Physical Basis for the Temperature-Based Melt-Index Method

ATSUMU OHMURA

Institute for Climate Research, Swiss Federal Institute of Technology, Zurich, Switzerland

(Manuscript received 17 April 2000, in final form 23 August 2000)

ABSTRACT

The close relationship between air temperature measured at standard screen level and the rate of melt on snow and ice has been widely used to estimate the rate of melt. The parameterization of the melt rate using air temperature usually takes a simple form as a function of either the mean temperature for the relevant period or positive degree-day statistics. The computation provides the melt rate with sufficient accuracy for most practical purposes. Because of its simplicity, it is often called a crude method and is rated as inferior to other more sophisticated methods such as the energy balance method. The method is often used with the justification that temperature data are easily available or that obtaining energy balance fluxes is difficult. The physical process responsible for the temperature effect on the melt rate is often attributed to the sensible heat conduction from the atmosphere. The simulation capacity of the temperature-based melt-index method, however, is too good to be called crude and inferior. The author investigated physical processes that make the air temperature so effective a parameter for melt rate. Air temperature has a more profound influence on melt than previously has been acknowledged. The influence of air temperature through the turbulent sensible heat flux is limited, however. The air temperature information is transferred to the surface mainly through longwave atmospheric radiation, which is by far the most important heat source for melt. Under cloudless-sky conditions, as much as 60% of the atmospheric emission is derived from within the first 100 m and 90% from the first 1 km of the atmosphere. When the sky is overcast with the cloud bottom within the first 1 km, more than 90% originates within this layer between the surface and the bottom of the cloud. When the sky is overcast with the cloud bottom higher than 1 km, the first 1 km of the atmosphere still makes up about 70% of the longwave irradiance at the surface, for which the air temperature measured at standard screen level is the single most influential factor. Wind speed is only weakly correlated with melt rate, because the main energy source for melting is longwave atmospheric radiation, followed by the absorbed global radiation, both of which are independent of the movement of the atmosphere.

1. Introduction

The cryosphere is an important component in the climate system because of the albedo and the latent heat of melt. The melt of snow and ice plays a vital role in the hydrological cycle by contributing to the stream discharge. The melt can be computed by applying the principle of heat conservation, whereby all heat supplying and consuming terms except for the latent heat of melt must be known. Because this is a laborious undertaking and the required information is not always available, the melt is often parameterized simply as a function of the air temperature. The justifications often used by the authors for this type of an index method are 1) its good performance in accuracy despite the simplicity, 2) a wide availability of the air temperature data in contrast to the limited sources of radiative and turbulent fluxes, and 3) the easy spatial interpolation possibility of air temperature. The accuracy of the temperature-based melt-index method is reexamined based on field data and literature. This simple method indeed makes it possible to estimate the rate of the melt with sufficient accuracy for most purposes. The author questioned how this seemingly simplistic method yields such a high accuracy. A physical explanation is presented below that has a more profound foundation than mere computational simplicity or the easy availability of input data.

2. Temperature-based melt-index methods

An attempt to relate the melt rate to air temperature has a long history. Hann (1908) and Ahlmann (1924) used air temperature to explain the altitude of glaciation, whereby the annual or summer mean air temperature was related to the annual ablation. Krenke and Khodakov (1966) examined relationships between the melt rate and short- and long-term mean temperature as well as the positive degree-day (PDD) for the regions of the ex-Soviet Union, Scandinavia, and the Alps, presenting useful empirical equations. This index method is still

Corresponding author address: Atsumu Ohmura, Institute for Climate Research, Swiss Federal Institute of Technology (ETH), Winterthurerstrasse 190, CH-8057 Zurich, Switzerland.
E-mail: ohmura@geo.umnw.ethz.ch

TABLE 1. Comparative energy balance observations for various snow and ice surfaces during the melt (W m^{-2}; for boreal forest, C and M were not separated).

	Net radiation R	Sensible heat flux H	Latent heat flux L_vE	Subsurface heat flux C	Melt M	Source
Sea ice (central polar ocean)	73	−6	−6	−6	−45	Vowinckel and Orvig (1964)
Glacier, Accumulation area (Upper Ice Station I)	20	29	3	−10	−41	Havens (1964)
Glacier, Ablation area (EGIG IV)	91	29	−20	−11	−90	Ambach (1963)
Tundra (Axel Heiberg Is.; interior tundra)	88	−29	−40	−12	−5	Ohmura (1982)
Tundra (Barrow); coastal tundra)	67	−13	−5	−15	−35	Weller and Homgren (1974)
Boreal forest	163	−67	−78	−19		Pugsley (1970)
Midlatitude country (Hokkaido)	52	12	−13	0	−51	Takeuchi et al. (1992)
Midlatitude city (Ottawa)	80	27	−79	0	−28	Gold and Williams (1961)

widely used by hydrologists and glaciologists (Reeh 1989; Lang and Braun 1990; Braithwaite 1995; Hock 1999). The way in which the air temperature is used for estimating the melt varies considerably depending on authors. Ahlmann (1924) and, more recently, Ohmura et al. (1996) simply used the summer mean temperature, but Braithwaite (1995) states that PDD generally gives a better estimate for the melt. Further, Reeh (1989) presents an analytical explanation as to why PDD performs better than mean temperatures.

