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Contribution of Alaskan glaciers to sea-level rise derived from satellite imagery

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Over the past 50 years, retreating glaciers and ice caps contributed 0.5 mm yr^{-1} to sea-level rise[1], and one third of this contribution is believed to come from ice masses bordering the Gulf of Alaska[2,3]. However, these estimates of ice loss in Alaska are based on measurements of a limited number of glaciers that are extrapolated to constrain ice wastage in the many thousands of others. Uncertainties in these estimates arise, for example, from the complex pattern of decadal elevation changes at the scale of individual glaciers and mountain ranges[4–7]. Here we combine a comprehensive glacier inventory with elevation changes derived from sequential digital elevation models. We find that between 1962 and 2006, Alaskan glaciers lost 41.9 ± 8.6 km^3 yr^{-1} of water, and contributed 0.12 ± 0.02 mm yr^{-1} to sea-level rise, 34% less than estimated earlier[2,3]. Reasons for our lower values include the higher spatial resolution of our glacier inventory as well as the reduction of ice thinning underneath debris and at the glacier margins, which were not resolved in earlier work. We suggest that estimates of mass loss from glaciers and ice caps in other mountain regions could be subject to similar revisions.

The extent and high turnover rates of glaciers in Alaska and northwest Canada (Fig. 1), hereafter 'Alaskan glaciers', make them a potentially important contributor to historical and future sea-level rise (SLR). With the exception of a few tidewater glaciers, most have retreated since the late nineteenth century[8]. Previous efforts to estimate their mass loss since the 1950s have relied on extrapolating site-specific measurements to the entire region[2,3,9]. In their landmark study, Arendt *et al.*[2], for example, used laser altimetry to measure elevation change on 67 glaciers covering 20% of the area of Alaskan glaciers.

We apply sequential digital elevation model (DEM) analysis[10] to estimate the mass loss of Alaskan glaciers over the period 1962–2006. A digital glacier inventory, created by merging glacier outlines derived from US and Canadian maps (see the Methods section), indicates that the total ice-covered area for the mid-to-late twentieth century is about 87,860 km^2 (Table 1). This estimate is higher than the total area (85,150 km^2) reported by Dyurgerov and Meier[9] and in the gridded inventory of Cogley[11] (80,430 km^2), but slightly lower than the 90,000 km^2 value used by Arendt *et al.*[2]. Our inventory, compiled from a variety of sources with different accuracies and different dates, better resolves small glaciers and emerging rock outcrops on large glaciers and icefields. For those few glaciers that advanced between the median date of the maps (1962) and of the satellite images (2006), the inventory was updated using 5 m (Spot5) and 15 m (ASTER) resolution satellite images.

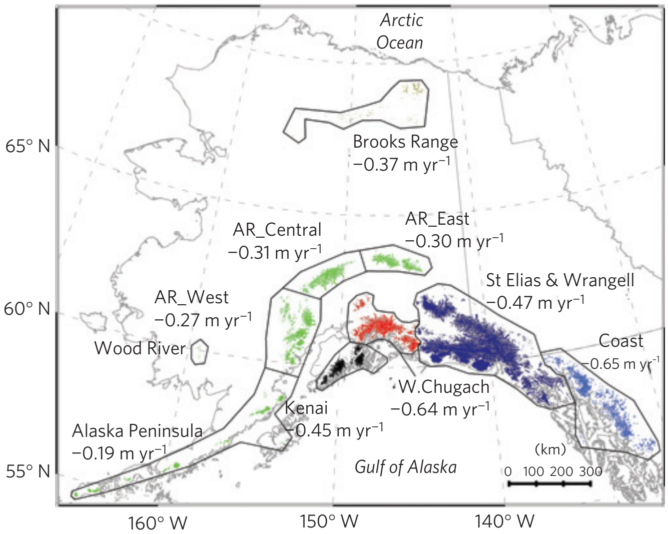

Figure 1 | Regional area-average glacier mass balance in northwest North America between 1962 and 2006. The boundaries and names of the different glaciated regions follow those of Arendt *et al.*[2], except that the Wrangell Mountains have been included in the St Elias Mountains. AR: Alaska Range.

