Climate reconstruction since the Little Ice Age by modelling Koryto glacier, Kamchatka Peninsula, Russia

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ABSTRACT. Based on the field data at Koryto glacier, Kamchatka Peninsula, Russia, we constructed a one-dimensional numerical glacier model which fits the behaviour of the glacier. The analysis of meteorological data from the nearby station suggests that the recent rapid retreat of the glacier since the mid-20th century is likely to be due to a decrease in winter precipitation. Using the geographical data of the glacier terminus variations from 1711 to 1930, we reconstructed the fluctuation in the equilibrium-line altitude by means of the glacier model. With summer temperatures inferred from tree-ring data, the model suggests that the winter precipitation from the mid-19th to the early 20th century was about 10% less than that at present. This trend is close to consistent with ice-core results from the nearby ice cap in the central Kamchatka Peninsula.

INTRODUCTION

Valley glaciers can serve as an important indicator of climate change due to their rapid response (Oerlemans and Fortuin, 1992; Oerlemans, 1994). The recent increase in the rate of glacier melting is assumed to be contributing significantly to a rise in the global sea level (Meier, 1984; Dyurgerov and Meier, 1997; Gregory and Oerlemans, 1998). Continuous monitoring of glacier mass balance can determine the influence of climate change upon glaciers. However, data regarding mass balances are spatially and temporally limited. The records of glacier terminal positions, including those obtained from geographical features, may solve these problems. However, glaciers tend to exhibit complex responses to climate change due to their dynamic processes, so fluctuations of the termini do not directly indicate the effect of the climate conditions. Several research groups have numerically modelled various glaciers to understand their behaviour over the past few decades and to predict the changes in glacier characteristics (e.g. Greuell, 1992; Oerlemans, 1997; Schmeits and Oerlemans, 1997; Zuo and Oerlemans, 1997; De Smedt and Pattyn, 2003).

We describe a numerical model of Koryto glacier, Kamchatka Peninsula, Russia, based on field measurements obtained in 2000, and we examine the validity of the model by comparing with records of the glacier retreat during the last half-century. Finally, we try to reconstruct the regional climate since the Little Ice Age (LIA), paying particular attention to the trends in precipitation.

KORYTO GLACIER

Kamchatka Peninsula extends into the North Pacific and is the largest glaciated region in northern Asia, with 448 glaciers covering ~905 km^2 (Muravyev, 1999b). Kronotsky peninsula is located on the east coast of Kamchatka (Fig. 1) and has 32 lower-altitude glaciers, some of which extend as low as 250 m a.s.l. Koryto glacier is the third-largest glacier in this region, with an area of 7.55 km^2 and a length of 7 km from the highest point at 1200 m to the terminus at 300 m a.s.l. (Muravyev, 1999a). It has a simple form, with a large accumulation area and a narrow ablation area. Glaciers in Kamchatka are usually covered with debris or volcanic detritus, but the surface of Koryto glacier is clean. The daily surface flow measurements in summer 2000 indicated that basal motion occurred throughout the glacier bed (Yamaguchi and others, 2003). Several terminal moraines were deposited between the LIA and 1930 (Solomina, 1999; Solomina and Calkin, 2003).

The amplitude of the annual mass balance of Koryto glacier, which is considered to be a factor in the glacier–climate regime (Meier, 1984), is the second largest among 50 glaciers in the Northern Hemisphere that have been studied from 1961 to 1990 (Dyurgerov and Meier, 1999). Thus, Koryto glacier is regarded as one of the most extreme maritime glaciers and should be very sensitive to climate change.

DESCRIPTION OF THE GLACIER MODEL

We use a one-dimensional model with the x axis along a flowline, based largely on the model of Oerlemans (1997). The gridpoint spacing is 50 m, and the three-dimensional shape is taken into account by parameterization of the cross-sectional geometry at each gridpoint. Cross-section S is trapezoidal, with the valley width at surface W_s, the ice thickness H and the average slope of the valley walls on each side of the glacier γ as follows:

$$S = 2H(W_s - H\tan\gamma). \qquad (1)$$

The values of W_s and γ were taken from the topographic maps made in 1960 by D.G. Tsvetkov and others (personal communication, 1982). The thickness H at each gridpoint was determined from the bedrock profile obtained from a radio-echo sounding survey in 2000 (Macheret and others, 2001).

