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Abstract

The long-term mass balance, daily runoff and internal water storage of the Bering Glacier, Alaska is determined with the PTAA model using meteorological observations at Cordova and Yakutat, plus the area-altitude distribution of the glacier. Precipitation and temperature observations collected daily at these low altitude stations are converted to snow accumulation and rain, and snow and ice ablation at each of the 93 area-altitude intervals that make up the total glacier. The transformation from meteorological observations to glacier precipitation and ablation is accomplished with algorithms in a deterministic computer program using 14 variable coefficients. The model is calibrated by the regression of one simulated balance variable with another, each day throughout the summer ablation season. Eighteen different combinations of balance variables are regressed using two-degree polynomials (for a total of 2196 regressions). The 14 coefficients are simultaneously adjusted with a simplex optimizing program until the average regression error is a minimum. The glacier has lost mass at the rate of one meter per year since 1950. The average snow accumulation (winter balance) is 1.7 m and average ablation (summer balance) is -2.7 m. Runoff (the sum of precipitation as rain plus total ablation) is 4.5 m per year; 22 percent of this is derived from the loss in glacier mass. Discharge from the glacier averages 3000 m$^3$ per second in August and has reached a daily peak of 9000 m$^3$ per second (which corresponds to an instantaneous peak approximately two times this value). There is strong evidence that Bering Glacier surges are triggered by several consecutive years of above-normal winter balances. There is also some indication that high discharges of water from the glacier and an increase in internal water storage may also be instrumental. Independent balance measurements made at points on the Bering Glacier and over the nearby Seward-Malaspina Glacier show fair to excellent agreement with the simulated balance.

Introduction
The Bering is the largest glacier in Alaska. Its area ranges from 4700–5800 km$^2$, depending on which of its tributaries are included, with an altitude range that extends from sea level to over 4650 m, or to nearly the summit of Mt. St. Elias (5490 m). Periodic surging is a striking characteristic of this glacier. Folded moraines indicate a 20-year surge cycle that has probably gone on for centuries, regardless of the climate or other seemingly obvious causative factors (Post, 1972, 1999). Figure 1 shows the approximate boundary of the Bering-Bagley Icefield used in this study (4773 km$^2$). Figure 2 is one of the many Austin Post photos of Bering Glacier.

Figure 1. Bering Glacier, Alaska. Map outline supplied by Yann Merrand, Quaternary Research Center, University of Washington. The glacier area used for this study is 4773 km$^3$. The ELA is approximately 1320 meters.
The purpose of this study is mainly to test the PTAA (precipitation-temperature-area-altitude) mass-balance model on one of the largest non-polar glaciers on the earth. Knowing both the current and historic mass balance of these large ice masses is especially important for determining their contribution to global sea-level changes (Meier, 1984). In addition, it is thought that an improved understanding of the influence of mass balance, runoff and internal water on glacier surges can be extracted from these results.

Input Data
Only two reliable long-term weather stations exist in the vicinity of Bering Glacier. Cordova lies about 125 kilometers to the west of the glacier and Yakutat about the same distance to the ESE. Figure 3 shows the location of these long-term weather stations with respect to the glacier.

![Location map](image.png)

**Figure 3.** Location map showing Bering Glacier and weather stations used to simulate the mass balance.

<table>
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<th>Table 1. Weather Stations</th>
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Daily precipitation at the glacier is based on both Cordova and Yakutat observations, times a multiplier that is a function of altitude. However, over much of the glacier there is a rain shadow effect that greatly reduces precipitation. The precipitation multiplier therefore
increases to a maximum at an altitude of 2090 meters and then decreases. At 3000 meters precipitation is approximately one-half of that observed on the coast at Yakutat.

The area-altitude distribution is derived from a compilation of 1 meter altitude increments of area formed into 93–50 meter altitude intervals, shown in Figure 4 (Merrand, 1999, personal communication).

![BERING GLACIER AREA-ALTITUDE DISTRIBUTION](image)

**Figure 4.** The area-altitude distribution of the Bering Glacier that was applied to calculate the mass balance. There are 93- 50 meter altitude increments. (Compiled in 1 meter increments by Yann Merrand from USGS dem, 3x3 arc-second resolution.)