We calculated ice elevation changes for nearly three quarters of the ice-covered areas in Alaska by subtracting an old DEM derived from map elevation contour lines from a recent DEM derived from Spot5 and ASTER images. To reduce systematic errors in our estimate of ice elevation changes, the two topographic data sets have each been adjusted to ICESat altimetric profiles (see the Methods section). Although random elevation errors are relatively high in the map (±45 m in the accumulation area[2]) and in the satellite DEMs (±15 m for ASTER (ref. 12), ±10 m for Spot5 (ref. 13)), they are reduced by averaging over vast regions (see Supplementary Notes).

The complexity of glacier wastage during the period 1962–2006 is illustrated by ice loss in the Western Chugach Mountains (Fig. 2). Most glaciers thinned, especially their low-elevation tongues (Fig. 3). A few glaciers, such as the tidewater Harvard Glacier thickened and advanced[14]. Columbia Glacier alone accounts for 42% of the ice loss in this mountain range. For this tidewater glacier, maximum thinning rates averaged 10 m yr^{-1} during 1957–2007, but thinning accelerated after 1980, with the onset of rapid, frontal retreat[15]. Maps of ice elevation change for other Alaskan regions reveal the pattern of glacier changes over the past 50 years

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Table 1 | Ice loss and mass balance in the different mountain ranges of Alaska

Region	Ice-covered area (km^2)	Surveyed area (%)*	Map date	Satellite date	Ice loss (km^3 yr^{-1} w.e.)	Area average mass balance (m yr^{-1} w.e.)
Brooks Range	598	1	1956	2002	0.18±0.05	−0.37±0.06
Alaska Range	15,834	67	1953	2004	4.33±1.4	−0.30±0.09
Kenai Peninsula	4,351	68	1950	2007	1.95±0.47	−0.45±0.11
Western Chugach Mountains	9,149	82	1954	2006	5.81±0.66	−0.64±0.07
St Elias and Wrangell Mountains	45,905	75	1968	2006	21.66±4.4	−0.47±0.09
Coast Mountains	12,026	72	1966	2007	7.88±1.6	−0.65±0.14
Alaskan Glaciers	87,862	73	1962	2006	41.9±8.6	−0.48±0.10

*The fraction of the ice-covered area where elevation changes were measured using sequential DEMs.

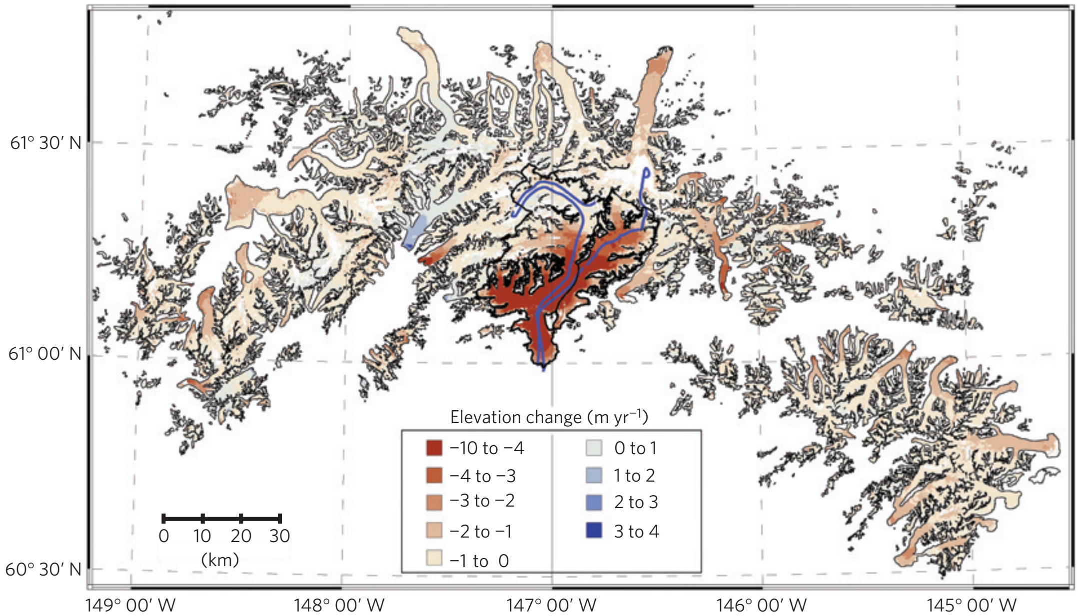

Figure 2 | Map of surface elevation change in the Western Chugach Mountains between the 1950s and 2007. The thin black line corresponds to our new ice inventory. The thick black line is the outline of Columbia Glacier. The locations of laser altimetry profiles used by Arendt *et al.*[2] to estimate the ice loss of Columbia Glacier are shown in blue. Regions where no reliable elevation changes could be measured are denoted in white.

