Effect of precipitation seasonality on climatic sensitivity of glacier mass balance

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Abstract

Numerical calculations are described, aimed at evaluating the influence of precipitation seasonality and seasonality concentration on climatic sensitivity of glacier mass balance. Equilibrium line altitudes (ELAs) are modeled using idealized meteorological variables, and then a warming test (+1 °C) is performed, which revealed that the effects of latitude and the annual precipitation amount are less than those of precipitation seasonality and its concentration. Calculation shows higher sensitivities for the glaciers located in a summer precipitation climate than for those located within a winter precipitation climate. Difference due to seasonality is enhanced with the seasonality concentration. The present study suggests that the earlier parameterization of glacier mass balance with annual precipitation amount is insufficient to describe the response of glaciers to climate change. Distribution of precipitation seasonality and its concentration suggest that many parts of the world have a summer-precipitation climate but their seasonality concentration is not significant worldwide. On the other hand, an extensive region of Asia and part of the Andes are located in the highly concentrated summer-precipitation climate. This suggests the high sensitivity of glacier mass balance in these regions to climate change. **Citation:** Fujita, K. (2008), Effect of precipitation seasonality on climatic sensitivity of glacier mass balance, *Earth Planet. Sci. Lett.., 276*, 14-19, doi:10.1016/j.epsl.2008.08.028.

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1. Introduction

Despite the relatively small amounts of ice stored as non-polar glaciers against the total amount of land ice on earth, these glaciers are thought to have had a significant effect on sea-level rise during the last century, since they have higher mass balance sensitivity than the large ice sheets of Greenland and Antarctica. Using glacier mass-balance data, Meier (1984) estimated about half of the sea-level rise was attributable to the shrinkage of non-polar glaciers. Meier's approach was based on the assumption that the thinning rate of glaciers was proportional to the annual balance amplitude, which is approximately equivalent to the annual precipitation on the glaciers. It is necessary to extrapolate the scarce mass-balance measurements to large areas and to a century time scale. Later studies summarized the climatic sensitivity of glacier mass balance (how much mass will be wasted against +1 °C warming) as a function of the annual precipitation on individual glaciers, and confirmed Meier's idea (e.g. Oerlemans and Fortuin, 1992; Braithwaite and Raper, 2002; Raper and Braithwaite, 2006).

Despite the efforts made in subsequent studies to obtain more precise estimates based on the observational data (Dyurgerov and Meier, 1997a; 1997b; Cogley and Adams, 1998; Kaser et al., 2006), major uncertainties still remain due to the serious lack of measurement data for the Asian region. Since the region includes an extensive glacier area (119 × 10^3 km^2, 18% of the total area of non-polar glaciers (Dyurgerov and Meier, 1997b)), any uncertainty will generate considerable estimation errors. In addition, previous numerical studies have applied their models, mainly validated with glaciers in Europe and the Americas, to glaciers worldwide (Oerlemans and Fortuin, 1992; Braithwaite and Raper, 2002; Raper and Braithwaite, 2006). A few studies using a numerical approach, on the other hand, have suggested higher glacier mass balance climatic sensitivity in the Asian highlands (Ageta and Higuchi, 1984; Fujita and Ageta, 2000), though their calculations were performed on specific glaciers in the Himalayas and Tibet. They concluded that snowfall during summer could, with warming, fall as rain, and the decrease in fresh snow would then reduce the glacier surface albedo, leading to greater ablation from the glacier. Because many glaciers are located within a summer accumulation climate pattern, not only in Asia but also in other regions such as South America, it is meaningful to evaluate the effect of precipitation seasonality on glacier mass balance and its climatic sensitivity. Fujita (2008) showed that the precipitation seasonality (summer-type vs. winter-type) significantly affects the climatic conditions and sensitivities of glacier mass balance using a heat-balance model with a simple meteorological pattern. Warming on the summer-type glaciers not only prolongs the melting period, but also causes a significant decrease in snow accumulation and an additional decrease in surface albedo due to a diminished summer snowfall that prevents the absorption of solar radiation and snow melt during the melt season (Fujita, 2008). In order to describe more comprehensively the influence of precipitation seasonality on the climatic sensitivity of the glacier mass balance, further calculations are performed here by adding the other factors of precipitation seasonality (other seasons and 'concentration of seasonality') to the calculation in Fujita (2008).

2. Mass-balance model

The energy-mass balance model used calculates the daily heat balance at the glacier surface, including radiation balance, sensible and latent turbulent heat fluxes, heat conduction into the glacier, and mass balance consisting of snow accumulation, melt, refreezing and evaporation:

$$\max[Q_M;0] = (1-\alpha)R_S + R_L - \min[\sigma T_S^4;315.6] + Q_S + E_V l_e + Q_G \quad . \qquad (1)$$