**The PTAA Model**

The computer model used for this study has been tested and applied for Columbia Glacier, Alaska and South Cascade Glacier, Washington (Tangborn, 1997, 1999). A detailed description of the model is provided in the South Cascade Glacier report.
Model Calibration

The balance model is calibrated by regressing one calculated balance variable against another (for example, the mass balance versus the ELA) to obtain the best fit, then using the resulting root mean least square error in a simplex optimization process to determine optimal coefficient values. It is the internal consistency that exists among the different mass balance variables that provides the means to produce meaningful balance results. Both linear and 2-degree polynomial regressions are used. The algorithms that convert meteorological observations to snow accumulation and snow and ice ablation are identical to those used in earlier applications of this model for Columbia and South Cascade Glacier. Seven different balance variables are applied to define eighteen regression pairs (there would be twenty-two if all combinations were used, however only those variables that are always positive were used as dependent variables).

Table 2 summarizes the calibration process. The least squares error (LS\_error) for the dependent variable is calculated for each day and for each regression pair, thus from June 1 to September 30 there are 2196 (122 x 18) regressions made and averaged for each iteration (Table 2 is an example for just one of these days). The calibration process used for the Bering Glacier varies slightly from earlier versions of the model in that the root-mean-square error instead of the R\(^2\) complement, (1-R\(^2\)), is used as the objective function:

\[
\text{rmse} = \frac{\text{LS\_error}}{Y_{\text{mean}}}
\]

where:

\[
\text{rmse} = \text{root-mean-square-error (average of all regressions)}
\]

\[
\text{LS\_error} = \text{Least squares error of polynomial regression}
\]

\[
Y_{\text{mean}} = \text{Dependent variable mean}
\]

To avoid having negative values for Y\_{mean}, the selected dependent variables are always positive (the two balances and the mean summer temperature are applied only as independent variables). Figure 5 demonstrates the 277 iterations comparing the rmse with the independently calculated ELA. The close approximation of the simulated to the observed ELA suggest that the balances and other variables generated by the model are physically real.
Figure 5. Calibration results demonstrating the relationship between the minimum error and the calculated ELA. Each point is one iteration out of a total of 277. The optimization error is the root mean square error divided by the mean of the independent variable used in the regression (see Table 2). The mean ELA determined by the model for the period is approximately 1320 m. The larger triangles represent the 15 manually-introduced coefficients used to start the simplex optimization. The smaller dots represent coefficients automatically determined in the simplex to find the minimum RMSE.

The average root-mean square-error for all regressions is used as the objective function in the simplex. When the minimum \( rmse \) is reached the calibration is complete and the determined coefficients can be applied for further calculations. Figures 6-10 are examples of these regressions plotted for just one day (September 30) of the 122 days.
Figure 6. The mean summer temperature (integrated over the total glacier) versus the ELA for each year of the 1950-96 period. The $R^2$ for this sample is 0.22 (regression number 9 in Table 2). Based on this relationship, a one degree C increase in summer temperature will raise the ELA approximately 90 meters.
Figure 7. The annual glacier balance versus the ELA for each year of the 1950-96 period. The $R^2$ for this sample is 0.91 (regression number 1 in Table 2).
Figure 8. The minimum balance at the glacier terminus versus the ELA for each year of the 1950-96 period. The $R^2$ for this sample is 0.63 (regression number 5 in Table 2).
Figure 9. The snowline altitude versus the ELA for each year of the 1950-96 period. The $R^2$ for this sample is 0.63 (regression number 13 in Table 2).
Results

Mass balance

The average annual balance for the 1950-96 period is -1.00 m, the winter balance is 1.73 m and the summer balance -2.73 m, thus the glacier has thinned nearly 50 m during this period. It is noteworthy but likely not significant that South Cascade Glacier also has had a –1 meter loss per year over about the same time period. The altitude distribution of the annual balance b(z), winter c(z) and summer (a(z) is shown in Figure 11. The annual balance for individual years (Figure 12) indicates that the balance has been positive only a few times in the past 47 years. The winter balance for individual years is shown in Figure 13.
Figure 11. The annual, winter and summer balance distribution for each 50 meter altitude interval, averaged for the 1950-96 period. The altitude of maximum precipitation is 2090 meters.
Figure 12. The annual balance for each year of the period. Only eight years show a positive annual balance. The average annual balance for the period of record is –1.0 meters.
Figure 13. The winter balance for each year of the period. The average winter balance is 1.73 meters. The 1966-67 and 1993-95 glacier surges occurred subsequent to one or two years of above normal winter balance.

