NOTES AND CORRESPONDENCE

On the Estimation of Antarctic Iceberg Melt Rate

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ABSTRACT

Estimates of Antarctic iceberg melt rates made from field observations, iceberg distribution statistics, laboratory experiments and theoretical studies give a wide range of values. Evaluation of the errors associated with each method allows for the quantitative first-order correction for both the effect of bubbles released from the melting ice on the convective heat transfer and the effect of other forms of iceberg deterioration besides sidewall melting. The results provide a best estimate for the melt rate of 5, 17 and 55 m year\(^{-1}\) at a temperature elevation above the freezing point of \(T_a = 2, 4\) and \(8^\circ\)C, respectively. Also, the laboratory, theoretical and field observations indicate that the melt-rate dependence is proportional to \(T_a^{1.5}\).

1. Introduction

The problem addressed in this note is: Can various melt rate estimates of Antarctic icebergs made from laboratory studies, field observations and iceberg distribution statistics be combined, evaluated and/or reconciled to yield a best estimate of actual iceberg melt rate? Potential users of such information vary from biologists studying polar sea ecosystems to engineers towing icebersgs. We first compare estimates derived from observations and statistics with those from extrapolation of laboratory and theoretical studies. We then examine the errors associated with each of these estimation methods, and the roles of wave erosion and concomitant calving and the release of trapped air bubbles from glacier ice during melting in an effort to account for the differences.

Conclusive field measurements that accurately describe the melting process have not yet been carried out. The impediments are substantial; safety precludes close approach to icebergs, and logistics further complicates the problem. Scientifically, there is uncertainty as to which parameters should be measured and to what precision. In a study of water masses in Muir Inlet, Alaska, Matthews and Quinlan (1975) develop the idea that ice melt from a submerged glacier face yields a characteristic, measurable \(T-S\) relation in nearby waters, but this approach cannot distinguish between mixing by entrainment during convection along the icewall, from which melt rate might be inferred, and mixing in the far field after the convective plume spreads into the interior of nearby waters. Observations by Greisman (1979) in d'Iberville Fjord, Northwest Territories, do not replicate the Matthews and Quinlan (1975) results; he hypothesizes that the discrepancy is due to the much colder \((-1.5^\circ)\)C water temperature in contrast to that in Muir Inlet \((4^\circ)C\). However, a different type of observation in d'Iberville Fjord has yielded a melt rate datum (D. Topham, personal communication). An iceberg was first sighted frozen fast in the fjord pack ice. As melting of the submerged portion proceeded in time, the surface of fast ice near the iceberg was distorted downward. Successive measurements of the distortion together with estimates of the submerged surface area of the berg yield a melt rate of \(\sim 2\) m year\(^{-1}\) in ambient water of \(-1.5^\circ\)C.

Still other estimates of iceberg melt rates (Morgan and Budd, 1978, hereafter referred to as MB) are based on observed iceberg distributions, geographical concentrations and dispersion rates away from the Antarctic continent and are given as a function of the ambient mean temperatures of the upper 200 m of high-latitude Southern Ocean waters. Their method uses observations of iceberg concentrations by Shil'nikov (1969) and size distributions by Romanov (1973) to compute the incremental decrease in berg lengths within successive 1° latitude bands north of the continent. Using estimates of transit
are only weakly useful in iceberg melt rate studies because these results do not include the simultaneous cooling and dilution at the ice-seawater interface nor does the experimental range of the Grashoff number for these results encompass those of icebergs.

2. Estimates based on observations and iceberg statistics

We plot in Fig. 1 the melt rate estimates of MB, together with the datum from d’Iberville Fjord, as a function of the thermal driving $T_d$ defined as the elevation of ambient water temperature above freezing point. MB fit a straight line to their data under the constraint of zero melt rate at zero $T_d$; the result is a melt rate coefficient of $-12 \text{ m year}^{-1}$ ($T_d$) $^{-1}$ constrained by definition to be independent of $T_d$. Both constraints have drawbacks. First, the MB data are more appropriately termed “wastage rates” rather than melt rates since their observations inherently reflect all processes of iceberg deterioration including aerial melting, calving and waterline erosion through wave action. Second, both Josberger (1979) and Greisman (1979) show that melt rate has a power-law dependency on $T_d$; the former reports an exponent of 1.6 from laboratory experiments while the latter gives a theoretically derived 1.4–1.6 dependent on the length scale of the vertical icewall. Still a further complication arises in that MB included basal melt data in their straight-line fit; we consider only sidewall melt rates here.

Line (a) in Fig. 1 is our linear regression to the nine MB sidewall melt data. The line is free of the zero melt rate at zero $T_d$ constraint and has a significantly lower least-squares residue (33%) than the MB line. The $16.8 \text{ m year}^{-1}$ melt rate at zero $T_d$ can be interpreted as sidewall wastage rate due to causes other than melting. If we attempt a power dependency curve fit, the result is curve (b) whose equation is $M = 24.4T_d^{-0.7}$, i.e., a melt rate coefficient which decreases with $T_d$; this is not consistent with the Josberger (1979) or Greisman (1979) results given above.