The dynamic behaviour of the glacier is described in terms of changes in H that are calculated from the mass continuity equation. Since the glacier is assumed to be composed of ice of uniform density, the conservation

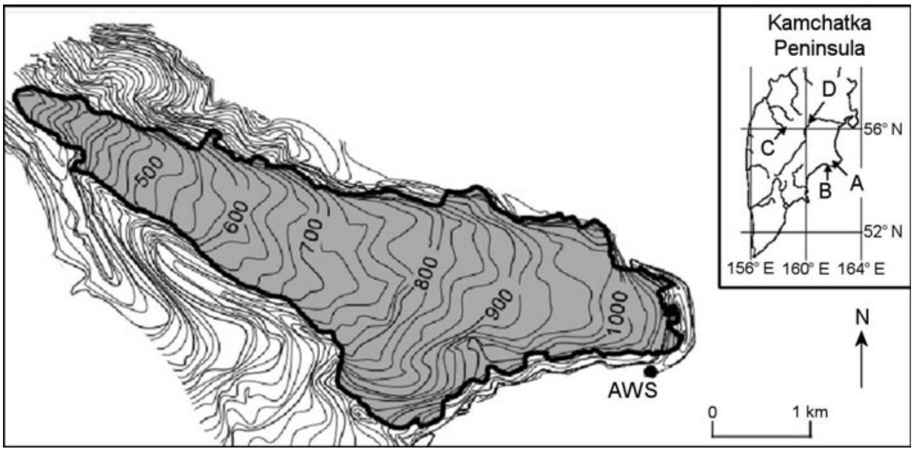

Fig. 1. Map of Koryto glacier (54°50′ N, 161°44′ E) with 20 m contour lines, based on a 1960 map. The results of a global positioning system (GPS) survey in 2000 were used to revise the elevations of the glacier surface. AWS marks the location of the automatic weather station. Features on inset map of Kamchatka peninsula include A: Kronotsky peninsula; B: Stopozh meteorological station; C: Esso; D: Ushkovsky ice cap.

equation for the ice volume is

$$\frac{\partial S}{\partial t} = -\frac{\partial (US)}{\partial x} + BW_s, \tag{2}$$

where U is the depth-averaged ice speed at x and B is the net balance at the surface. Melting of ice at the bottom is neglected. The driving stress τ_d is defined

$$\tau_d = -\rho g H \frac{dh}{dx}, \tag{3}$$

where h is the surface elevation. The surface ice speed U_s is calculated as the sum of the internal ice deformation U_d and the basal motion U_b, i.e.

$$\begin{aligned} U_s = U_d + U_b &= f_1 H \tau_d^3 + f_2 \frac{\tau_d^3}{H} \\ &= f_1 \left(-\rho g \frac{dh}{dx}\right)^3 H^4 + f_2 \left(-\rho g \frac{dh}{dx}\right)^3 H^2, \end{aligned} \tag{4}$$

where f_1 and f_2 are the flow parameters. In the model, the ice deformation speed U_d on the surface is multiplied by 0.8 to calculate the depth-averaged ice deformation speed (Paterson, 1994).

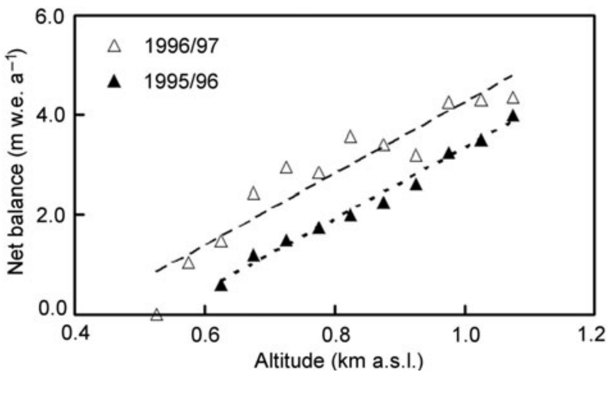

Fig. 2. Altitudinal distributions of net balance in the accumulation area. The black and white triangles indicate data from 1995/96 and 1996/97 respectively. The dotted line is a linear approximation of 1995/96 data with a gradient of 7.1, and the dashed line is for 1996/97 with a gradient of 7.2.

The balance B is assumed to vary linearly with elevation as

$$B(h) = \alpha (h - \mathrm{ELA}), \tag{5}$$

where α is the mass-balance gradient, which is given by different values for the accumulation and ablation areas, and ELA is the equilibrium-line altitude. Climate change is included in the model through fluctuations in the ELA.