There is strong evidence that several consecutive years of above-normal snow accumulation rates over the upper glacier will trigger a surge of the Bering Glacier. The cumulative deviation of winter balance from the long-term average shows a close link to the two major surges of this glacier that have been observed in the past half-century. Cumulative winter balance deviation is found by:

\[
\frac{dB_w}{C_{db}} = \frac{B_w - B_w}{dB_w}
\]

where:

\[
dB_w = \text{Deviation of winter balance from long-term mean}
\]
Figure 14 shows $C_{db}$ for the period of record and reveals the close connection that prior increases in mass over the glacier had on surge events in 1966-67 and 1993-95. The 1993-95 surge was much more prominent than the one in 1966-67 and produced a significant terminus advance (Molnia, 1993). The increased load of several cubic kilometers of ice on the upper glacier also appears to have affected ice flow rates. Interferometer studies based on SAR images show an acceleration of ice in the Bagley Icefield subsequent to the surge onset in 1993 (Fatland and Lingle, 1998). Prediction of glacier surges a year or two in advance of their onset would be feasible by updating the meteorological input files and applying the PTAA model on an annual schedule.

$B_w =$ Winter balance

$B_w =$ 1950-96 average winter balance 1.73 m(we)

$C_{db} =$ Cumulative deviation of winter balance

Figure 14. The cumulative winter balance deviation from normal reveal two periods when the winter balance was above normal for
several consecutive years, culminating in a surge that is thought to be due to the great increase in mass over the upper glacier. The observed 1993-95 surge was much larger than the one in 1966-67.

Figure 15 is the cumulative daily balance for the period of record that demonstrates the annual variations in mass exchange and also shows the rate that the glacier has lost mass nearly continuously for past 47 years.

Figure 15. Cumulative daily balance for the period of record. The average date of the maximum balance is May 12 (but ranges from April 1 to June 23), and the average date of the minimum is October 5 (ranges from September 15 to October 25).
The zero balance altitude (ZBA) is defined as the altitude that $b(z) = 0$. The maximum altitude of ZBA each year is then the equilibrium line altitude (ELA). The simulated ELA for each year is shown in Figure 16. The mean ELA for the period is approximately 1320 m., which is considerably higher than the ELA mean of 1037 m reported by Viens (Viens, 1997). The simulated ELA ranges from 800 in 1973 to 1920 m in 1996. The difference between observed and simulated ELAs may be due to difficulties in making precise time and space observations of this parameter.

Figure 16. The ELA for each year of the 1950-96 period. The mean is approximately 1320 m.

Comparison with Measured Balances

The eruption of Mt. Spurr in 1992 provides a horizon marker of volcanic ash that can be used to measure accumulation rates along crevasse walls. Two such point measurements were made in 1996 by the Quaternary Research Center at the University of Washington. In addition, the QRC measured total ice ablation over a one year period at 1000 meters altitude (Merrand, 1999, personal communication). The mean, long-term accumulation rate at 5500 meters altitude on Mt. Logan was determined by deep coring to be 0.5 m (we) a$^{-1}$ (Holdsworth, 1989). These four measurements are shown in Figure 17 together with the 1992-96 average winter, summer and annual balances simulated by the model.
Figure 17. The mass balance of Bering Glacier calculated by the PTAA model averaged for the 1993-96 period, compared with field measurements made in 1996 and a coring on Mt. Logan at 5500 meters altitude. The coring determined a long-term mean accumulation rate of 0.5 m(we) per year. In 1996, snow depth measurements were made in crevasses to the ash layer deposited by the eruption of Mt. Spurr in September 1992. The ablation measurement at 1000 meters was made for the July 1997 to August 1998 period. The mean ELA for the 1993-96 period determined by the model is 1500 m, which is about 200 m higher than normal.