The standard error of estimate (SEE) of the melt parameterized by the air temperature depends on the duration of the period considered. For the daily total ablation estimated by the daily mean air temperature on the Qamanarsup sermia, southwest Greenland, Braithwaite (1995) reports SEE of 10–20 mm day^{-1} which corresponds to about 35% of the mean value. The SEE for monthly total ablation rate calculated by PDD for Qamanarsup sermia and Nordbogletscher is 145 mm month^{-1} (13% of the means; R. J. Braithwaite 1999, personal communication). For the summer total ablation based on the summer mean air temperature (June–August), Ohmura et al. (1996) gives 70 mm (2.5% of the means). The data for the last work is obtained at five locations on the western slope on the Greenland ice sheet. The accuracies reported by these and other field experiments (e.g., Lang 1978) are sufficiently high for most practical objectives. There are, however, further efforts to improve the temperature-based melt-index method. Lang (1978) finds small improvements by incorporating additional variables such as global radiation and water vapor pressure. Curiously, wind speed has no significant influence on the melt rate. Lang (1968) and Lang and Braun (1990) report a better simulation capacity of the glacier melt with the air temperature measured outside the glacier rather than on the glacier. Hock (1999) tries to narrow the range of the error in estimating the melt distribution over an entire glacier by introducing additional variables such as direct radiation.

Whether the mean temperature or PDD is chosen, the melt rate on snow and ice can be calculated with an acceptable accuracy that is sufficient for practical purposes. The success of this type of approximation depends on an appropriate choice of the melt-to-temperature sensitivity factor (mm w.e. °C^{-1}, where w.e. is water equivalent) or the PDD factor (mm w.e. °C^{-1}). Because the melt rate is a term in the surface energy balance equation in which air temperature is involved in several terms, all energy balance information is summarily included in these factors. The variability of the PDD factor in time and space is analyzed qualitatively by Lang and Braun (1990) and quantitatively by Braithwaite (1995). An interesting finding of the latter work is the almost constant nature of the degree-day factor against variable wind speed for temperatures between 1° and 5°C, a common temperature range encountered during the melt season.

3. Previous analyses of the melt with energy balance considerations

In explaining energy sources available for the melt, most authors set out by describing the surface energy balance equation in the following manner:

$$R + H + L_vE + C + M = 0, \qquad (1)$$

where R is net radiation, H is sensible heat flux, L_v is latent heat of vaporization, E is evaporation rate, C is conductive heat flow in the subsurface, and M is latent heat for melting. The signs are taken as positive when the term is a heat source for the surface. Table 1 presents the energy fluxes on various surfaces for the melt period expressed in Eq. (1). From this result, a general conclusion can be drawn that the net radiation is usually the main heat source, followed by sensible heat flux. The heat conduction into the subsurface and the melt are usually two significant heat sinks, of which the latter is dominant. The latent heat flux of vaporization L_vE is variable and depends on the vapor pressure of the advecting atmosphere in comparison with the saturation vapor pressure at the melting temperature. This kind of an argument is widely used and often makes a substantial part of the conclusion in the publications concerned with energy balance/melt experiments. It is interesting to note that the net radiation, which is the major heat source, is often poorly correlated with the melt in comparison with the temperature (Braithwaite 1981). It is, however, useful to look at the balance situation in the manner expressed in Eq. (1) and Table 1. One sees, for example, how closely the sum of all terms on the left-hand side approaches 0. This test is used to evaluate the quality of the experiment, although this is by no means an absolute measure, because two or more errors may cancel each other. The energy conservation in the form of Eq. (1) is also used to estimate an experimentally difficult term to measure, such as latent heat of vaporization L_vE. In climate models and discharge forecast schemes, Eq. (1) is also used to determine the melt rate M, based on atmospheric and subsurface terms, the first four terms on the left-hand side. Beyond this kind of usefulness, considerations extracted from Eq. (1) do not offer much fundamental insight. It is basically a bookkeeping exercise. The considerations with Eq. (1) are useless to answer fundamental questions such as how the air temperature influences the melt or how the air temperature, despite the dominance by net radiation, contains so much information for the melt. It is necessary to look at the surface energy balance from a slightly different viewpoint. In this respect, Kuhn (1987) made an important step by treating longwave incoming radiation separately from the net radiation.

4. A thought on the surface energy balance equation

The basic problem of the considerations based on Eq. (1) is partly the fact that the net radiation consists of heterogeneous components. At least four irradiances with different origins and characteristics make up net radiation. If one wishes to use net radiation for understanding surface processes, one must necessarily examine in detail each of the individual components. When individual terms of the net radiation are expressly written, one gains more insight into the processes involved in the surface energy balance:

$$S(1 - a) + L\downarrow - \sigma T^4 + H + L_vE + C + M = 0, \tag{2}$$

where S is shortwave incoming radiation or global radiation, a is surface albedo, $L\downarrow$ is longwave incoming radiation, σ is Stefan–Boltzmann constant, and T is the surface temperature. Out of the seven terms in Eq. (2) there is one term that is fundamentally different from all others. It is the longwave outgoing radiation σT^4, which does not only mean the surface emission, but also provides a very accurate approximation of the longwave outgoing radiation by taking the long wave reflected radiation into account. This term is fundamentally different from the others in that it is totally passive in comparison with the others. It is incapable of altering itself spontaneously but is entirely determinable when all other terms are given. For example, the first term S, global radiation, can change its value when a number of conditions not explicitly written in Eq. (2) change, such as the solar constant, atmospheric transmittance, and cloud cover. At the same time, the changes in other terms in the equation do not necessarily influence S. This term is autonomous with respect to the other terms. Likewise, the sensible heat flux H is heavily dependent on wind speed and the surface roughness, which are not expressed in Eq. (2). One might say that upon determining all other terms the value of H can be known, very much like σT^4 just discussed. This statement is true as an arithmetic argument, but this fact does not mean that all other terms determine H. The circumstance is different for σT^4. This term, or the surface temperature, is determined entirely by other terms, or the climate system behaves ultimately in a manner so as to adjust its surface temperature toward the new equilibrium following the changes in external energy fluxes. There are mutual dependencies of different degrees between various terms in Eq. (2), however. For example, the subsurface heat conduction C is certainly influenced by the surface temperature T, but C is also influenced by the heat conductivity of the subsurface layers and the subsurface temperature profile, which are not directly expressed in Eq. (2). The longwave incoming radiation $L\downarrow$ is greatly influenced by T, because the important portion of $L\downarrow$ comes from the atmospheric layer close to the ground whose temperature and water vapor concentration are influenced by T. However, $L\downarrow$ reserves a certain independence from T, because $L\downarrow$ is partially a product of the temperature and greenhouse gas concentration profiles in the higher atmosphere and cloud layers that are not directly determined by T. The surface temperature T has a profound influence on M, because the melt happens only when the surface temperature reaches 0°C. Although T influences M in this manner, T alone cannot determine the rate of melt. Unlike the terms discussed above, the longwave outgoing radiation is totally determined by all other terms in Eq. (2). The surface emission and, hence, temperature are totally passive because the surface is an infinitesimally thin layer, which does not store heat. It is therefore incapable of influencing other terms by expending or storing energy on its own. It can only respond toward external heat fluxes by adjusting the rate of radiative emission, and hence T, to obtain an equilibrium. Once the surface reaches the melting point, the majority of the energy sources are funneled into the melt because the subsurface heat conduction forms a kind of a barrier to heat flow owing to the poor thermal conductivity of subsurface soil and ice. Under this condition, the melt term turns into an almost passive role very much like σT^4, which is now a constant. This condition makes it easier to predict the melt rate with the atmospheric information. Individual flux terms in Eq. (2) are presented for various zones of glaciers in Table 2 to find out how the melt rate can best be characterized.

5. Energy sources for the melt

For most glaciers, there are three energy sinks and three sources. The largest sink is the longwave outgoing

TABLE 2. Radiation and heat balance for glaciers and snow cover during the melt (W m^{-2}). Values in brackets are percentages of total sources and sinks. Observed period is day.month, year, where month 1 is Jan, etc.

Site	Coordinates	Altitude (MSL)	Global radiation S	Albedo a	Absorbed global radiation $S(1-a)$	Longwave incoming radiation $L\downarrow$	Sensible heat flux H
			Accumulation area				
Upper Ice Station I, Müller Ice Cap	79°41′N, 90°27′W	1530	260	0.79	55 (15)	273 (76)	29 (8)
Ice Cap Station, Devon Is.	75°30′N, 93°18′W	1320	272	0.83	47 (15)	266 (83)	10 (3)
Crest Station, Barnes Ice Cap	70°14′N, 73°55′W	1073	256	0.69	81 (23)	265 (75)	6 (2)
Camp A1, Barnes Ice Cap	69°43′N, 72°13′W	865	147	0.57	63 (18)	280 (79)	13 (4)
Carrefour, EGIG, Greenland	69°49′N, 47°26′W	1850	346	0.85	52 (17)	242 (78)	15 (5)
Penny Ice Cap	66°59′N, 65°28′W	2050	249	0.6	100 (26)	266 (70)	4 (1)
Kesselwandferner	46°59′N, 10°47′E	3240	218	0.72	61 (17)	275 (75)	31 (8)
Ewigschneefeld	46°33′N, 08°02′E	3366	280	0.74	71 (20)	286 (79)	4 (1)
			Near equilibrium line				
Ward Hunt Ice Shelf	83°12′N, 74°00′W	15	207	0.64	72 (20)	301 (80)	0 (0)
Main Ice, Meighen Ice Cap	79°58′N, 99°09′W	241	202	0.74	53 (15)	299 (85)	−4 (−1)
ETH Camp, Greenland	69°35′N, 49°16′W	1155	281	0.77	65 (19)	262 (76)	16 (5)
Blue Glacier	47°48′N, 123°43′W	2010	346	0.59	142 (32)	258 (57)	50 (11)
Vernagtferner	46°50′N, 10°45′E	2973	258	0.37	162 (34)	282 (60)	23 (5)
Hintereisferner	46°48′N, 10°45′E	2960	265	0.59	109 (27)	269 (66)	32 (8)
Rhonegletscher	46°37′N, 08°24′E	2820	278	0.64	99 (23)	256 (59)	81 (19)
No. 1 Glacier, Tenshan	43°06′N, 87°15′E	3910	232	0.56	101 (26)	270 (70)	17 (4)
			Ablation area				
Hans Tausen Ice Cap	82°49′N, 36°13′W	540	222	0.48	115 (27)	277 (66)	27 (6)
Kronprins Christian Land	79°55′N, 24°04′W	380	319	0.48	166 (32)	272 (52)	88 (17)
Lower Ice Station, White Glacier	79°21′N, 90°39′W	208	171	0.45	94 (20)	298 (63)	52 (11)
ditto	ditto	ditto	176	0.42	102 (24)	275 (65)	48 (11)
ditto	ditto	ditto	209	0.39	128 (27)	288 (62)	42 (9)
Sverdrup Glacier, Devon Is.	75°40′N, 83°15′W	300	141	0.49	72 (18)	283 (71)	31 (8)
Camp IV, EGIG, Greenland	69°40′N, 49°38′W	1004	272	0.56	120 (28)	285 (66)	29 (7)
Storglaciären	67°55′N, 18°35′E	1370	149	0.79	31 (9)	282 (80)	27 (8)
	ditto	ditto	172	0.5	89 (21)	278 (67)	49 (12)
Aletschgletscher	46°26′N, 08°04′E	2220	216	0.27	156 (31)	288 (58)	38 (8)