Heat for melting (Q_M) is obtained if the right-hand side of the equation is larger than zero. Absorbed short-wave radiation is calculated from surface albedo (α) and downward short-wave radiation (R_S). Downward long-wave radiation (R_L) is calculated from air temperature, relative humidity and the ratio of downward short-wave radiation to that at the top of atmosphere using an empirical scheme (Kondo, 1994). Upward long-wave radiation is obtained by the Stefan-Boltzmann constant (σ) and surface temperature in Kelvin (T_S) assuming black body for the snow/ice surface. A melting surface (0 °C surface temperature) releases an upward long-wave radiation of 315.6 (W m^{-2}). Sensible (Q_S) and latent ($E_V l_e$) turbulent heat fluxes are obtained by bulk methods. l_e is the latent heat for evaporation of water or ice, which is determined from the surface temperature. Conductive heat into the glacier ice (Q_G) is obtained by calculating the temperature profile of the snow layer and/or glacier ice. Absorption of short-wave radiation in snow and ice, which increases the temperature of the snow and ice, is taken into account (Fujita and Ageta, 2000). All heat components are positive when fluxes are directed toward the surface. Mass balance (B) at any location on the glacier is calculated as:

$$B = Ca - Q_M / l_m + E_V + R_F \quad . \qquad (2)$$

Solid precipitation (Ca, positive sign), which is determined along with air temperature, is equivalent to accumulation over the glacier. Mass is removed from the glacier as meltwater (Q_M / l_m, positive sign) and evaporation (E_V, negative sign). l_m is the latent heat for melting ice. A part of the meltwater is fixed to the glacier by refreezing (R_F, positive sign), if the glacier ice is cold enough (Fujita and others, 1996). The refreezing amount is calculated in the model by considering the conductive heat into glacier ice and the presence of water at the interface between the snow layer and glacier ice (Fujita and Ageta, 2000). Refreezing during winter and shorter cooling events are also calculated. Special attention should be paid to treatment of the surface albedo (α) as it varies enormously in space and time even on a single glacier (albedo declines down a glacier and during the course of the melt season). The albedo in the model was calculated according to the surface snow density, which changes with snow compaction. This treatment guarantees the feedback effect of albedo in climatic change experiments. Validation of the model was performed for a Tibetan glacier, considering parameters such as ice temperature, seasonal change in levels of surface snow (accumulation and ablation) and ice (meltwater refreezing), albedo at different altitudes, and altitudinal profile of mass balance (Fujita and Ageta, 2000). Detailed schemes have been described by Fujita and Ageta (2000) for the entire model and by Fujita (2007) for albedo and snow densification.

3. Simple meteorological input

Since our purpose is not to examine specific glaciers, but to evaluate the influence of the main precipitation seasons on glacier mass balance, we set up an ideal meteorological input. Daily air temperature has clear seasonality assuming the Northern Hemisphere, which described as:

$$T_d = v_a - A_a \sin\left[(d+91)\times\frac{2\pi}{365}\right] - I_w \cos\left[(d+91)\times\frac{2\pi\times 52}{365}\right] \quad (3)$$

Here, daily air temperature (T_d) at a given date of year (d) is described by its annual mean air temperature (v_a) with a seasonality defined by an annual range of temperature (A_a). A weekly periodic pattern suggesting a mesoscale atmospheric circulation is described with amplitude (I_w). This equation gives the dates of maximum and minimum temperature at the end of June and December. The annual range of temperature is assumed as 10 °C, with a 20 °C difference between maximum and minimum temperatures (Figure 1a). Transmissivity of solar radiation, relative humidity and wind speed (Figure 1b) are assumed as:

$$v_d = v_a + I_w \cos\left(d\times\frac{2\pi\times 52}{365}\right) \quad . \quad (4)$$

Here, a given daily variable (v_d) on a given day of the year (d) is described by its annual average (v_a). A weekly periodic pattern suggesting a mesoscale atmospheric circulation is described with amplitude (I_w). Multiplying transmissivity by solar radiation at the top of atmosphere (Figure 1c) gives the solar radiation at the surface. These equations give simple periodic patterns by values summarized in Table 1 as shown in Figure 1.

Daily precipitation is given as:

$$P_d = \frac{P_r}{\sum_{year} P_r} P_a$$
$$P_r = \max\left(\left\{I_a \cos\left[(d-d_{\max})\times\frac{2\pi}{365}\right] + I_w \cos\left[(d-d_{\max})\times\frac{2\pi\times 52}{365}\right]\right\};0\right) \quad . \quad (5)$$

Here, multiplying daily precipitation ratio ($P_r / \sum_{year} P_r$) by annual precipitation (P_a, m w.e.) gives daily precipitation (P_d). Precipitation seasonality is described by the date of precipitation maximum ($d_{\max}$). Seasonality concentration is given by a parameter (I_a). Precipitation pattern also has weekly fluctuation. Some examples are shown in Figure 2.

4. Sensitivities of mass balance to warming

First, we calculate the equilibrium line altitude (ELA, altitude where annual accumulation and ablation are equal) by changing air temperature (0.3 °C step, assuming 50 m altitude interval) under a given latitude and precipitation pattern, which is defined by seasonality, concentration and annual amount. The annual cycle starts from October. Calculated results of the fifth year are obtained for the stabilization of glacier ice temperature. Second, air temperature is everywhere increased by +1 °C in the sixth year, and the mass balance at the original ELA is obtained as a measure of climatic sensitivity. Much fine tuning was necessary in previous studies to obtain specific mass balances (averaged for the whole glacier area) (e.g. Oerlemans and Fortuin, 1992; Gregory and Oerlemans, 1998; Braithwaite and Raper, 2002;

