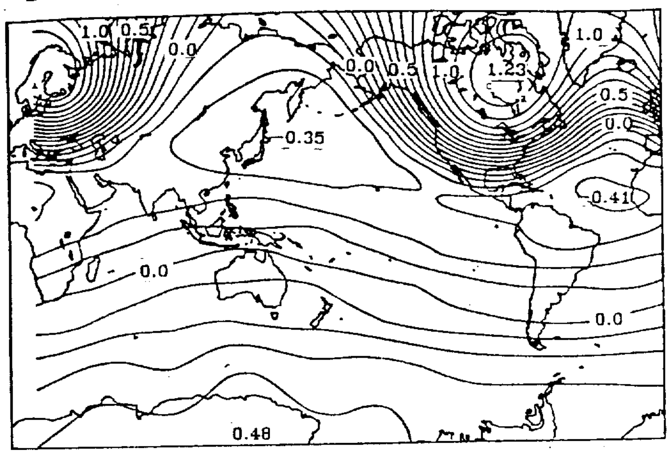

Figure 4. Present-day vertical velocity (cm/yr) of solid Earth's surface from simplified model of postglacial rebound. [WM]

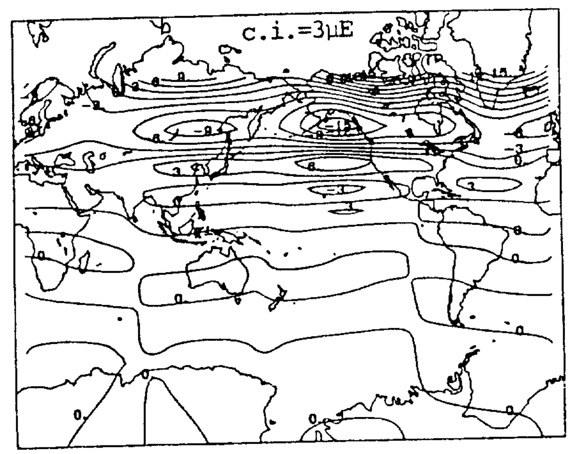

Figure 5. Predicted annual change (μE) in xz component of gravity gradient due to postglacial rebound.

3c. SEASONAL GLACIAL MASS FLUX

Secular and season glacial mass flux is large enough to produce a potentially detectable temporal variation in the gravity field. Meier [1984] suggests that the secular component of mass ablation from smaller glaciers (*i.e.*, Greenland and Antarctica excluded) is the dominant contributor to the nonsteric rise in global sea level. He also estimates season mass flux for each of 25 smaller glacier systems. By far, the largest among these estimates is the 225 km^3 (water equivalent) amplitude annual flux from the Gulf of Alaska Coastal Mountains. This flux occurs as annual accumulation and ablation over 88,400 km^2 with an areal mean amplitude of 2.54 m. The seasonal drop in the geoid (2 cm) produced by this winter-to-summer ablation is estimated (Fig. 6a) by assuming a source body which is a rectangular prism (100 x 884 km x 5.1 m) of density -1.0 x 10^3 kg/m^3. We searched the GEOSAT altimeter data from the Gulf of Alaska for such a geoid change. By averaging in space (over a swath of several groundtracks) and time (over five 17-day repeat cycles), we obtained an ascending and a descending sea surface profile to represent the Gulf in both the summer and winter of 1987. We subtracted the winter results from the summer to estimate seasonal change in the sea surface topography. We found this change (14cm) too large to be geoidal. More likely, it is a steric sea level change. This demonstrates the obstacle which oceanographic variability presents to detection of time variations in gravity from satellite altimeters.

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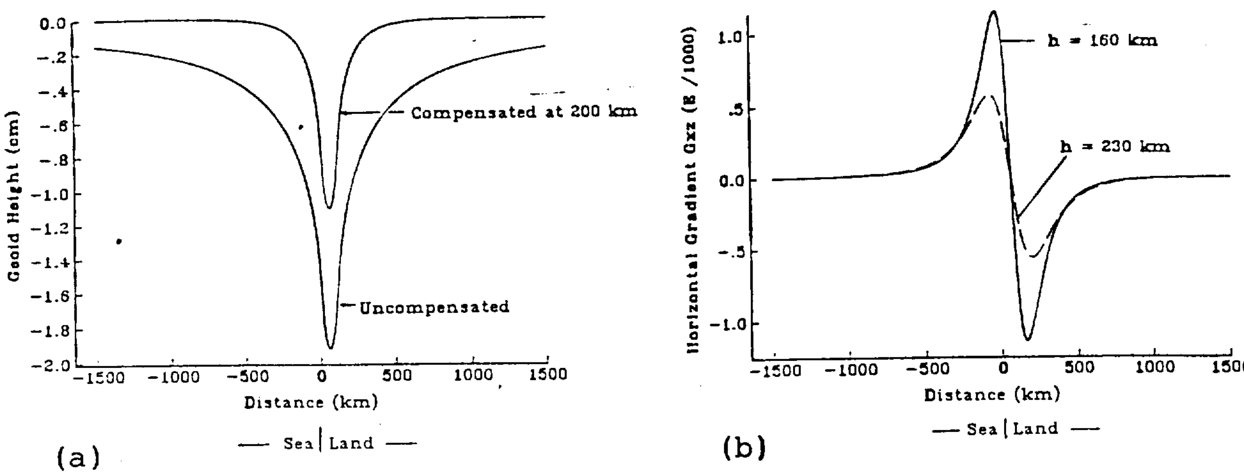

Figure 6. (a) Predicted geoid change (summer minus winter) due to seasonal mass ablation from southern Alaska glaciers. (b) Associated change in xz component of gravity gradient at elevation, h.

We also estimate (Fig. 6b) the change in the horizontal gravity gradients (xz) at satellite elevation for this seasonal glacial mass flux along coastal Alaska. At 160 km elevation, this signal has an amplitude of 10^{-3} E. A gradiometer with a sensitivity, again, of about 10^{-4} E is needed to reliably detect it.

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