and their distribution with altitude (Supplementary Figs S1,S2). The heterogeneous ice elevation changes within each mountain range result from differences in glacier dynamics (many glaciers are surge-type, lake-terminating or tidewater) and in climate sensitivity (effect of debris cover, and distribution of ice with altitude)[4,5,14]. The limited thinning or slight thickening at the highest elevations is consistent with the enhanced accumulation observed since 1950 in an ice core drilled at 5,340 m a.s.l. in the St Elias Mountains[16].

Surface elevation changes are converted to mass balances in each glaciated area (see the Methods section). Regional balances are spatially homogeneous in the northern and central parts of Alaska (about −0.3 m yr^{-1} water equivalent (w.e.)) but become more negative in the maritime icefields bordering the Gulf of Alaska (Fig. 1). This increased mass loss is attributed to the higher sensitivity of maritime glaciers to temperature change[17] and the rapid retreat of large, tidewater glaciers[2]. Between 1962 and 2006, Alaskan glaciers lost 41.9 ± 8.6 km^3 yr^{-1} w.e. (Table 1) and contributed 0.12 ± 0.02 mm yr^{-1} to SLR. It corresponds to 7.5% of a recent estimate of SLR (1.6 mm yr^{-1}) during 1961–2003 (ref. 18).

Our estimate of mass loss is supported by the lack of notable elevation bias in our data sets on ice-free terrain (Supplementary Table S2) and the agreement between our geodetically derived and field-based mass balances[19] for Gulkana and Wolverine glaciers although the time intervals differ (Supplementary Table S3).

To compare our results to Arendt *et al.*[2], we time-weighted their ice loss for 1962–1995 and 1995–2006 by assuming that the loss for 2001–2006 equals that measured for 1995–2001 and constructed a single estimate for 1962–2006. The extension for 2001–2006 is consistent with field-based annual mass-balance measurements[19] and with a recent analysis based on GRACE gravity fields[20]. Our estimate is 34% smaller than the 62.7 ± 19.9 km^3 yr^{-1} w.e. ice loss (0.17 ± 0.05 mm yr^{-1} SLR) based on airborne laser altimetry[2]. Uncertainties are large in both our estimate and the one of Arendt *et al.*[2] and arise mainly from the uncertainties of the old contour maps (Supplementary Table S1). As the same maps were used, however, this source of error is shared by both estimates and thus, we constructed an error estimate that applies only to differences in ice loss (Supplementary Notes). Our revised value is 20.8 ± 4.8 km^3 yr^{-1} w.e. lower than the laser-altimetry ice loss.

Regional extrapolation to unsurveyed ice masses is a potentially important source of uncertainty[14]. Unmeasured glaciers represented 80% of the total Alaskan ice-covered area in the laser altimetry study[2] but only 27% in our sequential DEM analysis. Three other factors, taken together, could explain why Arendt *et al.*[2] overestimated the ice loss.

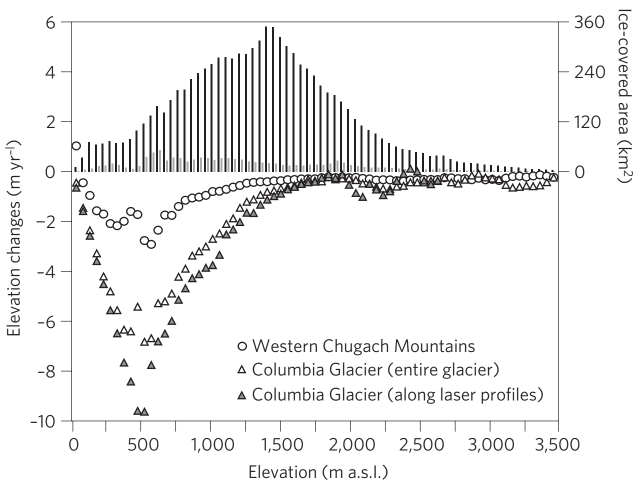

Figure 3 | Hypsometry and rate of ice elevation change versus altitude in the Western Chugach Mountains. Upper panel: The hypsometry (distribution of ice-covered areas with altitude) for the whole mountain range (9,149 km^2) and Columbia Glacier (1,066 km^2). Lower panel: 1957–2007 rate of ice elevation changes (averaged every 50 m elevation bins) extracted from the sequential DEM for the Western Chugach Mountains, the Columbia Glacier and along the altimetric laser profiles (see Fig. 2) surveyed by Arendt *et al.*[2].

