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Sensitivity of Glaciers and Small Ice Caps to Greenhouse Warming

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Recent field programs on glaciers have supplied information that makes simulation of glacier mass balance with meteorological models meaningful. An estimate of world-wide glacier sensitivity based on a modeling study of 12 selected glaciers situated in widely differing climatic regimes shows that for a uniform 1 K warming the area-weighted glacier mass balance will decrease by 0.40 meter per year. This corresponds to a sea-level rise of 0.58 millimeter per year, a value significantly less than earlier estimates.

The amount of ice stored in glaciers and small ice caps is only a fraction of the total amount of land ice on earth. Nevertheless, because glaciers have a shorter response time than the large ice sheets of Greenland and Antarctica, they are thought to be important for sea-level variations on a decadal and centurial time scale. The total volume of water contained in glaciers outside Greenland and Antarctica is not well known—it is probably equivalent to a sea-level rise between 0.3 and 0.7 m. Data on glacier area are more easy to obtain, and a current global estimate is 549 × 10^3 km^2 (*1*).

Global sea level has risen by 10 to 20 cm over the last hundred years (*2*). Explanation of this rise is a matter of debate. As it is known that mountain glaciers have retreated in most parts of the world since the late 19th century, it is likely that they made a contribution to the observed sea-level rise. An admirable attempt to quantify this contribution, using available mass-balance data of glaciers, was made by Meier (*3*), who arrived at a contribution on the order of several centimeters. Critical in his approach was the assumption that the climate sensitivity of glaciers is proportional to the annual balance amplitude (*4*), a measure of the mass turnover in a glacier. The assumption was needed for extrapolation of the scarce mass-balance measurements to large areas. In spite of the uncertainties, Meier's work made clear that, when considering sealevel fluctuations on a 100-year time scale, glaciers are a factor to deal with.

In this report, we follow a process-oriented approach and use an energy-balance model for the ice or snow surface to generate annual mass-balance for a selection of 12 glaciers, for which reliable mass-balance observations extending over a number of years exists. After having tuned the model to give a sufficiently accurate simulation of the observed mass-balance fields, we perform a sensitivity analysis.

The specific (or surface) mass balance of a glacier is the net amount of ice mass gained through a year at the surface of a glacier per unit area. We express the mass balance in meters of water equivalent per year. Precipitation, radiation, and air temperature are the most important factors determining glacier mass balance. For a specific region, the altitudinal gradient is most evident and is a direct consequence of the atmospheric temperature lapse rate. We refer to a mass-balance profile $B(h)$ as the balance that depends on altitude h. As elevation increases, $B(h)$ changes sign (from negative to positive) at the equilibrium line. The equilibrium line altitude E varies widely over the globe from greater than 4000 m in the subtropics, to 2900 m in the Alps, and to less than 1000 m in the subpolar and polar regions.

Observed mass-balance profiles for the set of 12 glaciers show a wide variety in character (Fig. 1). These are only a small part of all the glaciers and ice caps for which mass-balance measurements have been performed. The present selection represents well the wide range of climatological settings in which glaciers and small ice caps are found, perhaps with the exception of the tropics. Glaciers in the tropics, however, contribute a negligible amount to the total ice volume.

The mean specific balance of a glacier, denoted by B_m, is the mass balance averaged over the entire glacier. To obtain B_m for a specific glacier, the area-elevation distribution $S(h)$ (hypsometry) must be known. So

$$B_m = \frac{1}{S_T} \sum B(h_i)S(h_i)$$

Here the sum is taken over the entire glacier, where the index i refers to the elevation interval centered around h_i. S_T is the total area of the glacier. When a noncalving glacier is in perfect balance, $B_m = 0$.

The mass-balance model we use is based on the following equation:

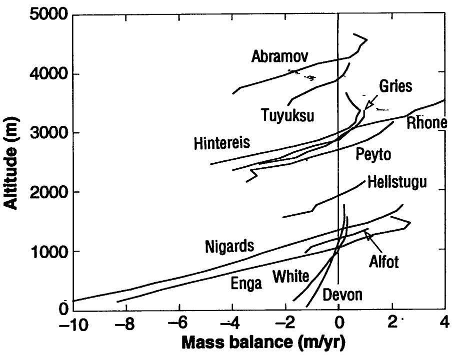

Fig. 1. Mass balance as a function of altitude for the 12 glaciers selected for this study (*5*). Note the wide variety in equilibrium line altitude and balance gradients (change of mass balance with altitude). Small balance gradients are found on glaciers in the drier polar and subpolar regions, like Devon ice cap and White glacier in the Canadian Arctic. Engabreen (Enga) and Nigardsbreen (Nigards), in western Norway, have a tremendous mass turnover and large balance gradients. The same applies to the Rhone glacier in the Swiss Alps. The mass turnover of Hintereisferner (Hintereis) (Austrian Alps) is moderate. Hellstugubreen (Hellstugu) is located in the drier part of southern Norway, and Peyto glacier is in the central Rocky Mountains in Canada. Abramov glacier and Tuyuksu glacier are located in central Asia and are more representative of glaciers in the subtropics.

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$$B = \int_{\text{year}} [(1 - f)min(0;-\Psi L) + P^*]dt$$

L is the latent heat of melt, Ψ is the energy balance at the ice or snow surface, and f is the fraction of melt water that does not run off but refreezes in the underlying snow pack when it is sufficiently cold. P^* is the rate at which solid precipitation is added to the surface. Thus there are two contributions to the mass balance: (i) melt and runoff whenever the surface energy balance becomes positive and (ii) mass gain at a rate P^*. The energy balance has many components: solar radiation, atmospheric (infrared) radiation, turbulent fluxes of heat and moisture, and energy used for heating up the upper snow or ice layers. All these components are calculated in the model according to schemes that are used widely in boundary-layer meteorology (6, 7). Special care was given to the treatment of the surface albedo (amount of reflected solar radiation), as even on a single glacier it varies enormously in space (albedo decreases down a glacier) and time (albedo decreases during the course of the melt season). The albedo was calculated in the model, as this guarantees that in climatic change experiments the albedo feedback is properly taken into account.

The simulations were made in a mode with a full daily cycle. The mass balance was generated on a one-dimensional grid such that grid points were 100 m apart in terms of surface elevation. Several years of integration were needed to obtain an equilibrium mass-balance profile. Climatological information was provided as input, taken from various climatological tables, maps, and some handwritten compilations. Reliable information on precipitation is the most difficult to obtain. Even for well-studied glaciers the altitudinal gradients in precipitation are only broadly known. For all glaciers studied a good match between observed and simulated mass balance could be obtained by adjusting the precipitation gradients within their range of uncertainty (Fig. 2). This result supports the view that energy-balance modeling is an appropriate tool to study glacier mass balance.

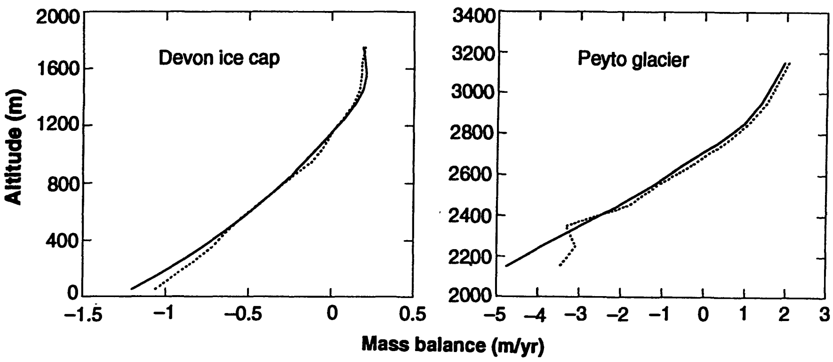

Fig. 2. Two examples of simulated mass-balance profiles (solid lines), compared with observations (dashed lines). The other simulations (not shown) have similar errors.

Many sensitivity experiments have been done. As the ultimate interest is in the changes in global ice volume and not so much in the balance profiles themselves, we concentrate on the mean specific balance B_m. Changes in B_m for a 1 K warming, uniform through the year, varied from -0.12 to -1.15 m/yr. This is a wide range. A regression analysis was carried out to see if a systematic relation between sensitivity and input parameters (latitude, mean temperature, precipitation rate, and so forth) could be found. It appeared that there was a significant relation only with annual precipitation, in such a way that glaciers in a wetter regime are more sensitive (Fig. 3). This relation is not unexpected and is also implied in Meier's (2) assumption that the mass-balance perturbation is proportional to the balance amplitude mentioned earlier.