The independently measured balance of the nearby Seward-Malaspina Glacier for the 1972-95 period was determined using a small aircraft laser altimeter GPS system developed by Keith Echelmeyer and Will Harrison at the Geophysical Institute at the University of Alaska (Li, et al. 1997). The area-averaged thickness change during this period for this glacier system is –23 meters, or –0.90 m(we) per year. For the same 23 year period, the Bering Glacier mean annual balance determined by simulation is also –0.90 m(we).

**Balance Flux**

As snow accumulates on the upper glacier there is a movement of mass down glacier that is defined as the balance flux. It is calculated from the uppermost area interval to the terminus by:
Where:

\[ M_{fx} = \frac{? \, a(z) \, b(z)}{0.9 \, A_g} \]

Where:

\[ M_{fx} = \text{Mass balance flux in } M^3 \text{ of ice per year} \]

\[ a(z) = \text{Area-Altitude fraction} \]

\[ b(z) = \text{Balance in meters} \]

\[ A_g = \text{Area of glacier in } m^2 \]

Positive values of \( M_{fx} \) indicate that the vertical flow component is down and negative means an upward flow component. Figure 18 is the mean mass flux of ice as a function of altitude. The mass balance flux is an important calibration variable because it is the primary cause of glacial erosion, which is thought to be the connecting link between the glacier’s AA profile and the climate. A significant increase in mass flux may also be a triggering mechanism for a glacier surge.
Figure 18. The mass balance flux is calculated by summing mass balance times area of each interval from the head to the glacier terminus.

Runoff

Total runoff from the glacier is the sum of precipitation as rain and the ablation of snow and ice. A more accurate runoff determination would take into account inflow and outflow of water from internal storage, however, until more is known about the mechanisms that control internal storage, the storage component of runoff will be omitted in this preliminary analysis. Figure 19 is the mean daily discharge from the Bering Glacier as determined from rain and ablation. Figure 20 is the daily average for both the period of record and for 1993. There is a large increase in simulated discharge from approximately July 10-20 in 1993 that is likely associated with an outburst of water observed at the terminus that month (Fleisher, et al., 1998).
Figure 19. Mean daily runoff from the Bering Glacier is the sum of daily ablation of snow and ice, plus precipitation as rain.
Figure 20. Runoff for the 1993 balance year compared with the 1950-96 average. From approximately July 10-20, the discharge was nearly two times the long-term average.

The annual runoff for individual years for the period of record is shown in Figure 21. Average annual runoff is 4.5 meters; ablation of snow and ice accounts for 61 percent of the total runoff (2.73 m) and precipitation as rain is 39 percent (1.77 m). The 1 meter annual loss in glacier mass is then about 22 percent of total runoff and precipitation (rain plus snow) 78 percent.
Figure 21. Annual runoff for the Bering Glacier. The yearly average is 4.5 m. Approximately 39 percent of annual runoff is from precipitation as rain and 61 percent is from ice and snow ablation. Loss in glacier mass (-1 meter per year) is 22 percent of the long term-average, thus precipitation accounts for 78 percent of total annual runoff.

**Internal Water Storage**

The internal storage of water within the glacier is derived from the simulated daily runoff record by assuming that when runoff from precipitation and ablation are low, water is stored within the glacier. Storage occurs at this time because ice flow or creep has closed drainage conduits that are usually kept open by high flows. When precipitation and ablation rates are high the sub-glacial drainage channels and conduits are opened and kept open as long as there is a large influx of meltwater. Water is then released from storage when ablation and precipitation rates are high, thus increasing discharge from the glacier to an even greater magnitude.