Raper and Braithwaite, 2006). Any change in the altitudinal distribution of precipitation, for instance, will seriously alter the glacier mass balance (Oerlemans and Fortuin, 1992). Spatial distributions and fluctuations in meteorological variables within a certain region, however, are not fully understood worldwide. On the other hand, conditions at the ELA are close to the average for the whole glacier. Thus, the conditions and climatic sensitivity to warming of the ELA are dealt with as indices of glacier mass balance in this study. Because a combination of weekly peaks of precipitation and air temperature significantly affects the calculated results, the 7-day averages of output are discussed hereafter. Weekly patterns of relative humidity and transmissivity are also changed consistently with that of precipitation (e.g. high humidity and low transmissivity with precipitation peak and vice versa). This assumption is based on a favorable correlation between precipitation and transmissivity, which was found in the observational data in/around Tibet (Matsuda et al., 2006). Although the lower humidity does not always mean higher transmissivity in the real world, the effect of a different combination of relative humidity and transmissivity is not significant in primary calculations (less than 0.1 m w.e. as the sensitivity). We do not discuss the combination of wind speed because its variability did not significantly affect the mass balance (Fujita and Ageta, 2000; Fujita et al., 2007).

Figure 3 presents the sensitivity calendar of ELAs to uniform warming. The effects of latitude (a), annual precipitation (b) and concentration (c) are depicted with the date of maximum precipitation (abscissa). Figure 3a shows the higher sensitivity at the lower latitude in a given seasonality. Since warming causes albedo lowering at any condition, stronger solar radiation increases more absorption of solar radiation and thus more melting at lower latitude. Figure 4a shows that the changes in positive degree-day sum by warming are not significant in association with the latitude. This implies that the sensitivities of glacier mass balance are affected not by the difference of increased temperature, but by the increased absorption of solar radiation in terms of latitude effect. The greater the annual precipitation, on the other hand, the higher the sensitivity in a given seasonality (Fig. 3b). This supports the basic idea in previous studies on the sensitivities of glacier mass balance (Meier, 1984; Oerlemans and Fortuin, 1992; Raper and Braithwaite, 2006). Oerlemans and Fortuin (1992) explained the reason why glaciers in a humid climate (humid-glaciers) are more sensitive rather than those in an arid climate (arid-glaciers). A significant fraction of the rainfall in annual precipitation will increase when temperature increases on the humid-glaciers, while this is not significant on arid-glaciers. In addition, 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. The relation implies that, for a given rise in annual temperature, the increase in melt is larger when the melt rate in the unperturbed state is higher. Therefore, mass-balance gradients, which imply sensitivities of glacier mass balance to warming, tend to become larger in humid-glaciers (Oerlemans and Fortuin, 1992). The calculations of this study lend support to the notion that the changes in positive degree-day sum by warming are the largest in association with the annual precipitation (Fig. 4b) rather than those with the other factors (Figs. 4a and 4c) (Oerlemans and Fortuin, 1992).

At the same latitude (Fig. 3a) or under the same annual precipitation (Fig. 3b), however, the differences among sensitivities associated with seasonality are significantly obvious. Different pattern of the positive degree-day (Fig. 4) against the sensitivity (Fig. 3) suggests that the degree of temperature increase is not always a significant parameter controlling the sensitivity of glacier mass balance. By comparing summer and winter seasonality in the same conceptual modeling, Fujita (2008) explained that warming on the winter-type glaciers only prolonged the melting period without changing snowfall in summer. On the other hand, warming on the summer-type glaciers not only prolonged the melting period, but also caused a significant decrease in snow accumulation and an additional decrease in surface albedo due to a diminished summer snowfall that prevents the absorption of solar radiation and snow melt during the melt season. Fujita (2008) also found that the contributions of decreased accumulation and lowered albedo, both of which were caused by altered precipitation phase (snow to rain), changed with annual precipitation. In a humid environment, a significant decrease in snow accumulation on the glaciers located within a summer accumulation pattern directly caused higher sensitivities. In an arid environment, on the other hand, the decreased summer snow induced accelerated melting by lowering the surface albedo and thus increasing absorption of solar radiation on the glaciers located within a summer accumulation pattern (Fujita, 2008).

In the numerical simulations, the effect of the concentration of precipitation seasonality is examined (Fig. 3c). The figure shows a significant alternation of climatic sensitivity attributed to the seasonality concentration. Without seasonality ($I_a = 0$), for instance, a difference associated with precipitation seasonality unsurprisingly does not appear. Obvious seasonality, however, enhances the contrast between summer and winter even under the same annual precipitation environment. Discrepancy among patterns of sensitivity and change in positive degree-day sum suggests that the change in surface albedo and accumulation more effectively alter the sensitivities than air temperatures. This simulation strongly suggests that the parameterization of climatic sensitivity of glacier mass balance in the previous studies does not sufficiently describe the primary feature of glacier responses to climate change. The sensitivities obtained using observational data, in which seasonality might be involved, should include the feature described in this study. Approximation with a linear regression, however, could compensate for these contrasts due to the biased number of glaciers examined in the Euro-American regions.