radiation σT^4, and this is almost constant at just below the blackbody emission for 0°C. This slight depression below 315.6 W m^{-2} (blackbody emission at the melting point of ice at normal temperature and pressure) is the result of nighttime refreezing and the subsequent drop of the surface temperature often observed with a clear sky. The second largest sink is the melt. This term is highly variable and is largely dependent on air temperature. The subsurface heat conduction C has some significance only on cold glaciers and is equivalent to about 10% of the heat used for M (Table 2). Because the longwave outgoing radiation $L\uparrow$ is almost constant and the heat conduction C very small, the computation of M can be made only with fluxes above the surface. This is fortunate, because the melt is the most significant ablation process for most glaciers. For snow- and ice-covered hydrological basins, melt is an important contribution to total water discharge. The energy sources are, in order of importance, the longwave incoming radiation $L\downarrow$, absorbed global radiation $S(1-a)$, and sensible heat flux H. The absorbed global radiation accounts for only a quarter the entire heat source. By far the largest energy source is the longwave incoming radiation. The longwave incoming radiation is about 10

TABLE 2. (*Extended*)

Latent heat flux L_vE	Subsurface heat flux C	Heat of melt M	Longwave outgoing radiation σT^4	Blackbody temperature of surface (K)	Observed period	Source
			Accumulation area			
3 (1)	−10 (−3)	−41 (−11)	−309 (−86)	272	5.6–26.8, 1960	Havens (1964)
−4 (−1)	−12 (−4)	−10 (−3)	−297 (−92)	269	1.6–31.8, 1962, 63	Holmgren (1971a,b)
2 (1)	−15 (−4)	−29 (−8)	−310 (−88)	272	5.7–11.8, 1962	Sagar (1966)
0 (0)	0 (0)	−40 (−11)	−315 (−89)	273	25.6–4.8, 1950	Ward and Orvig (1953)
−13 (−4)	−6 (−2)	0 (0)	−285 (−94)	266	1.6–27.7, 1967	Ambach (1977), Ambach and Markl (1983)
12 (3)	0 (0)	−50 (−14)	−306 (−86)	271	13.7–26.7, 1953	Orvig (1954)
−2 (−1)	0 (0)	−45 (−13)	−309 (−87)	272	11.8–8.9, 1958	Ambach and Hoinkes (1963)
−3 (−1)	0 (0)	−45 (−12)	−315 (−87)	273	3.8–19.8, 1973	Lang et al. (1977), Röthlisberger and Lang (1987)
			Near equilibrium line			
0 (0)	0 (0)	−60 (−16)	−316 (−84)	273	60 h in 6, 7, 1960	Lister (1962)
−4 (−1)	−11 (−3)	−35 (−10)	−311 (−85)	272	1.6–31.8, 1960–70	Taylor (1974)
−6 (−2)	−8 (−2)	−30 (−9)	−304 (−87)	271	3.6–31.8, 1991	Ohmura et al. (1994)
−4 (−1)	0 (0)	−131 (−29)	−316 (−70)	273	12.7–30.8, 1958	LaChapelle (1959)
4 (1)	0 (0)	−216 (−42)	−301 (−58)	270	21.8–31.8, 1950	Hoinkes and Untersteiner (1952), Hoinkes (1955)
−3 (−1)	0 (0)	−95 (−23)	−312 (−76)	272	15.7–18.8, 1971	Wagner (1979, 1980), Tanzer (1986)
−2 (<−1)	0 (0)	−167(−35)	−308(−65)	271	1.8–9.9, 1982	Funk (1985)
−15(−4)	0 (0)	−66 (−17)	−309 (−79)	272	1.6–31.8, 1986, 87	Calanca and Heuberger (1990)
			Ablation area			
−24 (−6)	−18 (−4)	−71 (−17)	−309 (−73)	271	2.7–5.8, 1994	Braithwaite et al. (1998)
−36 (−7)	−18 (−3)	−159 (−30)	−316 (−60)	273	8.7–27.7, 1993	Konzelmann and Braithwaite (1995), Braithwaite et al. (1998)
32 (7)	−13 (−3)	−149 (−31)	−317 (−66)	273	8.7–18.8, 1960	Andrews (1964)
1 (<1)	−22 (−5)	−89 (−21)	−315 (−74)	273	12.6–18.8, 1961	Müller and Roskin-Sharlin (1967), Müller and Keeler (1969)
9 (2)	−18 (−4)	−133 (−28)	−316 (−68)	273	16.7–31.7, 1962	Havens et al. (1965)
14 (4)	−11 (−3)	−84 (−20)	−316 (−77)	273	9.7–10.8, 1963	Keeler (1964)
−21 (−5)	−11 (−3)	−90 (−21)	−315 (−72)	273	26.5–7.8, 1959	Ambach (1963)
12 (3)	−1 (<−1)	−49 (−14)	−304 (−86)	271	7.6–17.9, 1993	Hock (2001)
3 (1)	−3 (−1)	−110 (−26)	−309 (−73)	272	5.7–6.9, 1994	Hock and Holmgren (1996)
14 (3)	0 (0)	−181 (−36)	−315 (−64)	273	2.8–27.8, 1965	Lang and Schönbächler (1967), Röthlisberger and Lang (1987)

times larger than sensible heat flux near the equilibrium line altitude. The sensible heat flux loses its significance in the accumulation area because the air temperature is lower and becomes more important in the lower ablation areas where air temperature is higher. Even in the ablation area, the sensible heat is less than 20% of the longwave incoming radiation. Longwave incoming radiation and sensible heat flux together account for three-quarters of the entire energy source for the melt. Both longwave atmospheric radiation and the sensible heat flux are strongly influenced by the air temperature above the glacier, and this is the main reason for the close relationship between the melt and the air temperature.

