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Multi-Year Elevation Changes Near the West Margin of the Greenland Ice Sheet from Satellite Radar Altimetry

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ABSTRACT

Mean changes in the surface elevation near the west margin of the Greenland ice sheet are measured using Seasat altimetry and altimetry from the Geosat Exact Repeat Mission (ERM). The Seasat data extend from early July through early October 1978. The ERM data extend from winter 1986-87 through fall 1988. Both seasonal and multi-year changes are measured using altimetry referenced to GEM T2 orbits. The possible effects of orbit error are minimized by adjusting the orbits into a common ocean surface. Seasonal mean changes in the surface height are recognizable during the Geosat ERM. The multi-year measurements indicate the surface was lower by 0.4 ± 0.4 m on average in late summer 1987 than in late summer 1978. The surface was lower by 0.2 ± 0.5 m on average in late summer 1988 than in late summer 1978. As a control case, the computations are also carried out using altimetry referenced to orbits not adjusted into a common ocean surface.

INTRODUCTION

After the last glacial maximum about 18,000 years before present (B.P.), the western margin of the Greenland ice sheet retreated from its position near the coast to near its present position [Weidick, 1984]. In a summary of marginal fluctuations inferred from geologic evidence, Weidick [1984] states that this occurred between ≥10,000 and 8000–6000 B.P., and that minor re-advances and retreats have occurred since then. A gap in the record, about which little is known, occurred between 6000 B.P. and the Little Ice Age due to destruction of proglacial deposits during re-advances. The Little Ice Age occurred between about 450 and 150 years ago [Grove, 1988]. An advance occurred in the 1880s, during the last neoglacial maximum, and a minor readvance occurred around 1920. According to Weidick [1984] the margin has been retreating since then, although the rate of retreat has slowed during recent decades. Continuous advance in some sectors, since at least the early part of the 20th century, has been superimposed on the general pattern of retreat.

More recent interest in the ablation area of West Greenland [e.g., Braithwaite and Olesen, 1989; Lingle et al., 1990] has resulted from the possibility that increased melting of the Greenland ice sheet may contribute to the increasing rate of sea level rise expected during the next century due to climatic warming, which is predicted because of increasing CO_2 and trace greenhouse gases. The National Research Council [1985] estimated that this ice sheet may account for 10–26 cm of sea level rise by A.D. 2100 [Bindschadler, 1985]. In contrast, the Intergovernmental Panel on Climatic

Change [IPCC, 1990] Report on Sea Level Rise estimates that the Greenland ice sheet may contribute 0.5–3.7 cm to rising sea level by A.D. 2030 [Warrick and Oerlemans, 1990]. If extrapolated, the latter estimate suggests a contribution of roughly 5.4 ± 4.1 cm from this source by A.D. 2100. The substantial difference between the NRC and IPCC estimates is a reflection of current uncertainty regarding the overall mass balance of the Greenland ice sheet.

The only direct measurement of a multi-year mean change in the surface elevation of a large area of the Greenland ice sheet (the southern half, approximately) has been made by Zwally et al. [1989], using satellite radar altimetry. They found that during the 7-year interval between Seasat and the Geosat Geodetic Mission (GM), there was a mean increase in the surface height equivalent to a linear rate of increase of 20 ± 6 cm yr^{-1}. Most of the measurements used in this determination were over the central regions of the ice sheet, due to relatively poor altimeter tracking near the margins. Zwally [1989] pointed out that this is equivalent to a sea level lowering of 0.2–0.4 mm yr^{-1}, depending on whether the increase of the surface height is short-term, consisting mostly of increased snow depth, or long-term, consisting mostly of increased ice thickness. (See also Douglas et al. [1990] and Zwally et al. [1990a].) The GM consisted of the initial 18 months following the spring 1985 launch of Geosat. Subsequently Geosat was maneuvered into an orbit geometry closely following the previous Seasat ground tracks in order to carry out the Exact Repeat Mission (ERM), which started in fall 1987 [e.g., Douglas and Cheney, 1990].