5. Global precipitation seasonality

The simulation suggests that the precipitation seasonality and its concentration strongly affect the sensitivity of glacier mass balance. In order to describe the present climate condition associated with precipitation seasonality, we examine the precipitation seasonality index (PSI) and the seasonality concentration index (SCI) by analyzing monthly data of precipitation from GPCP (Adler et al., 2003) and of surface air temperature from NCEP/NCAR (Kalnay et al., 1996). Both data are averaged for the period from 1981 to 2000. PSI is defined as the difference between the hottest month and the month of maximum precipitation. No difference and the maximum difference (six month) in PSI imply the summer and winter precipitation climate, respectively. Spring or autumn is not distinguished. SCI is defined as the standard deviation in the monthly precipitation divided by the annual average monthly precipitation for avoiding the effect of annual amount. Even if the same standard deviations were obtained such as 10 mm, for instance, seasonal variabilities could differ depending on their monthly averages (such as 10 mm or 100 mm). The maximum SCI is defined as 3.3 (only one month precipitation throughout a year).

Figure 5 shows PSI (upper) and SCI (bottom) of the average condition from 1981 to 2000. Distribution of PSI suggests that almost regions where glaciers potentially exist, are located under the summer-precipitation climate. It is climatologically plausible because of the higher humidity in the warmer environment. On the other hand, the western part of central Asia, the equatorial Andes, Patagonia, northern New Zealand, together with the eastern to southern parts of Greenland, are obviously located under the winter-precipitation climate. The distribution of SCI shows more regional features. Almost all regions show lower SCI, implying less seasonality concentration. A high SCI zone including glaciers appears only over the Himalayas, Tibet and Mongolia which are affected by the Asian monsoons. A part of southern Andes is also located under a higher concentration climate. This strongly suggests that the glaciers in these regions should have high sensitivities of glacier mass balance to temperature change, which have been mentioned with some observational data in the Himalayas (Fujita et al., 1997; 2001) and in the Andes (Kaser, 1999; Ceballos et al., 2006).

6. Conclusions

The climatic sensitivities of glacier mass balance were correlated only with annual precipitation in the previous studies by Oerlemans and Fortuin (1992) and Raper and Braithwaite (2006). Annual precipitation affects the temperature condition on the ELA (Ohmura et al., 1992; Fujita, 2008), and thus warming prolongs the melt season more effectively in a humid environment as shown in Fig. 4b (Oerlemans and Fortuin, 1992). Those authors validated their approaches by examining real glaciers worldwide in which both summer-type and winter-type glaciers might be included. However, their parameterization of the sensitivities with annual precipitation should neglect the effect of precipitation seasonality, which was involved in the observational data, because of the great number of examined glaciers sampled from the Euro-American climate. Fujita (2008) showed that the precipitation seasonality, more than the amount, significantly affected the climatic sensitivities of glacier mass balance using the same conceptual model. In addition to his study, we show that the concentration of seasonality more obviously alters the climatic sensitivity of glacier mass balance. The present study suggests that the parameterization of glacier mass balance with annual precipitation amount is insufficient to describe the response of glaciers to climate change. Distribution of precipitation seasonality (PSI) and its concentration (SCI) suggest that many parts of the world belong to the summer-precipitation climate but their seasonality concentration is not significant worldwide. On the other hand, the extensive region of Asia and a part of the Andes are located in the highly concentrated summer-precipitation climate. It suggests the high sensitivity of the glacier mass balance in these regions to climate change. The contribution of glacier shrinkage worldwide or climatic implication of glacier fluctuation has been examined by the numerical approach, in which mainly the effects of annual or seasonal changes in air temperature were discussed. The present study shows, however, that the changes in precipitation seasonality and seasonality concentration also alter the condition of glacier mass balance.

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Table 1. Values of variables used in the calculation.

Variables	Annual average (v_a)	Weekly amplitude (I_w)
Air temperature (°C)	–	3.0
Transmissivity of solar radiation	0.45	0.3
Relative humidity (%)	75	10
Wind speed (m s^{-1})	4.0	1.6
Precipitation ratio	–	10