6. Vertical structure of longwave atmospheric radiation and the melt

Because the atmosphere is only partially transparent in the wavelength of terrestrial radiation, the air temperature at the standard screen level alone cannot determine the total atmospheric emission. To understand the influence of the standard screen level air temperature on the incident longwave radiation, it is necessary to investigate the numerical importance of the emission from each atmospheric layer from the stratosphere down to the earth's surface. This problem was investigated with a radiation model. The radiation model of the mod-

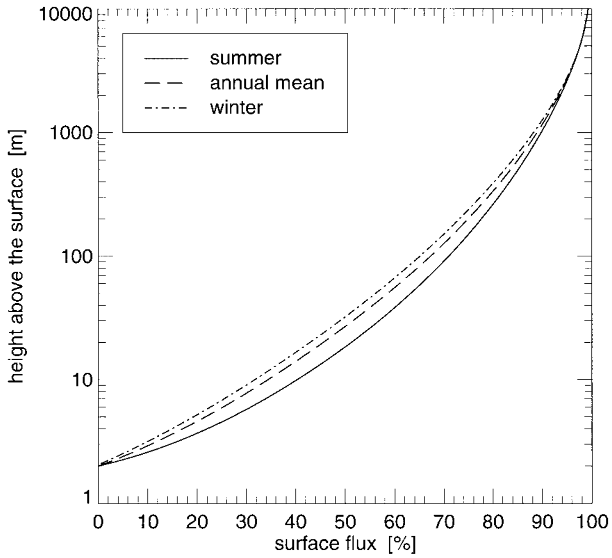

FIG. 1. Origin of longwave atmospheric radiation received at the surface in midlatitudes without clouds showing how the percentage of longwave incoming radiation received at the earth's surface is derived from the layer of the atmosphere whose upper limit is expressed in meters in the ordinate. For example, about 40% of the incoming radiation received at the surface is emitted by the first 10 m of the atmosphere in summer. The irradiance was computed with MODTRAN, using the *U.S. Standard Atmosphere, 1976.* The water vapor profiles were obtained from the mean humidity profiles observed at Payerne (46°49′N, 6°57′E, 491 m above mean sea level) in Switzerland.

erate-resolution transmittance model and code, (MODTRAN; Acharya et al. 1998) and the *U.S. Standard Atmosphere, 1976,* were used. To test the effect of various temperature and moisture stratifications, the midlatitude annual mean winter and summer temperature profiles of the standard atmosphere were used. Because the standard atmosphere does not provide water vapor concentrations, the humidity profiles were obtained from the 10-yr mean profiles (1981–90) from the radiosonde observations (annual mean, January, and July) at Payerne, Switzerland, (491 m above mean sea level), which can be assumed to be representative of midlatitudes. The results of the numerical experiments were presented in Fig. 1. The most important conclusion of this investigation is the finding that as much as 34% of the entire irradiance received at the surface originates within the first 10 m, 67% from the first 100 m, and 89% from the first 1 km of the atmosphere of the annual mean profile without cloud. Looking at the seasonal situations in detail, the first 100 m are responsible for the emission of 65% in winter and as much as 73% in summer. It can be concluded that about 2/3 of the longwave surface irradiance originates within the first 100 m in all seasons. The contribution of the first 1 km fluctuates seasonally between 87% and 90% of the total incoming longwave radiation. This difference is caused mainly by the seasonal change in precipitable water vapor corresponding to 11 and 25 mm for January and July, respectively. Almost 95% of the longwave radiation originates from within the first 2 km of the atmosphere, and at this level seasonal fluctuation becomes insignificant. If a cloud layer is present within this layer, the relative importance of the lower atmosphere further increases. If the cloud layer is present above 1 km, the relative importance of the first kilometer decreases. The influence of overcast conditions with stratocumulus at 2 km above the surface and altostratus at 5 km was examined with the annual mean atmospheric profile. The two experiments concluded that the first 1 km contributed 67% and 73%, respectively. It is, therefore, a thin near-surface layer of the atmosphere that plays an important role in determining the longwave atmospheric irradiance at the earth's surface. This thin layer can be interpreted as the first 1 km. The current result also suggests that the major process in the water vapor/longwave radiation feedback happens in this near-surface layer, and that the countermechanism against the enhanced greenhouse effect through the desiccation of the atmosphere above 3 km proposed by Lindzen (1990) is not relevant. Furthermore, the data from high-altitude sites, which are more relevant for the melt on Alpine-type glaciers, show that 60% of the longwave incoming radiation is emitted from the first 100 m of the atmosphere under the summer atmospheric profiles above Jungfraujoch (3580 m; Marty 2000). The relative importance of the near-surface atmosphere at high altitudes is due to the relative dominance by the emission from the 15-μm carbon dioxide absorption band in place of the diminishing effect of water vapor. In general, the air temperature measured at the standard screen level can be regarded as representative of the first 100 m of the atmosphere. Further, the atmospheric condition above the 100-m level is also highly correlated with that of the first 100 m. This is the reason why the standard screen-level air temperature is so influential to the longwave irradiance at the surface. This is also the main reason for the reliable performance of the empirical formulas based on the one-level temperature/longwave irradiance as proposed by Ångström (1916), Brunt (1932), Swinbank (1963), and many other authors. Brutsaert (1975) proposed essentially a one-level approximation of the longwave incoming radiation, based on the integration of the Schwarzschild equation. Although there is an effort to improve the computational accuracy by taking humidity into account (Ångström 1916; LeDrew 1975), many authors maintain that the air temperature alone is capable of simulating longwave incoming radiation with sufficient accuracy (Swinbank 1963; Ohmura 1981; Saunders and Bailey 1997). Arnfield (1979) rates the performance of the one-level temperature parameterization as equivalent to the accuracy of the pyrradiometer-based measurements of longwave incoming radiation. Because the longwave incoming radiation is by far the most powerful energy source for the melt and is mostly emitted from a thin near-surface layer, it is understandable that the air temperature measured at the standard screen level alone can explain much of the melt rate. Lang (1978) and Braithwaite (1995) found little influence of the wind speed on the melt rate, because the influence of the air temperature enters into Eq. (2) mainly through the longwave atmospheric radiation, which is independent of the movement of the atmosphere. Nevertheless, the air temperature is also an important quantity for determining the surface sensible heat flux, which is usually the third largest energy source for the melt. This situation makes the air temperature even more important for the melt computation. In this context, the finding by Lang (1968) that a better correlation of the melt was found with respect to the air temperature measured outside the glacier and not above the glacier can be recalled. The temperature measured outside the glacier carries substantial information both for longwave incoming radiation and for the advecting sensible heat for the glacier surface. In addition, this kind of air temperature contains the information on the shortwave incoming radiation. The albedo of the ground surrounding glaciers lies usually between 0.1 and 0.2 and is considerably lower than that of the glaciers. This condition allows for the air temperature to capture the information on shortwave radiation, which is the second largest energy source for the melt. Consequently, the air temperature measured outside the glaciers during the summer can be considered to contain physically substantial information for all three energy sources for the melt (Table 2).