Seasonal mean changes in the surface height of the West Greenland ablation area were measured by Lingle et al. [1990], using altimetry from the Geosat GM. Here we extend that study, by measuring multi-year mean changes in the surface height throughout a larger area that includes the ablation area and extends farther up the ice sheet to about the 2000-m contour (Figure 1). Elevation differences are measured at orbit crossover points during the Geosat ERM, and also between Seasat and the ERM. A crossover point is a location where an orbit ascending in latitude is later crossed by an orbit descending in latitude, or the reverse, so two measurements of the surface height, separated by a time interval, are obtained at approximately the same location. Seasonal and multi-year mean changes in elevation are estimated by averaging the crossover differences throughout the region.

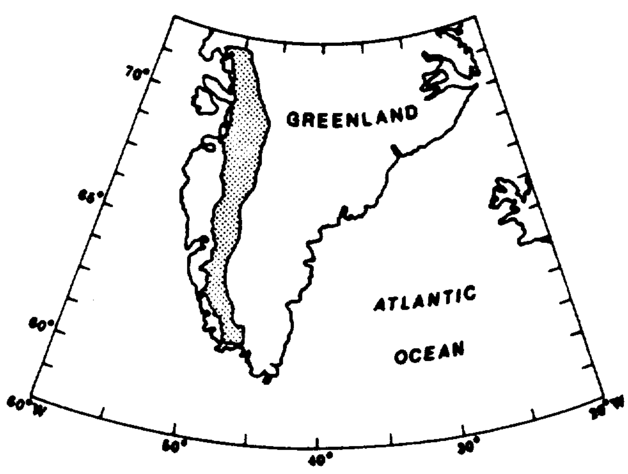

Figure 1. The stippled area is the measurement region, which includes the ablation area of the western Greenland ice sheet and extends up to the 2000-m elevation contour, approximately, and to Lat. 72°N.

ORBITS AND ORBIT ADJUSTMENT

Precision orbit determination is a fundamental aspect of altimetry, because the position of the satellite must be accurately known at the time of each measurement. Precision orbits are computed by iterating to a self-consistent solution among the observations from the tracking stations, which are widely separated on the earth's surface, and the equation of motion for the satellite, which must be satisfied everywhere. The numerical solutions of the equation of motion require an accurate model of the gravitational potential field, and nonconservative forces, primarily air drag and radiation pressure due to the solar wind and radiation upwelling from the earth's surface, must be taken into account [see, e.g., Stewart, 1985, pp. 260–309]. Precision orbits for the Geosat ERM have been computed by Haines et al. [1990] and Shum et al. [1990].

The most accurate model of the gravitational potential field (based entirely on satellite tracking observations) available at the time of this study was the Goddard Earth Model T2, or GEM T2 [Marsh et al., 1989]. This model was employed for re-computation by Haines [1991] of the Seasat orbits used in this study, as well as for computation of the Geosat ERM orbits [Haines et al., 1990]. The Greenland altimeter measurements employed here, which were obtained by two different satellites, are thus with respect to orbits computed using the same gravity model. The radial (vertical) precision of the GEM T2 Geosat ERM orbits is estimated to be about 0.35 m RMS [ibid].

Radial orbit error tends to result primarily from inaccuracies in the spherical harmonic expansions used to represent the gravitational potential field, i.e., the gravitational field is not equally well represented everywhere, and from errors in the estimates of air drag and radiation pressure, which must be made from other models. Radial orbit errors tend to be spatially well correlated, so it is necessary to be cognizant of the possibility that an apparent change through time in the height of a surface in a particular region, such as western Greenland, may be an expression of time-dependent orbit error. Although the 0.35 m RMS radial precision of the Geosat ERM orbits is excellent, time-dependent orbit error in particular regions can be larger than this.

The possibility of orbit error is taken into account by adjusting the Seasat and Geosat ERM orbits crossing Greenland into a common ocean surface, which is the Seasat/Geos-3 mean surface derived from the global Seasat and Geos-3 altimetric data sets [Marsh et al., 1986, 1990]. Short-arc adjustments are used; that is, the orbits are adjusted into the ocean surface immediately on either side of Greenland. A detailed description of the method is given by Zwally et al. [1990b]. Briefly, for each arc the differences (residuals) are computed between the ocean elevations derived from the data along that arc, and the smoothed Seasat/Geos-3 ocean elevations along the same arc, on both sides of Greenland. The residuals are plotted versus time for satellite travel along the arc, and a least-squares straight-line fit is performed. The corrections for the ice sheet measurements along the arc are then obtained by evaluating the linear function at the corresponding times. The underlying assumption is that if the orbit error were zero, the ocean surface measured along the arc would differ only negligibly from the mean ocean surface along the same arc.