Figure Legends

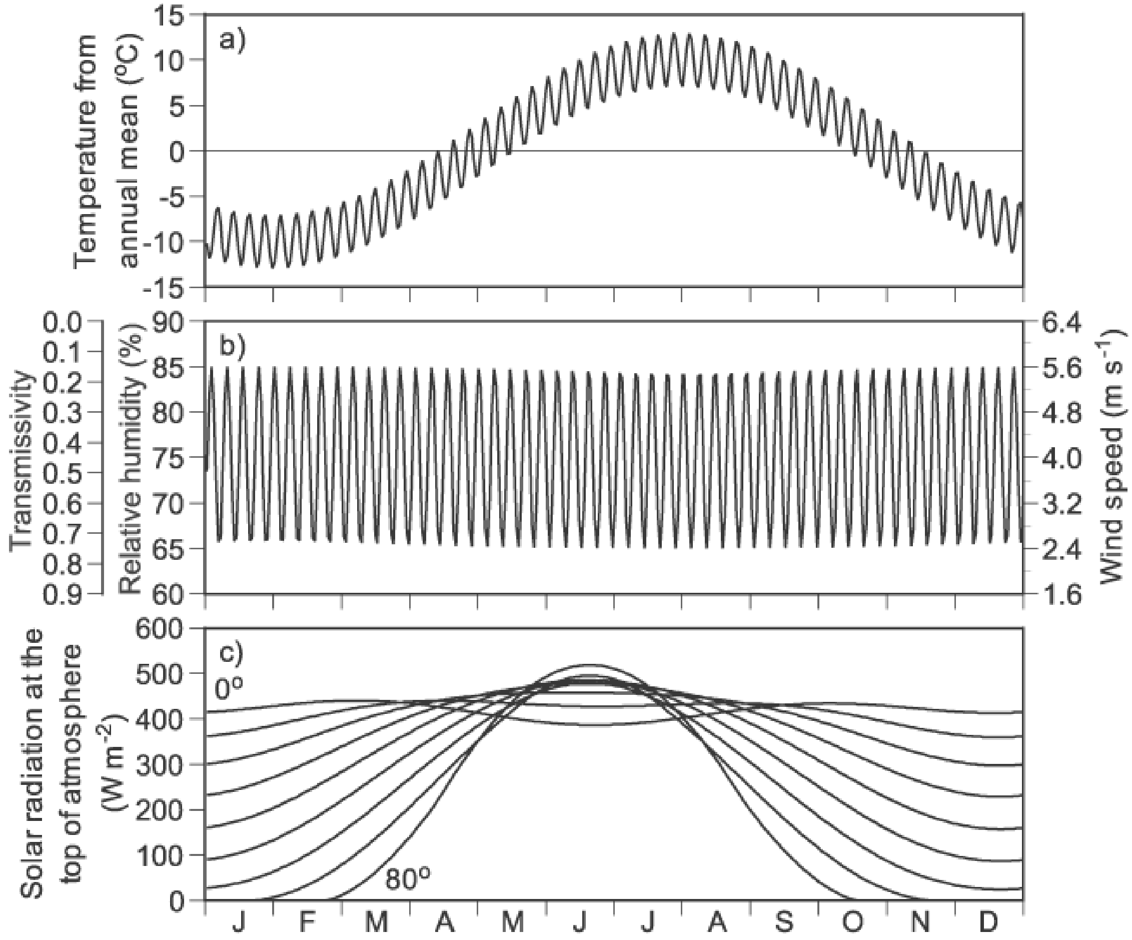
Fig. 1. Simplified meteorological variables for the model: (a) air temperature; (b) wind speed, relative humidity and transmissivity for solar radiation, expressed by a single curve with different ordinate axes; and (c) solar radiation at the top of the atmosphere. Multiplying solar radiation at the top of the atmosphere, which is obtained at each latitude, by the transmissivity gives the solar radiation at the surface.

Fig. 2. Examples of precipitation pattern. Left panels show changes in the date of maximum precipitation (d_{max} = 91, 182, 273, 364, top to bottom) with a given seasonality concentration (I_a = 4). Right panels show changes in seasonality concentration (I_a = 0, 4, 8, 12, top to bottom) with a given date of the maximum precipitation (d_{max} = 182).

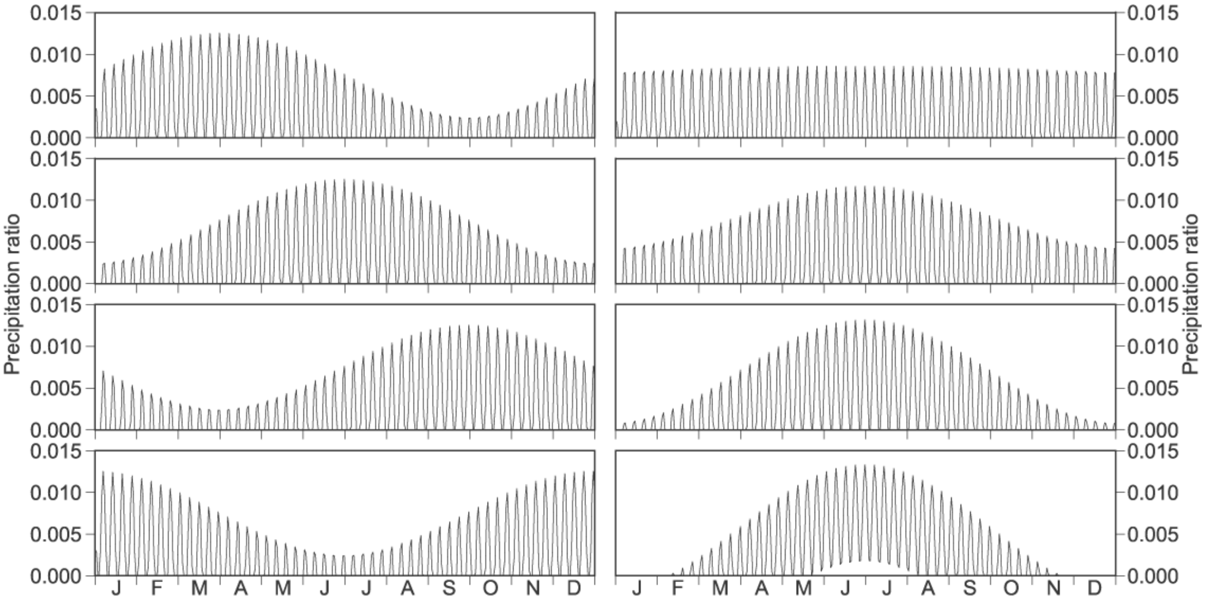
Fig. 3. Climatic sensitivity of glacier mass balance to uniform warming (+1 °C). a) Effect of latitude under a given annual precipitation (2 m w.e.) with a given concentration (I_a = 4). b) Effect of annual precipitation with a given concentration (I_a = 4) at a given latitude (40°). c) Effect of seasonality concentration under a given annual precipitation (2 m w.e.) at a given latitude (40°). Abscissa denotes the date of maximum precipitation in all panels.