DETERMINATION OF EACH PARAMETER

The flow parameters depend on many conditions including the bed features, debris content and crystal structure of the basal ice. The contribution of basal water to basal motion is not explicitly accounted for in Equation (3). The overall effects of these conditions are incorporated into the model as tuning parameters (f_1 and f_2) that vary from glacier to glacier (Oerlemans, 1998).

At Koryto glacier, the distribution of the surface flow speeds at six sites along the flowline (Yamaguchi and others, 2003), the ice thickness and the surface profile (Macheret and others, 2001) were measured. From the data, we determined the values of f_1 and f_2 to best fit the distribution of surface speeds: namely, $f_1 = 8.3 \times 10^{-25}\ \mathrm{Pa^{-3}\,s^{-1}}$ and $f_2 = 2.8 \times 10^{-21}\ \mathrm{Pa^{-3}\,m^2\,s^{-1}}$, assumed to be constant through time and space.

The mass-balance gradient α in the accumulation area was estimated from the altitudinal distribution of the net balance B measured at snow pits and crevasse walls during 1995/96 and 1996/97 (Fig. 2; Shiraiwa and others, 1997; Muravyev, 1999a). Both periods provided a similar value for α, so we used the average value of 7.1 m w.e. a^{-1} km^{-1} for the accumulation area.

No direct measurements of α were made in the ablation area. We therefore estimated the value of α in the ablation area as follows. First, we calculated the altitudinal distribution of ablation B_s in the accumulation area using the measured air temperatures T_a at an automatic weather station (Fig. 1; Matsumoto and others, 1997) with a lapse rate θ of 6.5°C^{-1} km^{-1}. The degree-day factor D varies with the surface conditions from 4.7 mm w.e. °C^{-1} d^{-1} for snow to 7.0 mm w.e. °C^{-1} d^{-1} for ice (Matsumoto and others, 2004). We then reconstructed accumulation B_w in the accumulation area using $B_w = B + B_s$. Next, assuming that the

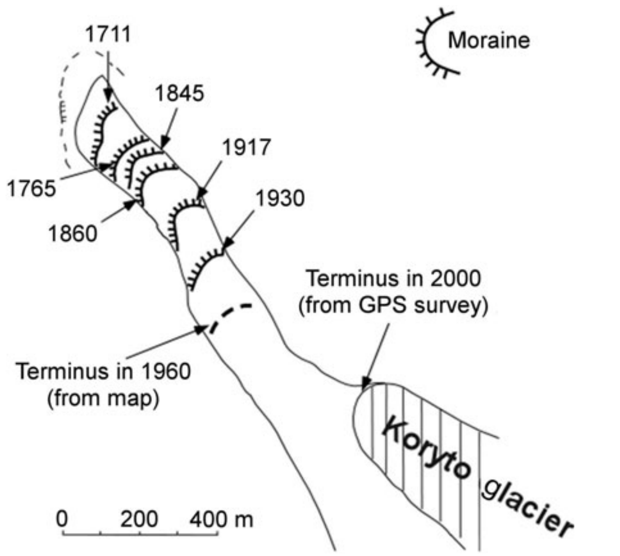

Fig. 3. Distribution of terminal moraines at Koryto glacier, modified from Solomina (1999) using the GPS data surveyed in 2000.

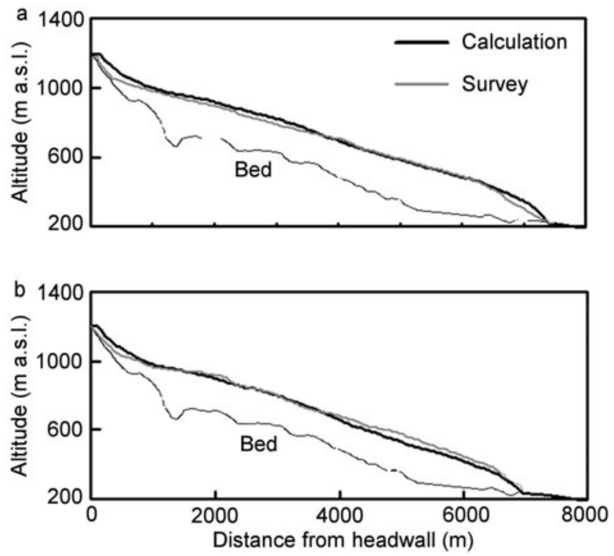

Fig. 4. Calculated surface profiles from the model (thick curves) and profile measurements (thin curves) from (a) 1960 and (b) 2000.

altitudinal gradient of accumulation η was constant from the accumulation area to the ablation area, the altitudinal distribution of B in the ablation area can be estimated from η and B_s calculated using T_a, θ and D. Thus, we determined α in the ablation area as 11.0 m w.e. a^{-1} km^{-1}.