It is important to note that this strategy cannot be expected to remove all traces of orbit error because the ocean surface is not static. Adjustment inaccuracy for an individual pass can be introduced, for example, by inadequately modeled ocean tides at the time of the pass, by changes in the height of the sea surface caused by mesoscale changes in atmospheric pressure, and possibly by seasonal to multi-year changes in the steric height and dynamic topography. It is assumed here, however, that on the whole this orbit adjustment procedure is sufficient to eliminate any apparent long-term trend that might otherwise appear as an artifact of time-dependent orbit error.

DATA

The Seasat and Geosat ERM altimeter data from the Greenland ice sheet were corrected for tracking errors, atmospheric effects, and solid earth tides as described by Martin et al. [1983] and Zwally et al. [1983]. These corrections were carried out prior to the orbit adjustment procedure described above. Data from the region of interest near the western margin of the ice sheet were selected using the mask shown in Figure 1. The lower elevation limit for acceptable data was specified, as a function of latitude, far enough up-glacier from the ice sheet margin to exclude waveforms back-scattered from coastal rocks and nunataks. The upper elevation limit coincides, approximately, with the 2000-m elevation contour.

MEASUREMENT NOISE AND CROSSOVER ERROR

The measurement noise levels in Geosat GM altimetry were estimated and mapped by Lingle et al. [1990] as a function of position throughout the West Greenland ablation area, using semivariogram methods. GM altimetry noise levels were similarly estimated and mapped by Lingle and Brenner [1989] within a series of 100 km x 100 km area blocks extending along the EGIG line [see, e.g., Seckel, 1977] from the ablation area to the central ice divide. These results, which are assumed to be characteristic of the Geosat ERM data (acquired by the same instrument on board the same satellite), indicate that average measurement noise levels near the west margin of the ice sheet, up to about the 2000-m contour, are about 10 m. The standard error of a Geosat ERM crossover difference is taken to be $\sqrt{2}$ (10) m or 14.1 m.

The Seasat and Geosat altimeters were generally similar in design, although improvements were incorporated in the Geosat version [MacArthur et al., 1987]. The Geosat altimeter yielded an instrument-induced along-track noise level of about 3 cm RMS over the oceans, compared to 5 cm RMS for Seasat (for 1 per second data in each case), but additional noise due to oceanographic effects caused both altimeters to have an average noise level, integrated over all frequencies, of about 8 cm RMS [Sailor and LeShack, 1987]. Over the ice sheets, the factors primarily responsible for high noise levels in the data are backscatter from off-nadir undulations and points upslope from nadir, and these factors were the same for both Seasat and Geosat. The Geosat measurement noise levels estimated and mapped by Lingle et al. [1990] and Lingle and Brenner [1989] (for 10 per second data) are therefore assumed to be characteristic of the Seasat altimetry.

The Seasat altimeter yielded less continuous data over the Greenland ice sheet due to a tracker that was less responsive over sloping and undulating surfaces. Consequently most of the valid Seasat–Geosat crossover differences are from the subregion of the area shown in Figure 1 that is above the ablation area, where the ice sheet topography is less rugged. The results of Lingle and Brenner [1989] suggest that in the 1300–2000 m elevation range, the average altimetry noise level is roughly 7 m. The standard error of a Seasat–Geosat ERM crossover difference is taken to be $\sqrt{2}$ (7) m or 9.9 m.

SEASONAL ELEVATION CHANGES, 1987–1988

Seasonal mean changes are first computed during the Geosat ERM by dividing the ERM into 91-day "seasons," with late summer defined to coincide with the Seasat time frame (July 10 through October 9). This definition fixes the times of the other seasons. The first season having data of sufficient quality to yield enough crossover differences for statistically valid averaging is late winter 1986-87 (i.e., January 8 through April 9, 1987). Subsequent seasonal mean changes in the surface height are computed with respect to that season. The method, which is described in detail by Lingle et al. [1990, pp. 160–161], is an adaptation of the methods of crossover analysis developed by Zwally et al. [1989]. The method includes cancellation of the ascending versus descending orbit bias [see Lingle et al., 1990, equation (6)]. The method used here differs from that described by Lingle et al. [1990], however, in that each crossover difference is not weighted in proportion to the inverse square of the measurement noise level in the local neighborhood of the crossover point. Rather, a representative noise level for the entire region shown in Figure 1 is estimated, as described above, and unweighted averaging of the crossover differences is employed. An edit level of 15 m is used to define acceptable Geosat ERM–Geosat ERM crossover differences.