Fig. 4. Change in positive degree-day sum at the ELA to uniform warming (+1 °C). a) Effect of latitude under a given annual precipitation (2 m w.e.) with a given concentration (I_a = 4). b) Effect of annual precipitation with a given concentration (I_a = 4) at a given latitude (40°). c) Effect of seasonality concentration under a given annual precipitation (2 m w.e.) at a given latitude (40°). Abscissa denotes the date of maximum precipitation in all panels.

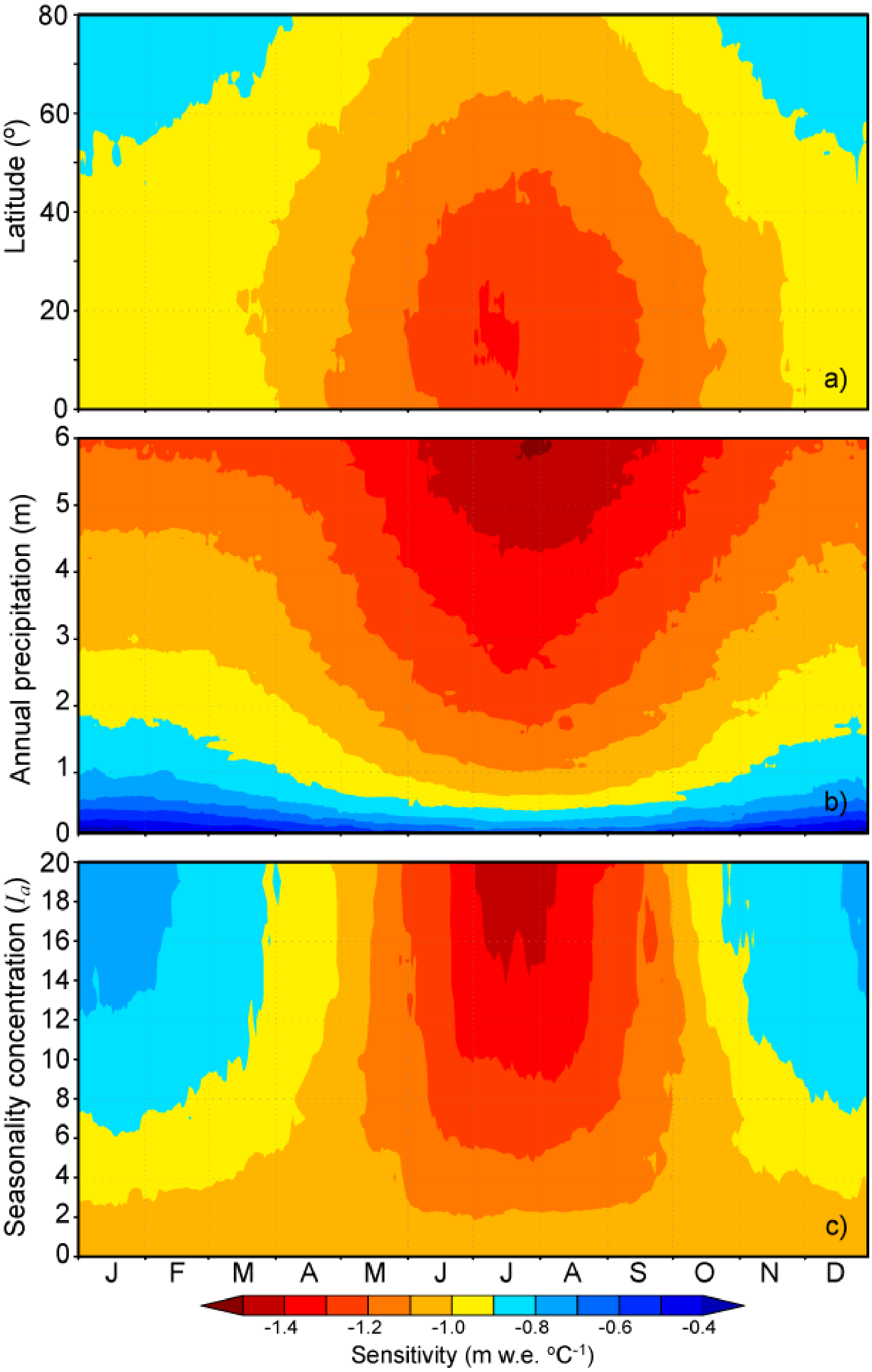
Fig. 5. Distribution of precipitation seasonality index (PSI, upper) and seasonality concentration index (SCI, lower). See text for the definition.