The model requires fluctuations in the ELA to introduce the influence of climate change, but ELAs on Koryto glacier are only known for the years 1960, 1971, 1982, 1984, 1996 and 1997 (Haeberli and Müller, 1988; Muravyev, 1999a). A method of estimating the glacier ELA is therefore required.

In general, ELAs depend on summer air temperature and winter precipitation, since the ELA is determined by the balance of summer ablation and winter accumulation. The nearest meteorological station to Koryto glacier is Stopozh meteorological station, about 50 km southwest of the glacier (Fig. 1 inset, point B). The mean summer air temperature T averaged over June–August and the winter precipitation P over October–May were measured from 1941 to 1995. Here, we assumed that the glacier ELA is a function of T and P:

$$\mathrm{ELA} = \mathrm{k}_1 P' + \mathrm{k}_2 \,\Delta T + \mathrm{k}_3, \qquad (6)$$

where P' indicates the ratio of winter precipitation to the average winter precipitation (1941–90) and ΔT represents the deviation of the summer temperature from the average mean summer temperature (1941–90).

Approximate values of the three constant factors (k_1, k_2, k_3) were obtained empirically using four pairs of meteorological data acquired from Stopozh meteorological station and the ELA (1960, 1971, 1982 and 1984). The values obtained for k_1, k_2 and k_3 were –746, 101 and 1491, respectively. The validity of Equation (6) using these parameters is examined below.

TESTING THE MODEL

Since each parameter in the model and the factors to estimate the ELA were derived independently of the fluctuation to the glacier terminus, we test their validity by comparing the calculated glacier profile and length with its measurements. The terminus positions of Koryto glacier were measured in 1960, 1971, 1982, 1997 and 2000, and the glacier surface profiles were surveyed along the flowline in 1960 and 2000. Using lichenometry, the latest, clearly defined terminal moraine was identified at 550 m downstream of the 2000 terminus with a formation date of 1930 (Fig. 3; Solomina, 1999). Because the terminal moraine is well established, we assumed that the glacier had been stable at that position for some years before 1930. We calculated the steady-state profile of the glacier when its terminus was located at the 1930 moraine, and used this profile and terminal position as the initial conditions in the calculation.

With this initial condition and the fluctuation of mass balance in terms of the inferred ELA, we calculated the surface profile and terminal position for each year from 1930 to 2000. Note that since the ELAs were only estimated as far back as 1941 (when meteorological measurements began at Stopozh), we simply assumed that the ELAs from 1930 to 1940 were equal to the average for 1941–50. In the same way, we compensated the lack of ELA data from 1996 to 2000 with the average value for 1990–95.

We compared the calculated glacier surface profiles with the measurements. Figure 4 represents comparisons in 1960 and 2000. Although the calculated profiles are overestimated in the upper reaches in 1960 and 2000, and near the terminus in 1960, the calculated and measured surface profiles generally showed good agreement. We next compared the calculated glacier lengths with the measurements (Fig. 5); they agreed well. Thus, these two tests indicated that the model provided a plausible simulation of Koryto glacier fluctuations.

DISCUSSION

Causes of the terminal retreat acceleration since the mid-20th century

The terminal retreat accelerated after the 1970s (Fig. 5). Yamaguchi and others (2003) argued that a cause of the acceleration was the decrease in winter precipitation. To validate this hypothesis, we conducted two simulations under the following conditions. Scenario I involves actual

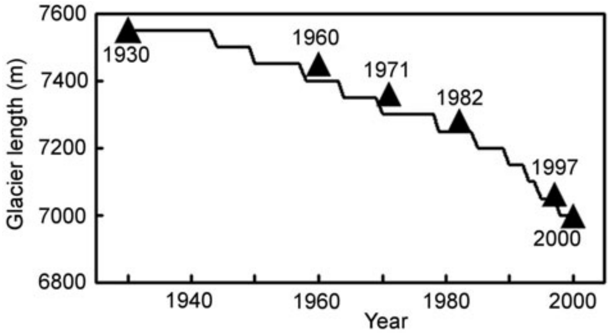

Fig. 5. Modelled (solid line) and measured (triangles) glacier lengths. The steps in the solid curve are a result of the 50 m grid resolution.

fluctuations in mean summer temperatures and an assumed constant winter precipitation after 1975 with the average value before 1975 (Fig. 6a), while scenario II describes actual fluctuations in winter precipitation and an assumed constant mean summer temperature after 1975 with the average value before 1975 (Fig. 6b). In Figure 6c, the simulated and measured glacier lengths are shown.