The change in the altitude of the equilibrium is not related in a clear way to precipitation (Fig. 3), or to any other meteorological quantity. The decisive factor affecting changes in the altitude of the equilibrium line is the balance gradient at the equilibrium line, which is not a simple function of any climatic parameter. This result implies that no clear relation exists between change in mean-specific balance and change in the altitude of the equilibrium line, even though such relations have been found for year-to-year variability on individual glaciers.

A logarithmic fit to the data points in Fig. 3 was used to make further estimates of total ice wastage from glaciers and small ice caps:

$$\delta B_m = -0.512 - 0.662 \log (P)$$

for $P \geq 0.22$ m/yr where P is the annual precipitation averaged over the glacier in meters per year.

There are several reasons why glaciers in a wetter climate are more sensitive, all more or less related to the observation that these glaciers have a larger mass turnover and thus extend to low altitudes with a relatively high air temperature. A significant fraction of the annual precipitation on these glaciers falls as rain. This fraction of rain will increase when temperature increases. For glaciers in dry regions, this effect is not important. Additionally, the relation between annual air temperature and total melt is not linear. With increasing temperature the melt rate increases, and the melt season becomes longer. This relation implies that, for a given rise in annual temperature, the increase in melt is larger when the melt rate in the unperturbed

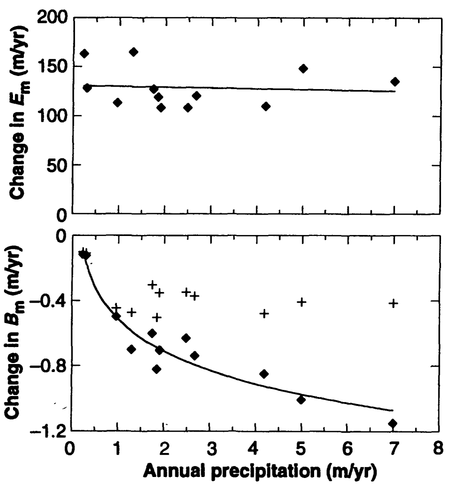

Fig. 3. Effect of a 1 K warming on the mass balance of the 12 selected glaciers, plotted in dependence of the annual mean precipitation. The upper panel shows the change in equilibrium line altitude (E_m); the lower panel shows the change in B_m (mean-specific balance). Crosses refer to an experiment where summer temperature only was raised by 1 K.

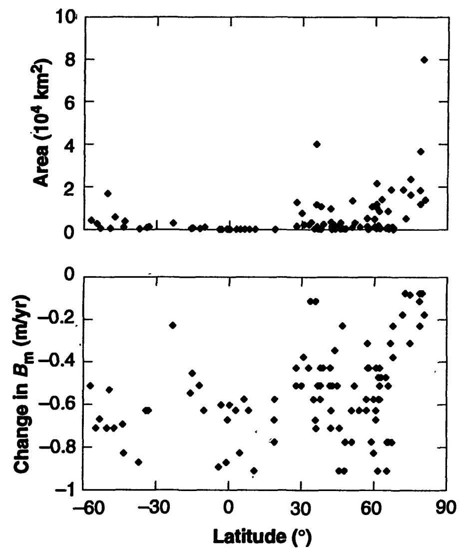

Fig. 4. The upper panel shows the 100 glacierized regions used in the calculation of mean sensitivity, plotted in a latitude-area diagram. The subpolar ice caps in the Northern Hemisphere dominate the picture. The lower panel shows the calculated changes in mass balance.

state is higher. Therefore balance gradients in the ablation zone tend to become larger. With regard to δB_m, the effect is greatest on glaciers in a wet climate. Finally, the albedo feedback contributes more to the sensitivity when precipitation rates are higher.

Results for the summer warming of +1 K (June, July, and August) show a cut-off in the increase of sensitivity in glaciers in moist climates. The occurrence of a cut-off is understandable because the melting on moist glaciers takes place not only in summer but also in fall and spring. Therefore the duration of the melt season will not increase when summer temperature goes up.

Further calculations were based on the assumption that each coherent glacierized region may be characterized by two quantities: the annual precipitation P_k and the glacier area A_k (Fig. 4). The subscript k now refers to the particular region. To arrive at a reliable estimate of the global mean sensitivity, we divided all glaciers into 100 regional groups (8) and assigned to each group a characteristic value of the annual precipitation, based on climatological maps and data from climatological stations. This is not a trivial matter, as glaciers and ice caps tend to form in the wettest parts of terrain with a pronounced topography. Precipitation rates taken from maps and weather stations will therefore underestimate P. On the basis of a comparison between the precipitation rates obtained from maps and from the mass-balance studies on the 12 selected glaciers, we used:

$$P_k = \max\{1.25P_{map}; 0.22\}$$

where P_k is given in meters per year.