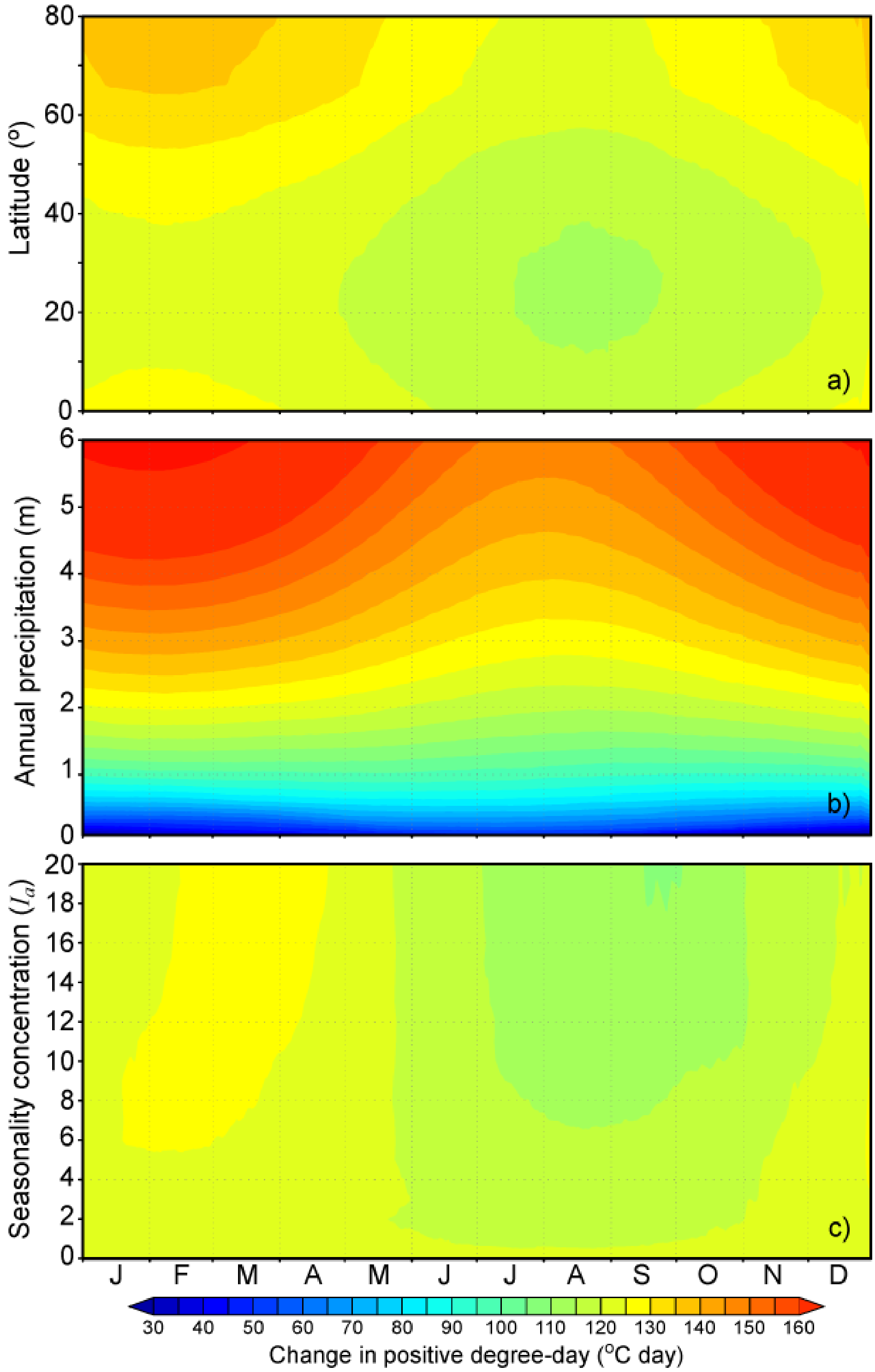
a)
Temperature from annual mean (ºC)
15
10
5
0
-5
-10
-15
b)
Transmissivity
0.0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
0.8
0.9
Relative humidity (%)
90
85
80
75
70
65
60
Wind speed (m s-1)
6.4
5.6
4.8
4.0
3.2
2.4
1.6
c)
Solar radiation at the top of atmosphere (W m-2)
600
500
400
300
200
100
0
0º
80º
J F M A M J J A S O N D



