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Growth of Greenland Ice Sheet: Measurement

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Measurements of ice-sheet elevation change by satellite altimetry show that the Greenland surface elevation south of 72° north latitude is increasing. The vertical velocity of the surface is 0.20 ± 0.06 meters per year from measured changes in surface elevations at 5906 intersections between Geosat paths in 1985 and Seasat in 1978, and 0.28 ± 0.02 meters per year from 256,694 intersections of Geosat paths during a 548-day period of 1985 to 1986.

DETERMINATION OF THE BALANCE between mass input and outflow of the polar ice sheets is needed for understanding of the ice-sheet response to climate change and the contributions to sea-level rise or fall. Measurement of elevation change by satellite altimetry offers a method of determining changes in ice volume and therefore mass balance (*1*). The 3-year operation of GEOS-3 radar altimeter from April 1975 to June 1978 (*2*), followed by the 3-month operation of the Seasat radar altimeter from July to September 1978, provided a time series of ice elevations, but the precision and spatial coverage of GEOS-3 was limited. The U.S. Navy Geosat radar altimeter (*3*), which was launched in March 1985, has provided a large number of recent repetitive measurements. We have determined changes in ice-surface elevations using data from GEOS-3, Seasat, and the first 18 months of Geosat. The estimated change in ice volume and its significance is discussed in a companion paper (*4*).

Changes in surface elevation were determined where successive sub-satellite paths intersect (Fig. 1). The measured elevation difference at a crossover point is $dH = H_2 - H_1 + E$, where H_2 and H_1 are the surface elevations during successive orbits at times t_2 and t_1, respectively, and E is the random measurement error from a distribution with a SD. The error for each elevation measurement is $E/\sqrt{2}$, which includes errors in the altimeter-range measurement and in determination of the vertical position of the orbit. The magnitude of E is determined from analysis for which $(H_2 - H_1)$ are small. Although E is usually larger than actual elevation changes, average changes can be obtained over areas of the ice sheet for time periods in which there are a sufficiently large number of measurements.

The range measured (Fig. 1) is to the average surface elevation in the "pulse-limited" footprint [maximum circular area from which radar reflection is simultaneously received by the altimeter (*2, 5*)]. The minimum pulse-limited footprint is 3.6 km in diameter for the GEOS-3 altimeter and 1.6 km for both the Seasat and Geosat altimeters, over smooth surfaces and larger over rough surfaces. Ranges are obtained at 0.66-km intervals along the satellite tracks; therefore, successive footprints overlap by 40% or more. Surface elevations at the crossover point are obtained by interpolation. Determination of the absolute surface elevation at satellite nadir would require correction for the slope-induced offset of the pulse-limited footprint from nadir (*6*), which is caused by the tendency of the pulse-limited footprint to be located at the closest surface lying within the larger "beam-limited" footprint, which is ten times the size of the pulse-limited footprint (*5*). However, for the purpose of studying elevation changes, correction for slope-induced errors is not necessary because the pulse-limited footprint is usually located at the same place on the surface during successive transits.

We corrected surface elevations for variations in the effective atmospheric path length, earth tides, and lags in the automatic radar-pulse tracking circuitry of the altimeter (*3, 4*). For GEOS-3 and Seasat, residual errors in the radial position of the satellite with respect to the center of the earth are reduced by reference of the orbital positions to a common ocean surface derived from the Seasat and GEOS-3 altimeter data. After orbit adjustment, the SD of the elevation differences is 4.7 m for GEOS-3–GEOS-3 crossovers and 1.0 m for Seasat-Seasat crossovers. The standard errors for single measurements are 3.3 m for GEOS-3 and 0.70 m for Seasat. The calculated SD for GEOS-3–Seasat differences is 3.4 m. Precise orbit information over ice-covered areas is included with the Geosat data (7). The SD of the elevation differences at 16,250 Geosat-Geosat crossovers for which the time difference between measurements is <15 days is 1.49 m. The Geosat single-measurement error is therefore 1.05 m. In these analyses, crossover differences greater than 10 m were discarded (15 m for GEOS-3) (*8*). The relative SD for Geosat-Seasat differences is therefore 1.26 m. The remaining errors are mainly a combination of altimeter measurement error over irregular surfaces and residual orbit errors.