The results under scenario I show that the glacier terminus remains almost constant after 1982. By contrast, the results under scenario II exhibit a similar fluctuation to the measured glacier termini, which represent a rapid retreat after 1982. These simulated results support the hypothesis that the rapid retreat of the glacier terminus since the 1970s is most likely due to the decrease in winter precipitation.

The standard deviations s of P' and ΔT from 1941 to 1995 were calculated as 0.22 and 0.70°C, respectively. The influences of the fluctuations in both parameters on the ELA were estimated using Equation (6). The fluctuation in ELA caused by s of P' is ±164 m, whereas that caused by s of ΔT is ±71 m, which is less than half of the former value. These results indicate that the ELA of Koryto glacier has mostly been influenced by the fluctuation in precipitation during the latter half of the 20th century, which supports the previous hypothesis.

Reconstruction of the fluctuation in ELA since the LIA

Using the model with the data on terminal moraines (Fig. 3), we now estimate the evolution of the ELA from the LIA to the present. Terminal moraines were lichenometrically dated by Solomina (1999). The oldest formed in 1711, which corresponds to the LIA. Although the reliability of the lichenometric method is difficult to evaluate, the accuracy is considered to be of the order ±15–20% (Solomina and Calkin, 2003). The positions of the moraines were determined with the global positioning system (GPS) in the 2000 survey, and these positions and dates were used to estimate the glacier lengths and retreat rates. Figure 7a clearly indicates that the retreat rate has not been constant since the LIA.

Using the retreat rates of the glacier, we are able to estimate the fluctuation in the ELA by the inverse method; that is, we modelled various ELAs by trial and error to fit the calculated to the measured retreat rates. First, we reconstructed the steady-state profile of Koryto glacier when its terminus was located at the oldest terminal moraine (1711). We then adopted this steady-state profile as the initial condition for this calculation. To calculate variations in the

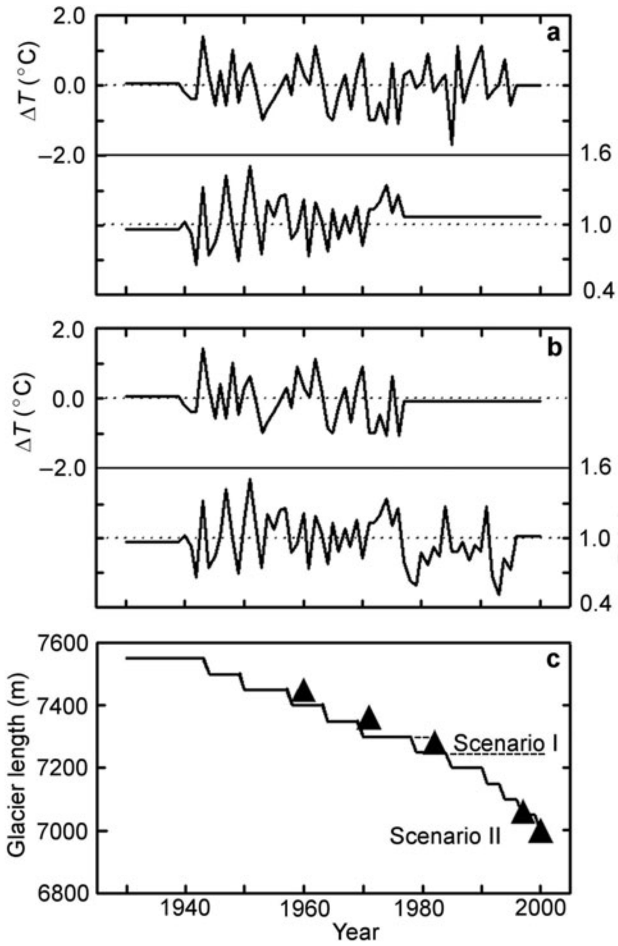

Fig. 6. Fluctuations in glacier lengths under different scenarios. (a) Scenario I: winter precipitation after 1975 assumed equal to the average value before 1975. (b) Scenario II: mean summer temperature after 1975 assumed equal to the average value before 1975. (c) Calculated glacier lengths from the model (solid curve and dotted curve) and measurements (triangles).