7. Conclusions

The parameterization of the melt rate by air temperature provides sufficiently accurate means for most practical purposes. The use of this method is justified more on physical grounds than previously has been assumed. The importance of the air temperature in estimating the melt is mainly due to the fact that the longwave atmospheric radiation is by far the most dominant heat source, and further to the fact that the majority of the atmospheric radiation received at the surface comes from the near-surface layer of the atmosphere (e.g., 90% from the first 1 km without cloud; 70% from the first 1 km under an overcast sky). The surface air temperature measured at the standard screen level is the most influential variable for determining the emission from this thin atmospheric layer. In addition, the secondary and tertiary heat sources such as shortwave radiation and sensible heat fluxes are also correlated with the air temperature. This situation adds to the importance of the air temperature as an index for calculating the melt rate.

Acknowledgments. The author is indebted to Dr. Regine Hock of the Climate Impact Research Centre, Kiruna. She provided the author with unpublished heat balance data for Storglaciären. Dr. Roger J. Braithwaite of the University of Manchester gave the author newly computed prepublication statistics of the melt observed on Qamanarsup sermia and Nordbogletscher, Greenland. The current work was partly financed by the Alliance for Global Sustainability within ETH–MIT–University of Tokyo and a grant from the Swiss National Science Foundation for the Summit Greenland Environment Observatory.

REFERENCES

Acharya, and Coauthors, 1998: *MODTRAN User's Manual, Version 3.7 and 4.0.* Air Force Research Laboratory, 83 pp.

Ahlmann, H. W., 1924: Le niveau de glaciation comme fonction de l'accumulation d'humidité sous forme solide (The level of glaciation as the function of the accumulation of solid precipitation). *Geogr. Ann.,* **6,** 223–272.

Ambach, W., 1963: Untersuchung zum Energieumsatz in der Ablationszone des Grönländischen Inlandeises (Investigation of the energy budget in the ablation area of the Greenland ice sheet) (Camp IV-EGIG, 69°40′05″N, 49°37′58″W). *Expédition Glaciologique Internationale au Groenland 1957–1960.* Vol. 4; *Meddelelser om Grønland.* Vol. 174, 311 pp.

——, 1977: Untersuchung zum Energieumsatz in der Akkumulatioinszone des Grönländischen Inlandeises (Investigation of energy budget in the accumulation area of the Greenland ice sheet). *Expédition Glaciologique Internationale au Groenland 1967–1968.* Vol. 4, *Meddelelser om Grønland.* Vol. 187, 44 pp.

——, and H. Hoinkes, 1963: The heat balance on an Alpine snowfield (Kesselwandferner, 3240 m, Oetztal Alps, August 11–September 8, 1958), *IAHS Publ.,* **61,** 24–36.

——, and G. Markl, 1983: Untersuchungen zum Strahlungshaushalt in der Akkumulationszone des Grömländischen Inlandeises (Investigation of radiation budget in the accumulation area of the Greenland ice sheet) (Station Carrefour 69°49′25″N, 47°25′57″W, 1850 m). *Expédition Glaciologique Internationale au Groenland 1967–1968.* Vol. 4, *Meddelelser om Grønland.* Vol. 187, 61 pp.

Andrews, R. H., 1964: Meteorology and heat balance of the ablation area, White Glacier, Canadian Arctic Archipelago—summer 1960 (Lower Ice Station: 79°26′N, 90°39′W, 208 m). Axel Heiberg Island Research Rep., Meteorology, No. 1, McGill University, Montreal, PQ, Canada, 107 pp.

Ångström, A., 1916: Ueber die Gegenstrahlung der Atmosphäre (On the atmospheric counter-radiation). *Meteor. Z.,* **33,** 529–538.

Arnfield, A. J., 1979: Evaluation of empirical expressions for the estimation of hourly and daily totals of atmospheric longwave emission under all sky conditions. *Quart. J. Roy. Meteor. Soc.,* **105,** 1041–1052.

Braithwaite, R. J., 1981: On glacier energy balance, ablation, and air temperature. *J. Glaciol.,* **27,** 381–391.

——, 1995: Positive degree-day factors for ablation on Greenland ice sheet studies by energy-balance modelling. *J. Glaciol.,* **41,** 153–160.