The seasonal mean fluctuations in the surface height during 1987-88 (the Geosat ERM) are shown in Figure 2, starting in late winter 1986-87. The standard errors of the mean ranged from ±0.3 to ±0.5 m for the computed changes relative to late winter 1986-87. The error limits shown in Figure 2 are larger than that, however, because the larger errors associated with the crossover differences between Seasat and the Geosat ERM are taken into account, as described in the next section.

Figure 2 shows that during the Geosat ERM the surface was highest on average during late winter, which includes part of spring, the time of maximum snow depth. In both 1987 and 1988 the surface was slightly lower on average during late spring, which includes part of summer, when melting was presumably in progress. During both years the surface height was lowest during late summer, when maximum surface lowering due to ablation is expected. The mean decrease in the surface height from late winter to late

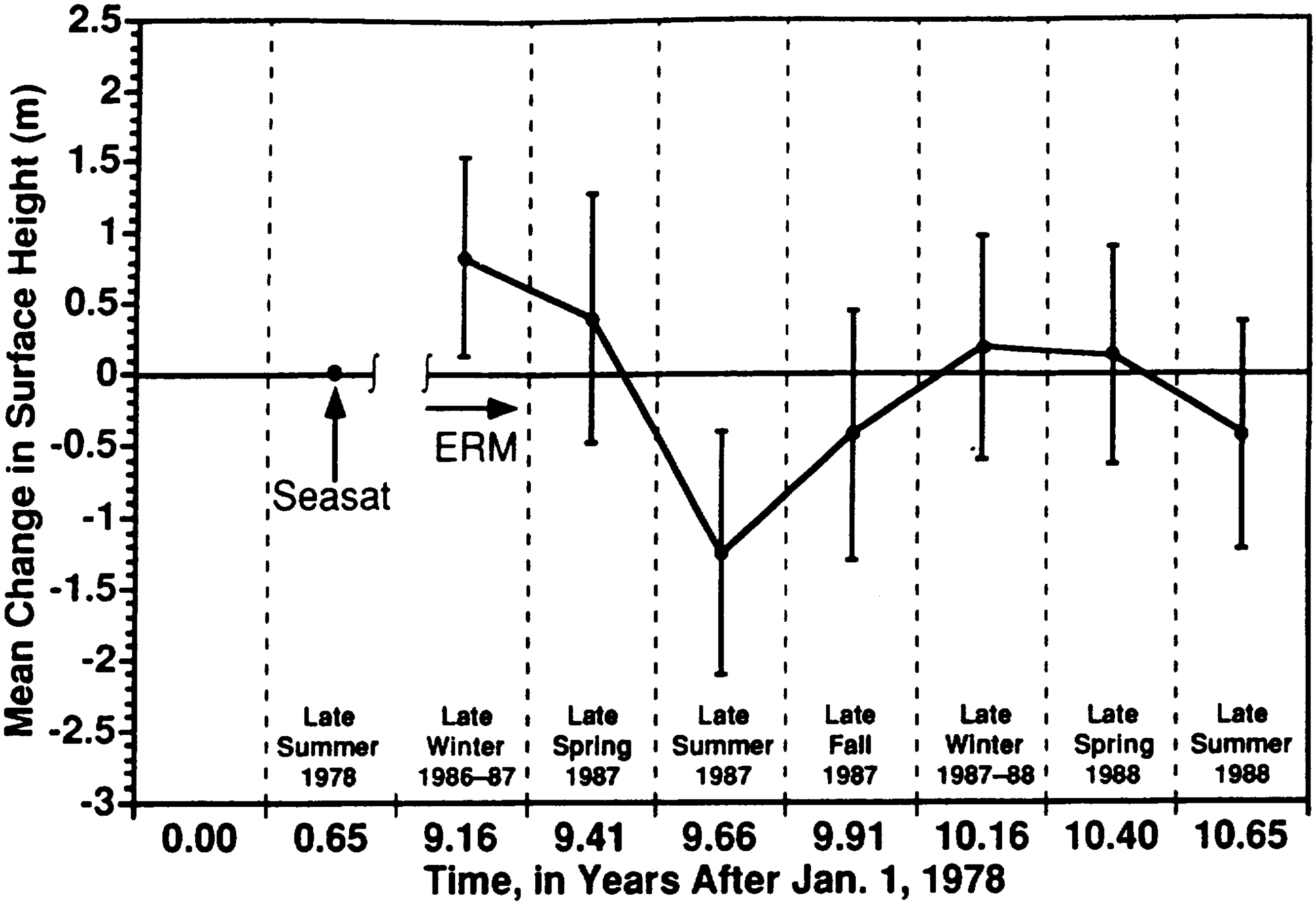

Figure 2. Mean changes in surface elevation near the west margin of the Greenland ice sheet, within the area shown in Figure 1, between late summer 1978 (Seasat) and 1987-88 (the Geosat ERM). Changes were computed using orbits adjusted into the Seasat/Geos-3 mean ocean surface. This figure corresponds to case 1 in Table 1. (Cases 1–4 in Table 1 are used to derive the mean height changes quoted in the abstract and conclusions, as described in the text.) Each error bar represents one standard deviation of the mean for the change computed with respect to the Seasat surface, which is the datum.

summer in 1987 was 2.1 ± 0.7 m. Between late winter and late summer in 1988, the mean decrease in the surface height was 0.6 ± 0.5 m.

MULTI-YEAR ELEVATION CHANGES, 1978 TO 1988