We propose a different interpretation of the MB data. Since the d’Iberville Fjord point in Fig. 1 is a datum under conditions free of wave erosion and associated calving, we separate the single, nearby MB datum from the remainder of their points on the basis that the latter do incorporate wastage by factors other than melting while the former may not. We then fit a quadratic polynomial to the remaining eight MB points as curve (c) in Fig. 1; its equation is

$$M = 39.3 + 2.78T_d + 0.47T_d^2 (\text{m year}^{-1}),$$

where the 39.3 m year$^{-1}$ is the zero $T_d$ wastage by factors other than melting. Under the assumption
that this wastage rate is independent of $T_d$ the curve (c) can be transposed to pass through the origin at $T_d = 0$, giving curve (d). The latter is then equivalent to a power-law dependency of melt rate on $T_d$ as

$$ M = 2.14T_d^{1.54} \text{ (m year}^{-1}) $$

(2)

the exponent is similar to that given to Josberger (1979) and Greisman (1979).

3. Estimates based on laboratory experiments

For 1 m long, vertical icewalls in a saltwater tank, Josberger (1979) and Josberger and Martin (1980) show that convective flow next to the ice becomes turbulent for Grashof numbers Gr $> 10^8$, where

$$ Gr = g(l^2/\nu^2)(\Delta \rho/\rho). \tag{3} $$

Here $g$ is the acceleration of gravity, $\Delta \rho/\rho$ the density difference between the icewall interface and the ambient water, $\nu$ the kinematic viscosity and $l$ is distance measured from the bottom of the ice. They report that the flow is turbulent for length scales $> 0.4$ m. For the zone of flow above the point of transition to turbulence, and $T_d \leq 12^\circ$C, Josberger (1979) gives the following melt-rate dependencies:

1) The simple average of melt rate measured at several points along the vertical ice face shows a $T_d$ dependency given by

$$ M = 2.45 \times 10^{-2} T_d^{1.4} \text{ (m day}^{-1}) $$

(4)

this laboratory-derived equation is plotted in Fig. 2 in units of m year$^{-1}$.

2) Both laboratory measurements and theoretical results indicate a melt rate that decreases as $z^{-1/4}$, where $z$ is vertical distance above the transition point, given by

$$ M_d(T_d,z) = 1.16 \times 10^{-2} T_d^{1.8}x^{-1/4} \text{ (m day}^{-1}). \tag{5} $$

To extend these results to iceberg length scales, Eq. (5) is vertically averaged over distance $d$, the iceberg draft, as

$$ M_d = (1/d) \int_0^d M_d(T_d,z)dz, \tag{6} $$

to obtain

$$ M_d = 1.55 \times 10^{-2} T_d^{1.6}d^{-1/4} \text{ (m day}^{-1}). \tag{7} $$

In Fig. 2 the curve $M_{200}$ is Eq. (7) evaluated for an iceberg of draft 200 m, a scale size typical of Antarctic icebergs (expressed in units m year$^{-1}$). Superposed in Fig. 2 is our power-law curve interpretation of the MB data [curve (d) from Fig. 1] as well as the MB data points. The melt rates from Eq. (7) are always lower than original MB data but only slightly less than the interpreted curve (d). However, an additional correction, that due to the effect of released air bubbles from glacier-derived ice, is appropriate before final comparison is drawn.

![Fig. 2. Comparison of extrapolated laboratory melt rates with interpretation of MB data: $M_1$, simple average melt rate along a 1 m icewall in a laboratory; $M_{200}$ vertically averaged laboratory melt rates scaled to iceberg draft $d$ of 200 m, incorporating the observed dependency of melt rate on $d^{-1/4}$; $M_d$, vertically averaged laboratory melt rate scaled to iceberg draft $d$ of 200 m, with no power dependency on $d$; transposed MB data (open circles); (d) from Fig. 1, curve fitted to transposed eight MB data; and (e) $M_{200}$, adjusted for bubble-release effect on melt rate.](image)

Glacier ice contains quantities of compressed air bubbles (Gow, 1968). We use the results of Josberger (1980) to estimate the effect of bubble release during melting on the melt rate and convection. Josberger (1980) shows that the specific bubble buoyancy $B_b$ integrated across the boundary layer at a level $z$ above the bottom of the ice wall may be larger than the dilution buoyancy $B_d$ for typical Antarctic glacier ice bubble concentrations given by Gow (1968). The exact ratio of $B_b$ to $B_d$ depends on $T_d$ and the vertical length scale. Some typical numbers are illustrative. For an iceberg with 200 m draft, the ratio of the total vertically integrated buoyancy $B_t$ to the vertically integrated dilution buoyancy alone $B_d$, at $T_d = 2, 4$ and $8^\circ$C is 1.41, 1.75 and 2.31, respectively (Josberger, 1980). Since turbulent heat transfer increases as the third root of the buoyant forcing (cf. Chapman, 1960), we make a first-order modification to the $M_{200}$ curve of Fig. 1 by multiplying melt rates at these values of $T_d$ by $(B_t/B_d)^{1/3}$ to get curve (e), Fig. 2.