First, their glacier inventory had a lower resolution and overestimated the Alaskan ice-covered area by 2%. Second, no correction was made for the insulating effect of debris cover. Thinning rates on debris-covered glaciers differ considerably from rates measured for non-debris covered ice when influenced by similar climate[10]. Our maps of ice elevation changes show that debris-covered glaciers experienced lower thinning rates. At low elevations on Bering Glacier, for example, we observe a twofold reduction of thinning rates under debris compared with debris-free ice. If elevation changes on debris-free ice were assumed to be representative of the whole Bering Glacier, the total ice loss for this glacier would be overestimated by 13%. Many Alaskan glaciers are partly debris-covered and the mass lost from these glaciers will be lower than for non-debris-covered ice. This effect is implicitly included in our sequential DEM analysis.

Third, for individual glaciers, Arendt *et al.*[2] measured elevation changes along two or three profiles, generally following a central flowline, and from this information the glacier-wide changes were estimated. Although seemingly innocent, this method can lead to a systematic overestimation of ice loss. The amount of downwasting cannot exceed the ice thickness, so thinning at the glacier margins is typically lower than that observed along its central flowline[21]. In contrast, sequential DEMs provide a nearly complete coverage of ice elevation changes and allow us to assess the magnitude of errors associated with this flowline sampling bias. For Columbia Glacier, one of the largest contributors to SLR among Alaskan glaciers, thinning is overestimated along laser altimetry profiles below 1500 m a.s.l. and the glacier-wide estimate of ice loss is inflated by 27% (Fig. 3 and Supplementary Fig. S3). Similar comparisons were made for other large glaciers in the Arendt *et al.*[2] data set and, in total, we found that altimetry-simulated ice loss exceeds actual ice loss by 22% (Supplementary Table S4). The magnitude and sign of the bias in the ice loss owing to central flowline sampling varies among the glaciers and depends on how reliably the profiles captured the across-flow variations in elevation change. We find no simple relation in our data that could be used to improve calculations of ice loss based on laser profiling.

We conclude that Arendt *et al.*[2] overestimated ice loss for Alaskan glaciers, an opposite conclusion to that of Larsen *et al.*[4] who also used sequential DEM analysis to find that Arendt *et al.*[2] underestimated the ice loss by more than a factor of two for a 14,500 km^2 ice-covered area in the Coast and southern St Elias mountains. In fact, our results are compatible with Larsen *et al.*[4] because in this area, our ice loss (12.5 ± 2.0 km^3 yr^{-1} w.e.) is close to the Larsen *et al.*[4] value (15 ± 4.0 km^3 yr^{-1} w.e.) and, therefore, remains higher than the ice loss previously reported by Arendt *et al.*[2].

Repeat airborne laser profiling is an efficient means to detect changes in the rate of ice loss along the centreline for a given glacier[22], but new approaches are required to scale these measurements up to an entire glacier and glaciated region. Where such data exist, we advocate the use of sequential DEMs to obtain a comprehensive view of glacier and ice-cap contributions to SLR for other regions that contain sizeable fractions of global ice cover. Such regionally integrated measurements could be compared with results obtained through grid-based mass-balance modelling[23] to resolve the relative role of surface mass balance and tidewater dynamics in the regional ice wastage, and thus better constrain the future glacier and ice-cap contribution to SLR (refs 17, 24). This is crucial given that SLR is one of the main socio-economic hazards associated with global warming[25].

Methods

Glacier inventory. In Alaska, our glacier inventory is based on the Digital Line Graph files that contain all water features digitized from the United State Geological Survey (USGS) 1:63,360-scale, 15 min topographic maps. In southeast Alaska, Digital Line Graph files were missing. Instead, we used the inventory compiled by Beedle[26] for the largest icefields complemented with smaller ice masses extracted from the USGS National Hydrographic Database. For the Yukon Territory, glacier extents were extracted from the 1:50,000-scale Canadian National Topographic Database and in British Columbia the 1:20,000-scale Terrain Resource Information Management database[6]. Minimal manual editing was carried out to correct some obvious errors; otherwise we rely on the capacity of the original cartographers to identify and outline each ice mass.