——, T. Konzelmann, C. Marty, and O. B. Olesen, 1998: Reconnaissance study of the glacier energy balance in North Greenland, 1993–94. *J. Glaciol.,* **44,** 239–247.

Brunt, D., 1932: Notes on radiation in the atmosphere, Part I. *Quart. J. Roy. Meteor. Soc.,* **58,** 389–420.

Brutsaert, W., 1975: On a derivable formula for long-wave radiation from clear skies. *Water Resour. Res.,* **11,** 742–744.

Calanca, P.-L., and R. Heuberger, 1990: Energy balance. *Glacial Climate Research in the Tianshan, Research Report on Project Glacier No. 1, 1985–1937.* Zürcher Geogr. Schriften A. Ohmura, A. H. Lang, F. Blumer, and D. Grebner, Eds. No. 38, Geogr. Inst., ETH, Zurich, Switzerland, 60–73.

Funk, M., 1985: Räumliche Verteilung der Massenbilanz auf dem Rhonegletscher und ihre Beziehung zu Klimaelementen (Spatial distribution of mass balance on the Rhone Glacier and its relation to climatic elements). Zürcher Geogr. Schriften, No. 24, Geogr. Inst., ETH, Zurich, Switzerland, 183 pp.

Gold, L. W., and G. P. Williams, 1961: Energy balance during the snow melt period at an Ottawa site. *IAHS Publ.,* **54,** 288–294.

Hann, J., 1908: *Handbuch der Klimatologie (Handbook of Climatology).* Verlag von J. Engelhorn, 394 pp.

Havens, J. M., 1964: Meteorology and heat balance on the accumulation area, McGill Ice Cap, Canadian Arctic Archipelago—summer 1960 (Upper Ice Station I: 79°41′N, 90°27′W, 1530 m). Axel Heiberg Island Research Rep., Meteorology, No. 2., McGill University, Montreal, PQ, Canada, 80 pp.

——, F. Müller, and G. C. Wilmot, 1965: Comparative meteorological survey and a short-term heat balance study on the White Glacier, Canadian Arctic Archipelago—summer 1962. Axel Heiberg Island Research Rep., Meteorology, No. 4, McGill University, Montreal, PQ, Canada, 68 pp.

Hock, R., 1999: A distributed temperature index ice- and snowmelt model including direct solar radiation. *J. Glaciol.,* **45,** 101–111.

——, 2001: A distributed energy balance model for complex topography and its application to Storgalciären, Sweden. *Water Resour. Res.,* in press.

——, and B. Holmgren, 1996: Some aspects of energy balance and ablation of Storglaciären, northern Sweden. *Geogr. Ann.,* **78,** 121–131.

Hoinkes, H., 1955: Measurements of ablation and heat balance on alpine glaciers. *J. Glaciol.,* **2,** 497–501.

——, and N. Untersteiner, 1952: Wärmeumsatz und Ablation auf Alpengletschern (Heat budget and ablation on Alpine glaciers): I Vernagtferner (Oetztaler Alpen), August 1950. *Geogr. Ann.,* **34,** 99–158.

Holmgren, B., 1971a: Climate and energy exchange on a sub-polar ice cap in summer, Arctic Institute of North America Devon Island Expedition 1961–1963, Part E, Radiation climate. Meddelande No. 111, Department of Meteorology, Uppsala University, Uppsala, Sweden, 53 pp.

——, 1971b: Climate and energy exchange on a sub-polar ice cap in summer, Arctic Institute of North America Devon Island Expedition 1961–1963, Part F, On the energy exchange of the snow surface at Ice Cap Station. Meddelande No. 112, Department of Meteorology, Uppsala University, Uppsala, Sweden, 111 pp.

Keeler, C. M., 1964: Relationship between climate, ablation, and runoff on the Sverdrup Glacier, 1963, Devon Island, N.W.T. AINA Research Paper No. 27, Arctic Institute of North America, 80 pp.

Konzelmann, T., and R. J. Braithwaite, 1995: Variations of ablation, albedo and energy balance at the margin of the Greenland ice sheet, Kronprins Christian Land, eastern north Greenland. *J. Glaciol.,* **41,** 174–182.

Krenke, A. N., and V. G. Khodakov, 1966: O svyazi poverkhnostnogc tayaniya lednikov s temperaturoy vozdukha (On the relationship of surface melt of glaciers with air temperature.) Materialy glyatsiologicheskikh issledovaniy (Data of glaciological studies), Khoronika obsuzhdeniya (Chronicle Discussions), No. 12, 153–164.

Kuhn, M., 1987: Micro-meteorological conditions for snow melt. *J. Glaciol.,* **33,** 24–26.

LaChapelle, E., 1959: Annual mass and energy exchange on the Blue Glacier. Contribution No. 42, Department of Meteorology and Climatology, University of Washington, Seattle, WA, 443–449.

Lang, H., 1968: Relations between glacier runoff and meteorological factors observed on and outside the glacier. *IAHS Publ.,* **79,** 429–439.

——, 1978: Ueber die Bedeutung der Lufttemperatur als hydrometeorologischer Informationsträger (On the significance of air temperature as hydrometeorological information carrier). Arbeiten aus Zentralanstalt für Meteorologie und Geodynamik, No. 31, 23/1–18.

——, and M. Schönbächler, 1967: Heat balance studies and runoff at the Aletschgletscher 1965. Internal Report VAW, ETH, Zurich, Switzerland, 30 pp.

——, and L. Braun, 1990: On the information content of air temperature in the context of snow melt estimation. *Hydrology of Mountainous Areas,* L. Molar, Ed., *IAHS,* 347–354.