4. Discussion

We now draw a comparison between our analysis of the MB data, curve (d) Fig. 2, and the laboratory
results scaled to icewalls of iceberg length and corrected for bubble release effect, curve (e). The graphs are not significantly different over the \( T_d \) range to 12\(^\circ\)C since the error in laboratory-derived estimates is \( \pm 5\% \) (Josberger, 1979) and the standard deviation of the MB data is \( -10 \text{ m year}^{-1} \). This result supports the power-law dependency of sidewall melt rate on the ambient seawater temperature: curvature goes as \( T_d^{1.6} \) for the laboratory-derived graph and \( T_d^{1.54} \) for the transposed curve fitted to MB data.

Extrapolation of laboratory results is questioned from two aspects: the mismatch in Grashoff numbers and the neglect of stratification effects. In experiments with Gr \( \approx 10^{10} \) melt rate decreases with vertical distance along the turbulent boundary layer. In the iceberg case, Gr \( \approx 10^{18} \), turbulent heat transfer to the ice may be large enough to yield a melt rate either independent of or increasing with \( z \). For example, in Fig. 2 we plot Eq. (7) with no \( z \) dependency; the result is essentially curve \( M \). At this time we cannot resolve this question but believe that (7) represents a minimum melt rate for Antarctic icebergs. With respect to stratification effects, Huppert and Turner (1978) and Huppert and Josberger (1980) found in laboratory studies that when the far field is stratified the convection occurs in multiple layers, thus limiting the vertical transport of fluid in the boundary layer. When extrapolated to oceanic stratification values typical of Antarctic seas the layer size may be of order 100 m, i.e., of vertical scale equivalent to real iceberg draft. The drawbacks of extrapolating from these stratification studies are the relatively strong density gradients of laboratory experiments in comparison to the polar ocean (Neshyba, 1978) and the small Grashoff numbers that result from layer thicknesses of a few centimeters.

Our analysis of the MB data is based mainly on a separation of observed wastage rates in iceberg sizes into two parts, one due to sidewall melting and the other to size loss by other mechanisms of deterioration. Assuming the latter part to be independent of \( T_d \) allows transporting the fitted polynomial to pass through the origin (Fig. 2) and hence direct comparison to the laboratory-derived curve. This assumption has the drawbacks that 1) mechanical stress due to thermal expansion in the outer portions of an iceberg is higher in warmer ambient waters, and 2) the surface wave regime to which the icebergs are exposed varies with latitude, increasing in intensity toward the roaring forties. Both factors would augment non-melt-caused wastage rate with increasing \( T_d \); the implication is that the curve (d) of Fig. 2 should show a lower than \( 1.54 \) exponent dependency of \( T_d \) and possibly show significantly lower melt rates than the laboratory-derived curve (e). We are unable to evaluate either of these drawbacks at this time. While Robe et al. (1977) attribute the major fraction of ice loss from North Atlantic icebergs to calving and melting at the waterline, and Josberger (1978) has measured waterline melting rates of a vertical icewall exposed to waves in a laboratory tank, it is not clear that these results are applicable to the larger and generally tabular Antarctic bergs: first, the tabular berg is highly stable in comparison to the North Atlantic berg which tends to roll often and thus continuously expose new faces to wave erosion, and second, tabular Antarctic bergs are often seen with deep wave-cut benches (Neshyba, personal observation) to imply that waterline erosion is limited in the absence of berg roll.

On the other hand, the MB results are based on the Romanov (1973) size distribution which is heavily skewed toward short berg lengths; for example, south of 65\(^\circ\)S average berg length is slightly less than 0.5 km with model lengths of 0.1–0.2 km. In contrast, the Gordienko (1960) distribution of lengths, also south of 65\(^\circ\)S, has an average of 1.5 km and a mode of \( \sim 0.4 \) km. The melt rate estimation procedure used by MB is particularly sensitive to the size distribution since they obtain the minimum length required for a berg to reach a given latitude by assuming complete melting of the number of smallest length bergs just needed to account for the loss in concentration given by Shil'nikov (1969) as a function of distance from the continent. Clearly, use of the Gordienko length scales in this method would have increased the derived incremental length loss as a function of latitude and hence the derived wastage rates. The implied consequence would be an exponent higher than 1.54 for curve (d) of Fig. 2, i.e., in a direction to offset the correction (not yet possible) attributed to wave erosion.

In summary these melt rate estimate analyses based on both laboratory and field observations are in statistical agreement insofar as we are able to correct quantitatively for the uncertainties in each. Sidewall melt rate increases from zero at \( T_d = 0\)\(^\circ\)C to about 10 m year\(^{-1}\) at \( T_d = 3\)\(^\circ\)C. For \( T_d \) between 5 and 12\(^\circ\)C, melt rate increases from \( \sim 25–100 \) m year\(^{-1}\), respectively. We believe that these represent near minimum values for iceberg sidewall melting rates.

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