Map DEMs. DEMs derived from the original map contour lines have been obtained from the USGS for Alaska, Geomatics Yukon for the Yukon Territory and GeoBC for British Columbia. In Alaska and the Yukon Territory, the vertical reference for altitude is the National Geodetic Vertical Datum 1929, which differs from the Earth Geopotential Model 1996 vertical reference used for recent satellite DEMs (ref. 4). To account for this difference, we systematically compared Alaska and Yukon Territory DEMs with ICESat data[27] on the ice-free terrain. On average, we found Alaska and Yukon Territory map elevations to be 2.5 m higher than ICESat data (standard deviation of 20–25 m), in good agreement with the 2.3 m offset calculated by Larsen and colleagues[4]. Thus, 2.5 m has been subtracted from Alaska and Yukon Territory DEMs before comparison with recent satellite data.

Satellite DEMs. Where available, we used a 40 m DEM derived from Spot5-HRS images acquired during the SPIRIT project[13]. Accuracy in glaciated areas is better than ± 10 m (refs 7, 13). Unreliable elevations have been masked using the score channel. No Spot5-HRS DEM is available in the Alaska Range, Alaska Peninsula and in part of the St Elias and Coast mountains. Some of these gaps were filled with ASTER 30 m DEMs calculated using the SILCAST software with an accuracy of ± 15 m (ref. 12). Images as close as possible to the end of the ablation period (mid-September in Alaska) were selected to minimize errors owing to seasonal elevation changes. The acquisition dates of all images are listed in Supplementary Table S5. Both ASTER and Spot5-HRS DEMs are automatically derived from stereo-imagery without ground control points and, thus, may contain some planimetric and altimetric biases[8,16]. These biases have been estimated and corrected using ICESat data acquired closest in time to the acquisition date of the satellite-derived DEM. For each ICESat footprint, the corresponding DEM elevation was extracted by bilinear interpolation. All data points for which the absolute elevation differences were greater than 70 m were considered as outliers. The planimetric shift was corrected by minimizing the standard deviation of these elevation differences[8]. Elevation differences were then plotted as a function of altitude and a least-squares adjustment was used to model the elevation bias[28]. The parameters of this adjustment (α, the slope and β, the vertical offset at sea level) are provided for each satellite DEM in Supplementary Table S5. When the same glaciated area was covered by different DEMs, we chose the one that had the lowest standard deviation of the elevation differences when compared with ICESat data.

Volume changes in unsurveyed areas. Glacier volume change for the 27% of unmeasured glaciated areas was estimated by integrating the measured elevation changes over the altitude distribution of unmeasured areas in each mountain range. This value was added to the measured changes to obtain a total volume change in each region. For the Brooks Range and most of the Aleutian Islands and Alaska Peninsula (Fig. 1), the USGS maps have a poor geodetic control[29] and the coverage using satellite data is limited, so that it was not possible to reliably measure elevation changes. In the Brooks Range, we use the specific mass balance of McCall Glacier[30] and following Rabus and Echelmeyer[29], assumed that this glacier was representative for the whole mountain range. In the Aleutian Islands and Alaska Peninsula, we applied the specific mass balance measured from sequential DEMs on two icefields around Mount Katmai (covering 580 km^2) to other ice-covered areas (totalizing 2630 km^2). The volume changes for these regions are more uncertain but they have small ice-covered areas and contribute less than 2% to the overall ice loss.

Conversion to regional mass balance and SLR. Total ice volume change in each region is converted to mass change assuming a constant density of 900 $kg\,m^{-3}$. After dividing the mass loss by the maximum ice extent, we obtain an area-weighted mass balance for each mountain range. The total mass loss from Alaskan glaciers is converted to changes in sea level after dividing by the area (362×10^6 km^2) of the global ocean[23].

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Author contributions

E.B. led the development of this study. All authors discussed the results and commented on the manuscript at all stages.

Additional information

The authors declare no competing financial interests. Supplementary information accompanies this paper on www.nature.com/naturegeoscience. Reprints and permissions information is available online at http://npg.nature.com/reprintsandpermissions. Correspondence and requests for materials should be addressed to E.B.