Internal water storage is simulated as follows:

\[ S(i) = ? XR(i) - Q(i) \]
where:

\[ S(i) = \text{Internal storage on day } i, \text{ in mm averaged over the glacier} \]

\[ XR(i) = \text{Runoff that is less than a maximum threshold (mm/day)} \]

\[ R(i) = \text{simulated runoff on day } i, \text{ (mm/day)}, \]

\[ Q(i) = \text{Discharge of water from storage on day } i, \text{ (mm/day)} \]

\[ = C_x S(i) \text{ when } R(i) > R_x \]

\[ R_x = \text{Runoff threshold (mm/day)} \]

\[ C_x = \text{Storage release coefficient} \]

The depiction of internal storage \((S(i))\) shown in Figure 22 is based on a threshold value \((R_x)\) of 10 mm (slightly less than the daily mean) and a release fraction \((C_x)\) of 0.10. These are approximately the same values used to calculate internal storage for Columbia Glacier and then used to relate it to that glacier’s flow rate. There was a unusual increase in internal storage in 1993 that reached a peak about May 15 and was nearly depleted by mid-June. The release of this amount of water over a 30 day period translates to an average discharge of over 600 cubic meters per second, which would be in addition to runoff generated by rain and ablation. The 1993-95 surge began at about the same time as the release of stored water, but a cause and effect connection cannot be made without further analysis. During surge conditions, water is thought to be slowly dispersed across the bed of the drainage system. When the glacier is not surging, water is transported rapidly through a conduits and tunnels (Bjornsson, 1998).
Figure 22. The internal storage of water within the glacier is determined from the daily runoff record. The 1993 peak occurred on May 15 and may be related to the surge that occurred at this time.

**Balance Dependence on Summer Temperatures**

The relationship between the annual balance and the mean summer temperature integrated over the glacier is demonstrated in Figure 23. The poor correlation ($R^2 = 0.28$) is thought to be due to a significant influence of the winter balance on the annual balance, and that the summer balance is determined by more than just the summer temperature (also by solar radiation, cloudiness and the surface albedo). The relation shown here (weak as it is) indicates that a one degree C increase in summer temperature produces a $-0.4$ m (we) balance. Therefore, if mass balance was entirely dependent on temperature, general warming over the past half-century would need to be $+2.5$ C (rather than the observed $+1$) to explain the $-1$ ma$^{-1}$ (we) loss during this period.
Figure 23. The dependence of the annual balance on the mean summer temperature integrated over the glacier suggests that a 1 degree increase in temperature will decrease the annual balance approximately 0.4 m (we).

Conclusions

Reasonable mass balance and other related variables that agree with measured balances on this and nearby glaciers can be calculated with this model. However, the mean ELA simulated by the model is 2-300 meters higher than has been reported from visual observations. The reason for this difference is as yet unknown but may be due to the difficulty of making precise ELA observations. The total mass lost from this glacier over the past half-century is approximately 200 cubic kilometers of water, equal to a 0.55 mm rise in sea-level. There appears to be strong evidence that several years of above-normal snow accumulation on the upper glacier will culminate in a surge. However, high runoff events and changes in internal water storage may also be instrumental. Further analysis is needed to prove or disprove if such linkages exist.
References


Merrand, Yann, 1999. Personal communication.


Post, Austin, 1999. Personal communication.


Acknowledgements

Thanks to Mark Dyurgerov, Jay Fleisher and Will Harrison for reading the initial manuscript and giving many helpful comments. Also thanks to Yann Merrand for providing the area-altitude tabulations and the crevasse/ash layer accumulation and the ablation measurements. Much appreciation to Austin Post for his insight based on long-term observation and understanding of the Bering Glacier.

Table 2. Eighteen Regressions Used for Optimization of Coefficients

(RESULTS FOR SEPTEMBER 30)

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**KEY**

- **BND** = GLACIER BALANCE
- **ZBA** = ZERO BALANCE ALTITUDE
- **SLA** = SNOWLINE ALTITUDE
- **AAR** = ACCUMULATION-AREA RATIO
- **EXM** = BALANCE FLUX
- **TBS** = MEAN SUMMER TEMPERATURE
- **XMEAN** = MEAN OF INDEPENDENT VARIABLE
- **YMEAN** = MEAN OF DEPENDENT VARIABLE
- **RSQ** = R-SQUARED
- **LSERROR** = LEAST SQUARES ERROR
- **SDEV** = STANDARD ERROR
- **RMSE** = ROOT MEAN SQUARE ERROR

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