Two methods were used to obtain the rate of change of surface elevation from a set of

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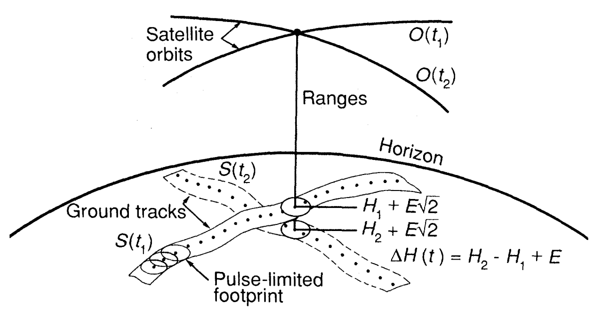

Fig. 1. Crossover method for measuring changes in surface elevation, $S(t)$, from radar-altimeter measured elevations, $H(t)$, on successive orbital paths (O) of the satellite. Horizontal location of the crossover point is determined within a few meters. The relative error, E, for measurement of elevation change, dH, at a single crossover is about 1.4 m.

Table 1. Average rates of change in ice-sheet elevation and relative error at a single crossover of the satellite radar altimeters. N, number of crossovers.

Satellite	N	dH/dt (m/year)	Maximum dt (years)	Interval	Relative error (m)
GEOS-3–Seasat	657	0.11 ± 0.14	3.2	April 1975 to June 1978	3.4
Geosat-Seasat	5,906	0.20 ± 0.06	7.0	July 1978 to October 1985	1.26
Geosat-Geosat	256,694	0.28 ± 0.02	1.5	April 1985 to September 1986	1.49

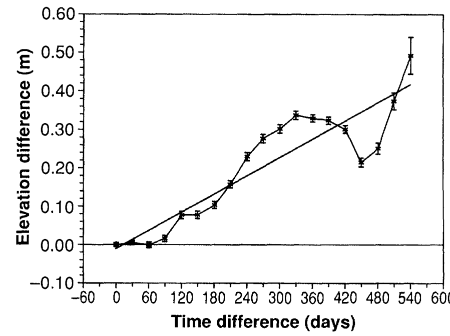

Fig. 3. Average dH_i of Fig. 2 in 30-day intervals versus dt (see text). Nonlinear components caused by seasonal effects are smaller than the linear trend of 0.28 m/year.

crossover measurements. If a sufficient number of measurements were made during two distinct periods separated by a relatively large time interval (dt), then the surface velocity is the average crossover height difference divided by the time interval $[\Sigma(H_2 - H_1)_i/N)/dt$, where $(H_2 - H_1)_i$ is the elevation difference at the ith crossover and N is the number of crossovers]. This method is appropriate for comparing the 3 months of Seasat measurements with Geosat measurements made 7 years later.

The second method, the dH/dt method, is appropriate for a set of crossovers that tend to have randomly distributed time intervals. The slope of a linear fit to the crossover differences, $dH_i = (H_2 - H_1)_i$, versus their time intervals, $dt_i = (t_2 - t_1)_i$, gives the thickening ($dH/dt > 0$) or thinning rate ($dH/dt < 0$). In this method, we implicitly assume that dH/dt is a linear function of time and that the data are randomly distributed about a linear trend. The SE of the slope and the SD of the points about the linear fit can only be used to assess the statistical significance of a linear trend if cyclical components in the data are small.

The dH/dt method has several advantages. All crossovers created by a series of altimeter measurements can be used. The effect of possible seasonal changes in either the surface elevation or radar back-scattering properties is reduced, because a specific dt_i, for example, might have summer-to-winter and winter-to-summer measurements. Also, the value of the dH intercept at $dt = 0$ can indicate measurement biases.

We first applied the dH/dt method to GEOS-3 and Seasat data, which extended to 65.1°N and 72.07°N, respectively. The average rate of increase in surface elevation south of 65.1°N obtained from 525 crossover differences was 0.35 ± 0.17 m/year (*9*). Subsequent analysis with 127 more crossovers gave a dH/dt of 0.11 ± 0.14 m/year (Table 1). The SD of the 657 dH_i values with respect to the linear fit is 3.48 m, consistent with the 3.40-m estimated relative error (Table 1). A similar analysis of 626 GEOS-3–GEOS-3 crossovers gave a dH/dt of 0.07 ± 0.23 m/year. Although these three values of dH/dt are positive and overlap within 1 SD, the confidence regarding ice-sheet thickening based on these data alone is not high.

The dH/dt analysis for 548 days of Geosat measurements between 1 April 1985 and 30 September 1986 (Fig. 2) gives 0.284 ± 0.004 m/year for 256,694 crossover differences. The ice sheet area covered is from the southern tip of Greenland to 72°N. The SD of the points about the line is 1.14 m, for an iterative 3 SD editing with SD convergent to 2%. In the calculation of SE of the slope, we assume that errors among the dH_i are independent, even though some correlation may be expected among the crossovers along a given orbit. Depending upon the magnitude of the residual orbit error compared to the altimeter measurement error, the number of crossovers per orbit, and the degree of correlation from one crossover to another, the true error on the slope for this case could be as much ± 0.02 m/year.