ELA, we assumed that the ELA was constant during the period from one moraine to the next. Except for the initial condition, the glacier did not reach a steady state, i.e. the glacier geometry changed with time even when the terminus was located at a moraine.

Figure 7b illustrates the calculated ELAs since 1711. The changes in the calculated glacier length and the measured position of terminal moraines are also depicted. This figure indicates that the ELA was sometimes located at a lower position than in the previous period, even though the glacier had retreated since the LIA. Such a situation occurred when the retreat rate was smaller than in the previous period (e.g. 1765–1845, 1860–1917), together with the delay action due to the response time of the glacier. According to the model, the response time at Koryto glacier is approximately 120 years, which agrees with the order of the value estimated from the mean surface slope (Haeberli and Hoelzle, 1995).

Reconstruction of precipitation since the LIA

Equation (6) indicates that the ELA is determined by a combination of the summer temperature (ablation) and the winter precipitation (accumulation). However, in the 18th and 19th centuries, there were no temperature records near Koryto glacier. As a result, we used a proxy temperature in

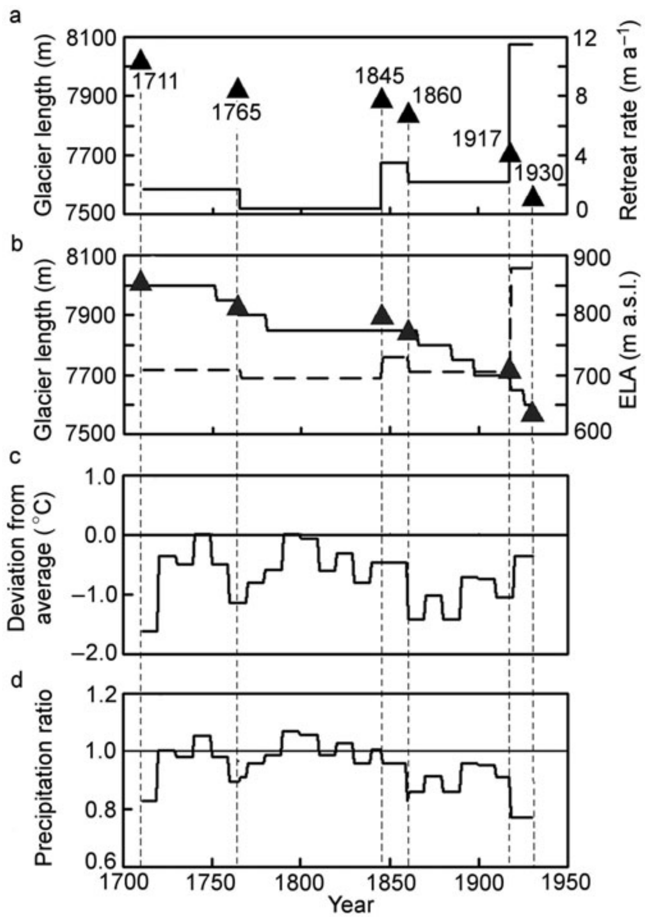

Fig. 7. Input data and results of the model calculation since the LIA. (a) Terminal retreat rates (solid lines) estimated from the moraine positions (triangles). (b) Modelled ELAs (dashed lines) since 1711 with the variation in the calculated glacier lengths (solid lines) and the moraine positions (triangles). (c) Decadal means of the early-summer temperature (May–June) at Esso inferred from tree-ring analyses (Gostev and others, 1996). These data are shown as deviations from the average values (1941–90). (d) Modelled winter precipitation rate P'. The solid lines are the reconstructed decadal means of winter precipitation from the model, normalized to the average (1941–90) values.

early summer (May–June) that was estimated from tree-ring data (Fig. 7c; Gostev and others, 1996) at Esso in central Kamchatka (Fig. 1, point C). We then reconstructed the winter precipitation using the following procedure. Decadal averages of the proxy temperature and their deviations ΔT from the average value (1941–90) for early summer in the Esso region were calculated. The same changes were assumed for the summer temperature between Esso and Koryto glacier. Equation (6) was then used to estimate the fluctuation in P'.