Multi-year mean changes in elevation near the west margin of the Greenland ice sheet are computed by defining the Seasat time frame, late summer 1978, as the initial season. Mean elevation changes between Seasat and the successive seasons of the Geosat ERM were first computed as described by Lingle et al. [1990], with the difference that there is an 8.5-year time gap between late summer 1978 and late winter 1987 (the first season of usable ERM data). Otherwise the method used was as described above, with unweighted averaging of crossover differences employed and a representative standard error assigned to all crossover differences.

The results showed a seasonal cycle during the Geosat ERM relative to late summer 1978, but were less than satisfactory because the discontinuous nature of the Seasat data, combined with data of varying spatial continuity during the Geosat ERM, resulted in two seasons (late fall 1987 and late winter 1987-88) having insufficient valid crossover differences for a statistically meaningful measurement of mean elevation change. (An edit level of 30 m is used to define acceptable Seasat–Geosat ERM crossover differences.) These results are not shown.

An alternative method is adopted instead, consisting of computation of the mean elevation change between Seasat and the first 91-day season of the Geosat ERM having useable data (late winter 1986-87). Subsequent elevation changes are determined by "adding-on" the time series of subsequent seasonal changes during the Geosat ERM, determined as described above. This method does not decrease the standard error of the mean for the Geosat ERM seasonal changes computed relative to the surface during the Seasat time frame, because the error of each change relative to the initial Geosat ERM season is combined with the additional (larger) error of the computed change between the surface during the Seasat time frame and the surface during the initial season of the Geosat ERM. This approach does, however, give a clearer and more continuous picture of the mean changes in the surface height during the Geosat ERM relative to the surface during the earlier Seasat time frame. The results are shown in Figure 2.