——, B. Schädler, and G. Davidson, 1977: Hydrological investigations on the Ewigschneefeld - Gr. Aletschgletscher. *Zeitschr. Gletscherk. Glazialgeol.,* **12,** 109–124.

LeDrew, E. F., 1975: The energy balance of a mid-latitude alpine site during the growing season, 1973. *Arctic and Alpine Res.,* **7,** 301–314.

Lindzen, R. S., 1990: Some coolness concerning global warming. *Bull. Amer. Meteor. Soc.,* **71,** 288–299.

Lister, H., 1962: Heat and mass balance at the surface of the Ward Hunt Ice Shelf, 1960. AINA Sci. Rep. No. 19, Arctic Institute of North America, Montreal, PQ, Canada, 54 pp.

Marty, C., 2000: Surface radiation, cloud forcing and greenhouse effect in the Alps. Dissertation, No. 13609, ETH, Zurich, Switzerland, 167 pp.

Müller, F., and N. Roskin-Sharlin, 1967: A High Arctic climate study on Axel Heiberg Island, Canadian Arctic Archipelago-summer 1961. Part I: General meteorology. Axel Heiberg Island Research Rep., Meteorology, No. 3, McGill University, Montreal, PQ, Canada, 82 pp.

——, and C. M. Keeler, 1969: Errors in short-term ablation measurements on melting ice. *J. Glaciol.,* **8,** 91–105.

Ohmura, A., 1981: Climate and energy balance on arctic tundra: Axe Heiberg Island, Canadian Arctic Archipelago, spring and summer 1969, 1970, and 1972. Zürcher Geographische Schriften, No. 3, Swiss Federal Institute of Technology (ETH), Zurich, Switzerland, 448 pp.

——, 1982: Climate and energy balance on the Arctic tundra. *J. Climatol.,* **2,** 65–84.

——, T. Konzelmann, M. Rotach, J. Forrer, M. Wild, A. Abe-Ouchi, and H. Toritani, 1994: Energy balance for the Greenland ice sheet by observation and model computation. *Snow and Ice Covers; Interactions with the Atmosphere and Ecosystems,* H. G. Jones, T. D. Davies, A. Ohmura, and E. M. Morris, Eds., IAHS 85–94.

——, M. Wild, and L. Bengtsson, 1996: A possible change in mass balance of Greenland and Antarctic ice sheets in the coming century. *J. Climate,* **9,** 2124–2135.

Orvig, S., 1954: Glacial–meteorological observations on icecaps in Baffin Island. *Geogr. Ann.,* **36,** 197–318.

Pugsley, W. I., 1970: The surface energy budget of central Canada. Publication in Meteorology, No. 96, Department of Meteorology, McGill University, Montreal, PQ, Canada, 58 pp.

Reeh, N., 1989: Parameterization of melt rate and surface temperature on the Greenland ice sheet. *Polarforschung,* **59,** 113–128.

Röthlisberger, H., and H. Lang, 1987: Glacial Hydrology. *Glaciofluvial Sediment Transfer,* A. M. Gurnell and M. J. Clark, Eds., John Wiley and Sons, 207–284.

Sagar, R. B., 1966: Glaciological and climatological studies on the Barnes Ice Cap, 1962–64. *Geogr. Bull.,* **8,** 3–47.

Saunders, I. R., and W. G. Bailey, 1997: Longwave radiation modeling in mountainous environments. *Phys. Geogr.,* **18,** 37–52.

Swinbank, W. C., 1963: Long-wave radiation from clear skies. *Quart. J. Roy. Meteor. Soc.,* **89,** 339–348.

Takeuchi, Y., Y. Kodama, N. Ishikawa, and D. Kobayashi, 1992: The effect of snow cover on the thermal characteristics of the surface layer. *Low Temperature Science, Ser. A,* **51,** 63–76.

Tanzer, G., 1986: *Berechnung des Wärmehaushalts an der Gleichgewichtslinie des Hintereisferner* (Computation of the heat budget on the equilibrium line of the Hintereisferner). Diplomarbeit, University of Innsbruck, Innsbruck, Austria, 103 pp.

Taylor, B., 1974: The energy balance climate of Meighen Ice Cap, N. W. T., Vol. 1. Ph.D. thesis, Department of Meteorology, McGill University, Montreal, PQ, Canada, 66 pp.

Vowinckel, E., and S. Orvig, 1964: Energy balance of the Arctic, V. The heat budget over the Arctic Ocean. *Arch. Meteorol. Geophys. Bioklimatol., Ser. B,* **14,** 303–325.

Wagner, H. P., 1979: Strahlungshaushaltsuntersuchungen an einem

Ostalpengletscher während der Hauptablationsperiode, Teil I: Kurzwellige Strahlung (Investigation of radiation budget on a glacier in the eastern Alps during the main ablation period, Part I: Shortwave radiation). *Arch. Meteorol. Geophys. Bioklimatol., Ser. B,* **27,** 297–324.

——, 1980: Strahlungshaushaltsuntersuchungen an einem Ostalpengletscher während der Hauptablationsperiode, Teil II: Langwellige Strahlung und Strahlungsbilanz (Investigation of radiation budget on a glacier in the eastern Alps during the main ablation period, Part II: Longwave radiation and net radiation). *Arch. Meteorol. Geophys. Bioklimetol, Ser. B,* **28,** 41–62.

Ward, W. H., and S. Orvig, 1953: The heat exchange at the surface of the Barnes Ice Cap during the ablation period. *J. Glaciol.,* **2,** 158–168.

Weller, G., and B. Holmgren, 1974: The microclimates of the Arctic tundra. *J. Appl. Meteor.,* **13,** 854–862.