The area-weighted mean change in B_m for a uniform 1-K warming is −0.395 m/yr. As the ratio of glacier area to ocean area is 0.00146, this value corresponds to a sea-level rise of 0.577 mm/yr. In comparison, Meier's calculation showed that from 1900 to 1961 glaciers contributed 28 mm (best estimate) to sea-level rise (*3*). A linear fit of Jones' (*9*) temperature data for this period yields a global annual mean temperature rise of 0.29 K. The same procedure applied to Hansen and Lebedeff's (*10*) data gives a rise of 0.41 K. For the purpose of estimating a sensitivity we assume that the temperature rise has been 0.35 K. Meier's estimate of 28 mm of sea-level rise in 60 years would then convert to a rate of 1.33 mm/yr for a 1 K warming. This value is twice that obtained by our modeling approach. However, the error assigned to Meier's estimate was 0.73 mm/yr, so our value is just at the lower bound of his estimate.

The discrepancy can to a large extent be explained by the large spatial variability of temperature trends, which make the empirical estimate of glacier sensitivity given above invalid. Also, in earlier estimates of glacier sensitivity, the subpolar ice caps, which have such a low sensitivity, were assigned too small of a weight. Here we only modeled two subpolar ice caps, both located in Canada. In future studies other small ice caps in the Arctic should be considered.

It has been argued that glacier melt resulting from increasing temperatures could be offset by higher precipitation. We investigated this notion by increasing the precipitation rate in the model by 10% in the case of a 1 K warming. Such a change would be large indeed, at least on the global scale. The result is a reduction of the rate of sea-level rise to 0.329 mm/yr. It thus appears that even a significant increase in precipitation cannot compensate for increased melting, and further shrinkage of glaciers and small ice caps must occur in a warmer climate.

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Sulfate Cooling Effect on Climate Through In-Cloud Oxidation of Anthropogenic SO_2

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Anthropogenic SO_2 emissions may exert a significant cooling effect on climate in the Northern Hemisphere through backscattering of solar radiation by sulfate particles. Earlier estimates of the sulfate climate forcing were based on a limited number of sulfate-scattering correlation measurements from which a high sulfate-scattering efficiency was derived. Model results suggest that cloud processing of air is the underlying mechanism. Aqueous phase oxidation of SO_2 into sulfate and the subsequent release of the dry aerosol by cloud evaporation render sulfate a much more efficient scatterer than through gas-phase SO_2 oxidation.

Sulfur dioxide (SO_2) emissions from fossil fuel combustion may have a climate cooling effect through the backscattering of solar radiation by sulfate particles in cloud-free regions (*1*, *2*). Sulfate aerosol forms in the atmosphere by oxidation of SO_2 either in the gas phase followed by condensation of the sulfuric acid produced or in the aqueous phase of clouds that subsequently evaporate (*3*). In the atmosphere SO_2 and sulfate have a combined lifetime of about 1 to 1.5 weeks, during which they can be transported several thousands of kilometers (*4*). Because vertical exchange between the lower and upper troposphere is relatively slow, the sulfate forcing on climate is exerted predominantly in the lower troposphere of the Northern Hemisphere, where anthropogenic SO_2 emissions are strongest. The mean negative climate forcing by sulfate in the Northern Hemisphere has been estimated at about -1.1 W m^{-2} (*1*), comparable to the positive forcings of about 1.5 and 0.95 W m^{-2} caused by increases of CO_2 concentrations and of other greenhouse gases, respectively (*5*). Sulfate climate forcing in the Southern Hemisphere may be about -0.1 W m^{-2} (*1*). In this report we show that in-cloud SO_2 oxidation followed by cloud evaporation, which results in the addition of sulfate only to that subset of particles that has served as cloud condensation nuclei, strongly influences the scattering properties of the aerosol.

The above estimates of sulfate climate forcing are critically dependent on the optical thickness δ of the aerosol layer, described in Lambert's law of radiation extinction

$$E_\lambda = E_{0,\lambda} e^{-\delta} \quad (1)$$

$E_{0,\lambda}$ is the solar irradiance incident to the aerosol layer (W m^{-2}), E_λ is the irradiance reaching the Earth's surface. Aerosol ex-

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