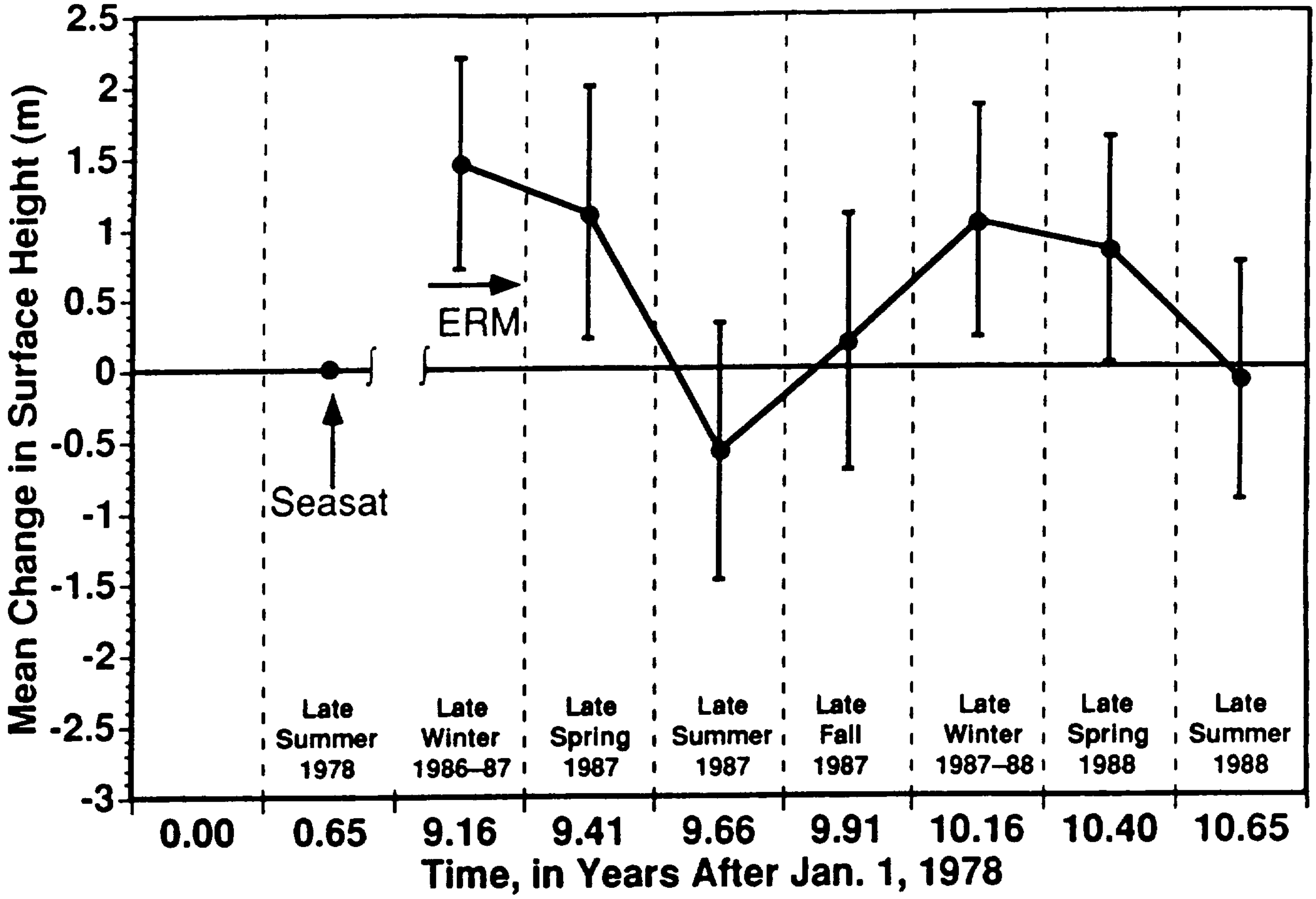

Figure 3. Control case: mean changes in surface elevation near the west margin of the Greenland ice sheet, within the area shown in Figure 1, computed using orbits *not* adjusted into the Seasat/Geos-3 mean ocean surface. The computation is otherwise identical to that shown in Figure 2. These results are analogous to case 1 in Table 1. (The mean height changes quoted in the abstract and conclusions are computed using adjusted orbits, and are derived from cases 1–4 in Table 1, as described in the text.)

CONTROL CASE: MULTI-YEAR ELEVATION CHANGES, 1978 TO 1988, COMPUTED WITHOUT ORBIT ADJUSTMENTS

Figure 3 shows the results of the crossover analysis described above, carried out using altimetry referenced to GEM T2 orbits that were not adjusted into the Seasat/Geos-3 mean ocean surface. Comparison of Figure 3 to Figure 2 shows that the effect of orbit adjustment is to cause a slight downward shift of the curve showing seasonal mean changes in the surface height during the Geosat ERM, relative to the Seasat surface. Otherwise, the curve showing seasonal mean changes during the Geosat ERM is quite similar. Figure 3 suggests that there may have been a negative mean change in the surface height between late summer 1978 and the same season in 1987, but the standard error of the mean overlaps zero. The mean change in the surface height between late summer 1978 and the same season in 1988 is near zero (in Figure 3).