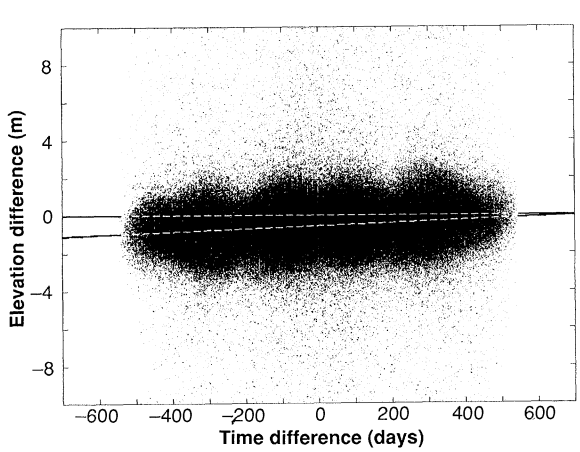

Fig. 2. Elevation differences dH_i at 256,694 Geosat-Geosat crossovers versus dt_i. Negative dt_i indicate that the elevation measurement, H_2, on an orbit ascending in latitude preceded the elevation measurement, H_1, on an orbit descending in latitude. Intercept at $dt = 0$ shows an ascending-descending orbit bias of −0.48 m. The rate of increase in surface elevation is 0.28 ± 0.02 m/year.

In our analysis, $dH = H_a - H_d$, where H_a is the elevation obtained on the orbit path ascending in latitude and H_d is the elevation on the path descending in latitude, regardless of whether $t_a > t_d$ or $t_a < t_d$. The dH intercept at $dt = 0$, obtained by averaging all dH_i for which $|dt_i| < 15$ days, gives an ascending-descending orbit bias equal to −0.48 ± 0.01 m. This bias indicates that Geosat descending orbit calculations are systematically too high relative to ascending orbits in the vicinity of Greenland.

Although the dH/dt method reduces the effect of possible seasonal variations in the measured elevation, a seasonal modulation in the deviation of the dH_i values about the linear fit is evident in Fig. 2. Minimal deviations appear at about 0, 6, and 12 months, and maximal deviations at about 3 and 9 months. To examine the effect of this semiannual variation on the linear trend analysis, we averaged the dH_i in 30-day intervals $[dH_m = \Sigma(H_2 - H_1)_i/n$, where n is the number of measurements in the interval]. The dH_m for $dt < 0$ were then averaged with the corresponding values for $dt > 0$ (Fig. 3). Although there are obvious nonlinear components, the variations about the linear fit are not large and are nearly symmetrical. Linear fits to the mean values give 0.289 ± 0.032 m/year and 0.279 ± 0.020 m/year, before and after averaging the mean values for $dt < 0$ and $dt > 0$, which are consistent with the slope in Fig. 2.

Elevation changes between Geosat and Seasat measurements are more meaningful for mass balance studies because of the 7-year interval between the two satellites; however, comparisons may be influenced by differences in the orbit calculations, or their relative ocean-geoid levels. Coarse-grid information received from the Navy on their altimeter measurements of the ocean surface in the vicinity of Greenland with Geosat is sufficient, however, to estimate the possible bias with respect to the Seasat measurements. The estimated Geosat-Seasat elevation bias over Greenland is 0.4 ± 0.4 m, which we treat as a correction with a systematic error. In other respects, the Geosat and Seasat altimeters are similar in design, and the same range-correction retracking algorithm was used over ice. We accounted for the ascending-descending orbit bias by analyzing the crossovers of ascending Geosat orbits with Seasat separately from those with descending Geosat orbits, and then averaging the two results. We avoided seasonal biases by comparing the Seasat data for 15 July to 10 October 1978 with Geosat data for the same period of 1985. The resulting Geosat-Seasat average elevation difference for 5906 crossovers is 1.785 ± 0.014 m. After correction for the Geosat-Seasat orbit bias, it is 1.385 ± 0.414 m. The average rate of change over the 7-year interval at these crossover locations is thus 0.20 ± 0.06 m/year. The altimeter measurements (Table I) thus show that the southern Greenland ice sheet has been thickening since the mid-1970s.