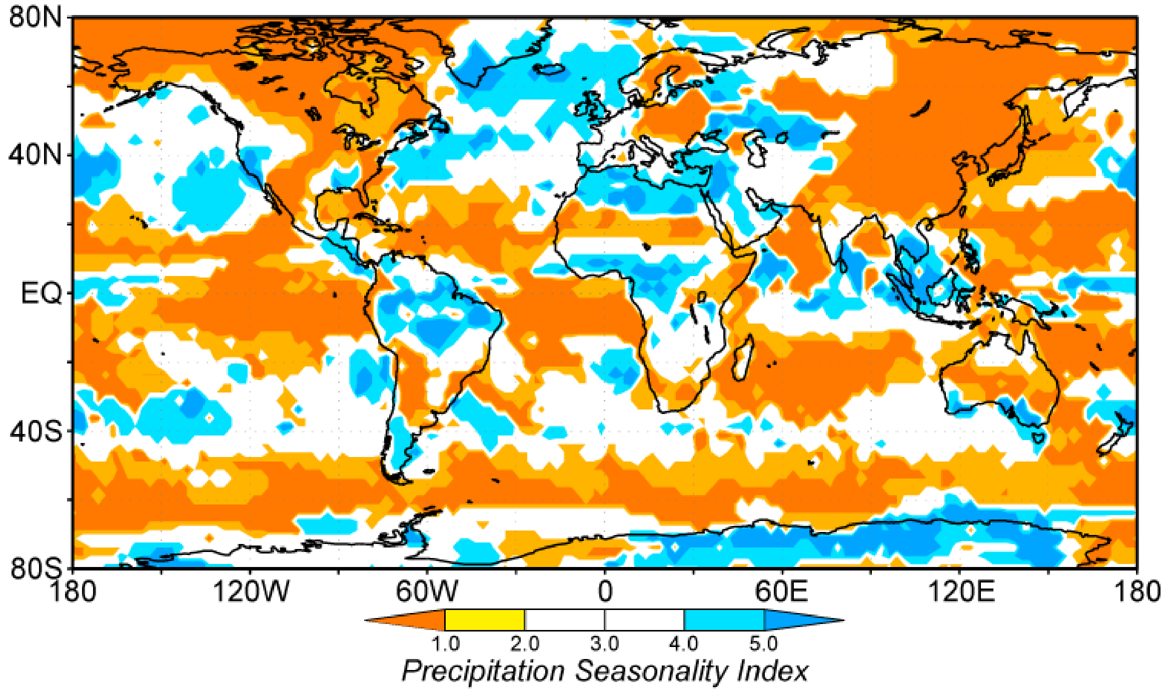
0.015
0.010
0.005
0.000
Precipitation ratio
J
F
M
A
M
J
J
A
S
O
N
D

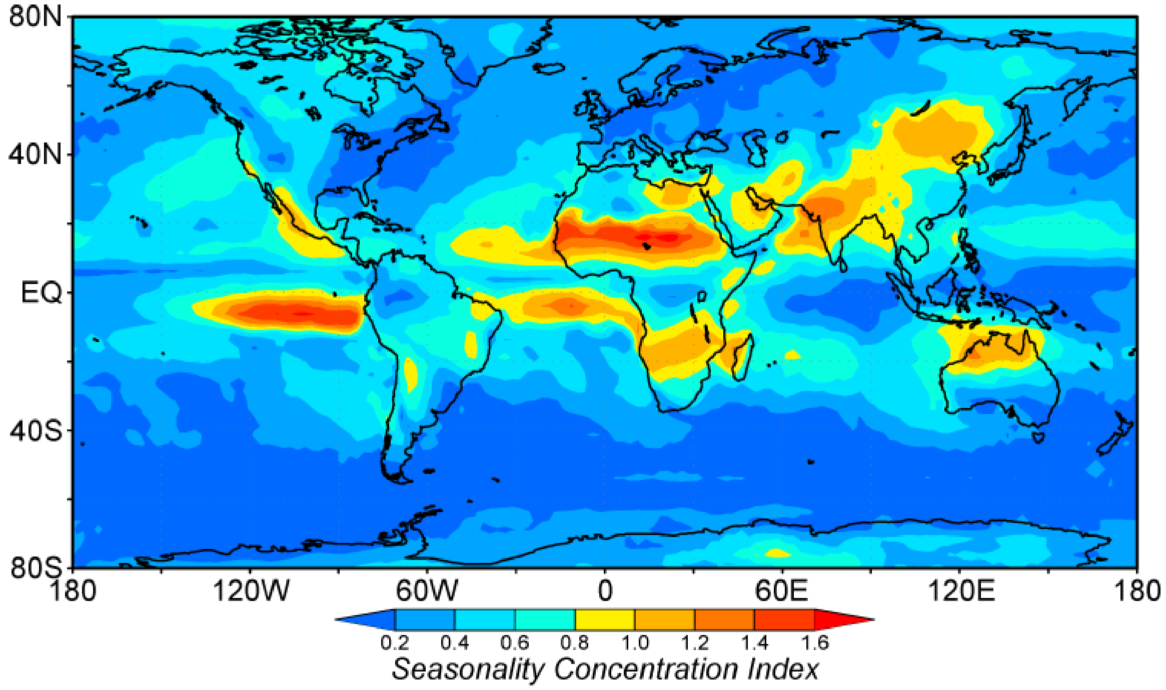


80
60
40
20
0
Latitude (°)
a)
6
5
4
3
2
1
0
Annual precipitation (m)
b)
20
18
16
14
12
10
8
6
4
2
0
Seasonality concentration (I_a)
c)
J F M A M J J A S O N D
-1.4 -1.2 -1.0 -0.8 -0.6 -0.4
Sensitivity (m w.e. °C^{-1})



Latitude (°)
80
60
40
20
0
a)
Annual precipitation (m)
6
5
4
3
2
1
0
b)
Seasonality concentration (I_a)
20
18
16
14
12
10
8
6
4
2
0
c)
J F M A M J J A S O N D
30 40 50 60 70 80 90 100 110 120 130 140 150 160
Change in positive degree-day (°C day)


80N
40N
EQ
40S
80S
180
120W
60W
0
60E
120E
180
1.0
2.0
3.0
4.0
5.0
Precipitation Seasonality Index


80N
40N
EQ
40S
80S
180
120W
60W
0
60E
120E
180
0.2
0.4
0.6
0.8
1.0
1.2
1.4
1.6
Seasonality Concentration Index