Figure 7d shows the reconstructed winter precipitation, from which two tendencies were obtained after careful examination. First, the winter precipitation from the mid-19th to the mid-20th century was about 90% of the present precipitation (1941–90). Second, despite a considerable fluctuation in the precipitation, the average precipitation for the period from the 18th to the mid-19th century was almost identical to the present values.

An ice core with a depth of 212 m was retrieved from the summit of Ushkovsky ice cap (3900 m a.s.l.) in central Kamchatka (Shiraiwa and others, 2001) (Fig. 1, point D), and

Table 1. Comparison of P' calculated from model and by reconstructed annual precipitation from an ice core at Ushkovsky ice cap

Period	Model	Ushkovsky ice cap
1860–1930	0.88	0.93
1765–1860	0.99	1.06
1711–65	0.97	–

analyzed for annual precipitation (Shiraiwa and Yamaguchi, 2002). We compared the modelled winter precipitation to the core data (Table 1). Each value of P' indicates the mean precipitation over each period from one moraine to the next, normalized by the average value for the period 1941–90.

Table 1 indicates the following tendencies. Between 1860 and 1930, P' estimated from the model shows a precipitation decrease (0.88), and that derived from the ice core at Ushkovsky ice cap exhibits a similar slight decrease (0.93). These results suggest that a decrease in winter precipitation of about 10% since the mid-19th century is quite likely to have occurred over Kamchatka Peninsula. The values of P' between 1765 and 1860 show a different trend: modelled P' is almost the same as at present (0.99) while P' at Ushkovsky ice cap indicates a slight increase (1.06). The causes of the slight difference in trends between modelled and actual data are not clear. Possibilities include the fact that the data from Ushkovsky ice cap represent annual and not just winter precipitation, but also the difference in precipitation regime between the high mountains and the low-altitude glacier.

CONCLUSIONS

A glacier model was developed using field data from Koryto glacier and tested against additional data from the glacier. The results showed that the model correctly predicted the terminus positions as well as the change in the surface profiles for the relevant period.

Employing the model in several numerical experiments, we determined that the rapid decrease in glacier length since the mid-20th century has likely been caused by the recent decrease in winter precipitation. We also concluded that the winter precipitation in the area of the glacier may have decreased by about 10% from the mid-19th to the mid-20th century. This trend is approximately consistent with the results of an ice-core analysis from Ushkovsky ice cap.

This study demonstrated that numerical modelling was effective in the reconstruction of winter precipitation variations in a valley glacier where data on glacier fluctuations and temperatures are available.

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REFERENCES

De Smedt, B. and F. Pattyn. 2003. Numerical modelling of historical front variations and dynamic response of Sofiyskiy glacier, Altai mountains, Russia. *Ann. Glaciol.,* **37**, 143–149.

Dyurgerov, M.B. and M.F. Meier. 1997. Year-to-year fluctuations of global mass balance of small glaciers and their contribution to sea-level changes. *Arct. Alp. Res.,* **29**(4), 392–402.

Dyurgerov, M.B. and M.F. Meier. 1999. Analysis of winter and summer glacier mass balances. *Geogr. Ann.,* **81A**(4), 541–554.

Gostev, M.G., G. Wiles, R. d'Arrigo, R. Jacoby and P. Khomentovsky. 1996. Early summer temperatures since 1670 AD for central Kamchatka reconstructed based on a Siberian Larch tree-ring width chronology. *Can. J. Forest Res.,* **26**(11), 2048–2052.

Gregory, J.M. and J. Oerlemans. 1998. Simulated future sea-level rise due to glacier melt based on regionally and seasonally resolved temperature changes. *Nature,* **391**(6666), 474–476.

Greuell, W. 1992. Hintereisferner, Austria: mass-balance reconstruction and numerical modelling of the historical length variations. *J. Glaciol.,* **38**(129), 233–244.

Haeberli, W. and M. Hoelzle. 1995. Application of inventory data for estimating characteristics of and regional climate-change effects on mountain glaciers: a pilot study with the European Alps. *Ann. Glaciol.,* **21**, 206–212.

Haeberli, W. and P. Müller, *eds.* 1988. *Fluctuations of glaciers 1980–1985 (Vol. 5).* Wallingford, Oxon, IAHS Press; Nairobi, UNEP; Paris, UNESCO.