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8. The distribution of most of the crossover elevation differences is Gaussian; however, some differences greater than 3 SD of the Gaussian distribution are believed to be caused by measurement of ranges to different locations on the surface on successive orbits. In some cases along a single orbit, two ranges to different places on the surface, differing in elevation by as much as 20 m, are indicated by simultaneous, double-peaked radar returns (*5*). To eliminate such cases, crossover elevation differences greater than 3 SD of the primary Gaussian distribution were discarded in the analysis of elevation changes.
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Growth of Greenland Ice Sheet: Interpretation

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An observed 0.23 m/year thickening of the Greenland ice sheet indicates a 25% to 45% excess ice accumulation over the amount required to balance the outward ice flow. The implied global sea-level depletion is 0.2 to 0.4 mm/year, depending on whether the thickening is only recent (5 to 10 years) or longer term (<100 years). If there is a similar imbalance in the northern 60% of the ice-sheet area, the depletion is 0.35 to 0.7 mm/year. Increasing ice thickness suggests that the precipitation is higher than the long-term average; higher precipitation may be a characteristic of warmer climates in polar regions.

THE MASS BALANCE OF THE GREENland and Antarctic ice sheets is of current interest, largely because of its direct relation to global sea level, which appears to be rising by 2.4 ± 0.9 mm/year (*1*). Although both thermal expansion of the ocean (*2*) and melting of small glaciers (*3*) contribute to sea-level rise, the major source of water is undetermined. Also, the possibility of enhanced ice-sheet melting in a warmer climate (*4*) is of concern. Glaciers respond to both precipitation and temperature in such a manner, however, that enhanced precipitation may offset increases in surface melting (*5*).

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Each year, approximately 3000 km^3 of water is exchanged between the ocean and the ice sheets of Greenland and Antarctica, a volume equivalent to 8 mm of water from the entire surface of the world's oceans. The uncertainty in ice-sheet mass balance has been at least ±30% of the annual mass exchange (*6*); this uncertainty is equivalent to ±2.4 mm/year of sea-level change. Recently, Meier *et al.* (*7*) estimated that there has been a small positive balance for both Greenland (−0.1 ± 0.4 mm/year of sea-level change) and Antarctica (−0.6 ± 0.6 mm/year). In contrast, recent total flux estimates (*8*) of annual snow accumulation, iceberg discharge, and peripheral melting of the Antarctic ice sheet indicate that the net ice loss has been 750 km^3/year, which is 35% of the mass input and equivalent to 1.9 mm/year of sea-level rise.

Satellite radar altimetry measurements show that the surface elevation of the Greenland ice sheet south of 72°N (Fig. 1) increased from 1978 to 1986 (*9*). The measured elevation change varies with latitude (Fig. 2), and the errors are larger at lower latitudes and lower ice-sheet elevations mainly because of the smaller number of crossovers (*10*). The largest elevation increases were over the southern dome around 63.5°N and in the central region near 72°N during 1985 to 1986.

The spatially averaged elevation changes, obtained by analyzing the crossover differences in ice-sheet elevation bands (Fig. 3) and weighting those values by the fractional area in each band (0.12, 0.14, 0.20, 0.31, and 0.23 for lower to higher elevations), are 0.233 ± 0.041 m/year for 1978–1985 Geosat-Seasat measurements and 0.239 ± 0.030 m/year for 1985–1986 Geosat-Geosat measurements. In southern Greenland, the equilibrium line (boundary between net ablation and net accumulation) is at ~1200 to 1500

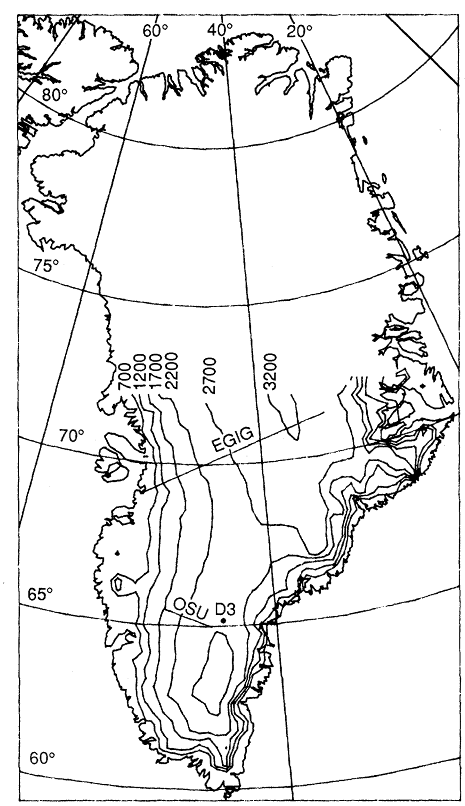

Fig. 1. Map of Greenland showing surface elevations in region covered by satellite radar altimetry and locations of surface measurements (EGIG, D3, and OSU) of elevation change.