ALTERNATIVE COMPUTATIONS OF LATE SUMMER CHANGES

In this paper, the primary focus is on determination of mean changes in the surface height between late summer 1978 and the same season in 1987 and 1988. This change can be determined in four ways, referred to below as cases 1 through 4, by making use of the different Geosat ERM seasons having sufficient crossovers with the Seasat orbits for valid averaging. Seasat and Geosat ERM orbits that were adjusted into the Seasat/Geos-3 mean ocean surface are employed in all of these cases. Case 1 is described above. In case 2, the Seasat orbits are crossed with the Geosat ERM orbits of late spring 1987, in order to fix the curve showing the ERM seasonal changes (analogous to Figure 2) relative to the mean surface height during the Seasat time frame (the datum in Figure 2). The heights of the late summer 1987 and 1988 surfaces are then calculated relative to late spring 1987, and corrected for the height change between late summer 1978 and late spring 1987.

In case 3, the Seasat orbits are crossed directly with the Geosat ERM orbits of late summer 1987, and with the ERM orbits of late summer 1988. In case 4, the Seasat orbits are crossed with the Geosat ERM orbits of late spring 1988, in order to fix the curve showing the ERM seasonal changes (also analogous to Figure 2) relative to the mean surface height during the Seasat time frame (the datum in Figure 2). The remainder of the calculation is analogous to case 2. The results of the 4 cases are shown in Table 1.

Table 1 shows that although the four cases yield differing results, the error bars overlap. The results are thus self-consistent to within one standard deviation of the mean. The mean change in the surface height between late summer 1978 and late summer 1987 is taken as the unbiased

Case	Late Summer 1978 to Late Summer 1987 (m)	Late Summer 1978 to Late Summer 1988 (m)
1	-1.26 ± 0.85	-0.43 ± 0.80
2	-1.16 ± 1.02	-0.33 ± 0.98
3	0.14 ± 0.78	-1.66 ± 1.08
4	0.24 ± 0.87	1.07 ± 0.82

Table 1. Alternative calculations of mean changes in surface height, late summer to late summer (adjusted orbits).

weighted average of the four cases, with each case weighted in proportion to the inverse square of its standard error. The result is -0.43 ± 0.44 m. The mean change in the surface height between late summer 1978 and late summer 1988, determined in the same way, is -0.18 ± 0.45 m.

DISCUSSION: VARIABLE PENETRATION DEPTH

Ridley and Partington [1988] found that the effect of surface and volume scattering can cause error in the measured range to the snow surface of as much as 1.1 to 3.3 m, depending on the ratio of surface to volume scattering, if the return wave forms are retracked using an assumption that the snow surface lies at the location on the leading slope of half the peak power. That is, the apparent surface measured by the altimeter is lower by an amount that depends on the predominance of volume scattering. These authors also state [Ridley and Partington, 1988, p. 621] that "the error is much smaller if the Martin et al. (1983) method of retracking is used, as the function fitted to the return is sensitive to the inflection point."

The method of Martin et al. [1983] was used to retrack the altimeter waveforms used to derive the elevation measurements employed in this study, as noted in the section on data.

Suppose, however, that surface scattering predominates in summer, when melting and refreezing tend to result in formation of a surface crust on snow, and in exposed bare ice in the ablation area, and that volume scattering predominates is winter, when lower temperatures tend to result in a dry snow surface. Then the seasonal mean changes in the surface height measured by the altimeter would be minimized, since the penetration depth would be maximum during late winter and spring, when the surface is highest, and minimal during late summer when the surface is lowest. The measured seasonal mean changes shown in Figures 2 and 3 and in Lingle et al. [1990] are not overly small, however, relative to the changes one would expect from the ablation rates measured by Braithwaite and Olesen [1989], and the accumulation rates mapped by Radok et al. [1982] (with the exception, perhaps, of the measured mean change between late spring 1988 and late summer 1988, shown in Figure 2). The results of this study and of Lingle et al. [1990] suggest, therefore, that seasonal changes in penetration depth do not preclude measurement of seasonal mean changes in the surface height with altimetry, when the crossover differences are averaged over a suitably large region. The central focus of this study—measurement of mean changes between 1978 and 1987-88—should be unaffected by seasonal variations in penetration depth, in any case, because the mean changes in surface height are measured between the same season (late summer) in each year.