Macheret, Y., E. Vasilenko, A. Glazovsky, O. Moshina, D. Miroshnichenko and T. Shiraiwa. 2001. Radio-echo sounding of Koryto glacier (Kamchatka, Russia), 2000. *In* Wakatsuchi, M. and T. Hara, *eds. Proceedings of the International Symposium on Atmosphere–Ocean–Cryosphere Interaction in the Sea of Okhotsk and the Surrounding Environment.* Hokkaido, Hokkaido University. Institute of Low Temperature Science, 160–161.

Matsumoto, T. *and 8 others.* 1997. [Meteorological observations by Automatic Weather Stations (AWS) in alpine regions of Kamchatka, Russia, 1996–1997.] *Low Temp. Sci., Ser. A,* **56**, 53–68. [In Japanese with English summary.]

Matsumoto, T., R. Naruse, K. Konya, S. Yamaguchi, T. Yamada and Y.D. Muravyev. 2004. Summer water balance characteristics of Koryto Glacier, Kamchatka Peninsula, Russia. *Geogr. Ann., Ser. A*, **86**(2), 169–178.

Meier, M.F. 1984. Contribution of small glaciers to global sea level. *Science,* **226**(4681), 1418–1421.

Muravyev, Y.D. 1999a. Mass balance of glacier in condition of maritime climate: Koryto Glacier in Kamchatka, Russia. *In* Naruse, R., *ed. Cryospheric studies in Kamchatka II.* Hokkaido, Hokkaido University. Institute of Low Temperature Science, 51–61.

Muravyev, Y.D. 1999b. Present-day glaciation in Kamchatka: distribution of glaciers and snow. *In* Naruse, R., *ed. Cryospheric studies in Kamchatka II.* Hokkaido, Hokkaido University. Institute of Low Temperature Science, 1–7.

Oerlemans, J. 1994. Quantifying global warming from the retreat of glaciers. *Science,* **264**(5156), 243–245.

Oerlemans, J. 1997. A flowline model for Nigardsbreen, Norway: projection of future glacier length based on dynamic calibration with the historic record. *Ann. Glaciol.,* **24**, 382–389.

Oerlemans, J. 1998. Modelling glacier fluctuations. *In* Haeberli, W., M. Hoelzle and S. Suter, *eds. Into the second century of worldwide glacier monitoring: prospects and strategies.* Paris, UNESCO, 85–96. (Studies and Reports in Hydrology 56.)

Oerlemans, J. and J.P.F. Fortuin. 1992. Sensitivity of glaciers and small ice caps to greenhouse warming. *Science,* **258**(5079), 115–117.

Paterson, W.S.B. 1994. *The physics of glaciers. Third edition.* Oxford, etc., Elsevier.

Schmeits, M.J. and J. Oerlemans. 1997. Simulation of the historical variations in length of Unterer Grindelwaldgletscher, Switzerland. *J. Glaciol.,* **43**(143), 152–164.

Shiraiwa, T. and S. Yamaguchi. 2002. Reconstruction of glacier mass balances and climate changes in the Kamchatka Peninsula. *J. Geogr. [Tokyo],* **111**(4), 476–485.

Shiraiwa, T., Y.D. Muravyev, S. Yamaguchi, G.E. Glazirin, Y. Kodama and T. Matsumoto. 1997. Glaciological features of Koryto glacier in the Kronotsky Peninsula, Kamchatka, Russia. *Bull. Glacier Res.,* **15**, 27–36.

Shiraiwa, T. *and 8 others.* 2001. Characteristics of a crater glacier at Ushkovsky volcano, Kamchatka, Russia, as revealed by the physical properties of ice cores and borehole thermometry. *J. Glaciol.,* **47**(158), 423–432.

Solomina, O.N. 1999. Gornoye oledeneniye Severnoy Evrazii v golotsne [Mountain glaciation of Northern Eurasia in the Holocene]. Moscow, Nauchniy Mir.

Solomina, O. and P.E. Calkin. 2003. Lichenometry as applied to moraines in Alaska, U.S.A., and Kamchatka, Russia. *Arct. Antarct. Alp. Res.,* **35**(2), 129–143.

Yamaguchi, S., R. Naruse, S. Sugiyama, T. Matsumoto and Y.D. Muravyev. 2003. Initial investigations of dynamics of the maritime Koryto glacier, Kamchatka, Russia. *J. Glaciol.,* **49**(165), 173–178.

Zuo, Z. and J. Oerlemans. 1997. Numerical modelling of the historic front variation and the future behaviour of the Pasterze glacier, Austria. *Ann. Glaciol.,* **24**, 234–241.

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