COMPARISON TO FIELD MEASUREMENTS AND OTHER WORK

The multi-year mean changes in the surface height computed in this study can be compared to mean changes in the surface height measured by optical leveling on the lower EGIG line within the ablation area at about Lat 69.5°N. Bauer et al. [1968] found the surface lowered at a mean rate of 0.3 m yr^{-1} between 1948 and 1959. Seckel [1977] found the surface lowered at a mean rate of 0.24 m yr^{-1} between 1959 and 1968. (These results are also summarized by Reeh [1985].) Over an area including the lower EGIG line and measuring about 22 km parallel to the ice sheet margin by 6 km perpendicular to the margin, Thomson et al. [1986] measured a mean surface lowering of 14 m between 1959 and 1982 using photogrammetric methods, which is equivalent to a mean lowering rate of 0.61 m yr^{-1}. A 0.24–0.61 m yr^{-1} mean rate of surface lowering during the 9–10 year measurement period considered here would be significantly greater than the error range shown in Figures 2 and 3, and should thus be recognizable in the altimeter data. If the ablation area of West Greenland is thinning, however, the rate of thinning should decrease upglacier from the margin because of the parabolic nature of ice sheet profiles [Paterson, 1981]. The EGIG optical leveling results, in fact, show thickening of the inland ice sheet above the equilibrium line [Reeh, 1985]. The results of this study are determined using altimeter crossover differences from both below and above the equilibrium line, with most being from higher elevations (up to about 2000 m). Thus, if the ice sheet is still thinning close to the western margin, while thickening above the equilibrium line, as determined by earlier field workers, this may be consistent with the near-zero mean change determined in this study using crossover differences averaged over the whole area (Figure 1).

If there was a linear mean decrease of the surface height between late summer 1978 and the same season in 1987, the results obtained here indicate the rate of surface lowering would have been 0.05 ± 0.05 m yr^{-1}. Between late summer 1978 and the same season in 1988, however, the assumption of a linear mean decrease leads to a conclusion that the surface lowered at a mean rate of 0.02 ± 0.05 m yr^{-1}. The error bars for these two cases overlap, but the difference, and the seasonal mean changes shown in Figures 2 and 3, indicate that the interannual variability is significant and a uniform rate of surface lowering during the 9 to 10 year measurement period is unlikely.

The seasonal mean elevation changes in the West Greenland ablation area measured by Lingle et al. [1990] suggest an increasing trend for the surface height during the Geosat GM. Zwally et al. [1989] and Zwally [1989] measured an increased surface height in all elevation intervals between Seasat and the Geosat GM (1978 to 1985-86), as well as during the Geosat GM, including the 700–1200-m interval, which generally coincides with the ablation area. However, the results obtained in this study (Figures 2 and 3) suggest, if anything, a decreasing trend for the surface height superimposed on the seasonal mean changes during the Geosat ERM, while the Seasat to Geosat ERM measurements show no statistically significant mean change in the surface height between late summer 1978 and the same season in 1987 and 1988, throughout the area shown in Figure 1. These apparent discrepancies may be related to short-term variability.

CONCLUSIONS

The results of this analysis indicate that the surface of the Greenland ice sheet near the western margin, throughout the region including the ablation area and extending up to about the 2000-m elevation contour and to Lat. 72°N (Figure 1), was lower by 0.4 ± 0.4 m on average in late summer 1987 than in late summer 1978. The surface was lower by 0.2 ± 0.5 m on average in late summer 1988 than in late summer 1978. This result was obtained from unbiased weighted averaging of the four sets of results shown in Table 1, of which case 1 (only) is shown in Figure 2. The possible effects of time-dependent orbit error were minimized by adjusting the GEM T2 Seasat and Geosat ERM orbits into the Seasat/Geos-3 mean ocean surface [Marsh et al., 1986, 1990].

The mean changes in elevation between late summer 1978 and the same season in 1987 and 1988 were also computed using altimetry referenced to GEM T2 orbits that were not adjusted into the Seasat/Geos-3 mean ocean surface. The results, shown in Figure 3, are analogous to Figure 2 and case 1 in Table 1. These results also indicate no statistically significant mean change in the surface height during the 9–10 year measurement interval.

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