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New estimations of precipitation and surface sublimation in East Antarctica from snow accumulation measurements

Abstract Surface mass balance (SMB) distribution and its temporal and spatial variability is an essential input parameter in mass balance studies. Different methods were used, compared and integrated (stake farms, ice cores, snow radar, surface morphology, remote sensing) at eight sites along a transect from Terra Nova Bay (TNB) to Dome C (DC) (East Antarctica), to provide detailed information on the SMB. Spatial variability measurements show that the measured maximum snow accumulation (SA) in a 15 km area is well correlated to firn temperature. Wind-driven sublimation processes, controlled by the surface slope in the wind direction, have a huge impact (up to 85% of snow precipitation) on SMB and are significant in terms of past, present and future SMB evaluations. The snow redistribution process is local and has a strong impact on the annual variability of accumulation. The spatial variability of SMB at the kilometre scale is one order of magnitude higher than its temporal variability (20–30%) at the centennial time scale. This high spatial variability is due to wind-driven sublimation. Compared with our SMB calculations, previous compilations generally over-estimate SMB, up to 65% in some areas.

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1 Introduction

Precipitation over Antarctica is recognised as an important climate variable. Snow accumulation (SA) rate or surface mass balance (SMB) on the Antarctic Plateau is the sum of precipitation, sublimation/deposition and wind-blown snow. Large gaps in observations mean that any estimate of the current mass input involves a large error factor (Genthon and Krinner 2001; Rignot and Thomas 2002). Spatial SMB is known to vary greatly (e.g. Richardson et al. 1997; van den Broeke et al. 1999; Frezzotti et al. 2002a; Frezzotti et al., in press). Representative observations of the SMB are important in estimating the characteristics of spatial and temporal variability at local scales (< 10 km^2) and at the scale of a drainage basin. Snow redistribution changes the topography, and the topography in turn alters the wind field in a feedback system between the cryosphere and atmosphere. On the local scale, there is continual interaction between processes such as wind, snow precipitation, sublimation and SMB variations; in particular, the surface-energy balance and katabatic wind patterns are closely inter-related.

Antarctica is the highest and flattest of the Earth's continents, but small changes in slope have a strong impact on wind direction and speed (Frezzotti et al. 2002a). It has long been known that slope and curvature can play an important role in the SMB; for example, concave depressions accumulate snow at the expense of convex rises (Black and Budd 1964; Whillans 1975; Pettré et al. 1986; van den Broeke et al. 1999; Liston et al. 2000; Frezzotti et al. 2002a). A large area of the plateau, where the slope along the wind direction is

higher than 4 m km^{-1}, has a nil or slightly negative SMB (Frezzotti et al. 2002b).

Surface mass balance studies based on numerical weather prediction models (e.g. Bromwich 1988; Genthon and Braun 1995; Turner et al. 1999) yield results which are comparable to those based on the interpolation of observations, but the horizontal resolution of the meteorological fields is not better than ~100 km and is affected by systematic biases (Gallée et al. 2001; Genthon and Krinner 2001). It is known that sublimation in Antarctica is not negligible (e.g. Stearns and Weidner 1993; Bintanja 1998; Gallée 1998; Gallée et al. 2001), and methods for describing spatial and temporal variability must be further developed (Cullather et al. 1998). The snowdrift process is not explicitly included in numerical weather prediction and general circulation models (Gallée et al. 2001; Genthon and Krinner 2001). One of the largest areas of uncertainty regarding SMB is the role of surface and wind-driven sublimation (e.g. Genthon and Krinner 2001; Turner et al. 2002). The fraction of precipitation on the grounded ice which is returned to the atmosphere through sublimation ranges from 6 to 25% in high-resolution models (van der Broeke 1997; Genthon and Krinner 2001). Van der Broeke (1997) pointed out four areas (east Dronning Maud Land, Lambert Glacier basin, Victoria Land and the southern part of Ross Ice Shelf) clearly stand out where sublimation on an annual basis removes more than 70% of the precipitation. Some authors even note that, due to small-scale spatial variability in net SMB, it is impossible to determine 'true' precipitation or SMB in large sectors of the Antarctic (Turner et al. 1999).

Alley (1988) found that a depth hoar layer lost 25% of its mass, probably by migrating upward to the atmosphere; even larger mass loss during sublimation is documented by Alley et al. (1990). In wind crust area, very light winds (<2 m s^{-1}) could be sufficient to initiate snowdrifts on this smooth surface (Bintanja et al. 2001), particularly in the case of fresh snow falls. Wind scouring does not allow the SA (due to surface smoothness) during the winter period, while low albedo of some snow surfaces promotes sublimation during summer. Both factors have a positive feedback effect on the development of permanent wind crust.

At Vostok Station, accumulation stake network allowed determining a mean accumulation rate of 23 ± 4.4 kg m^{-2} a^{-1} and a precipitation rate of 26 ± 4.5 kg m^{-2} a^{-1} for the last 30 years. The total sublimation, for the period November–February, varies from 4.4 to 2 kg m^{-2} a^{-1} of total annual precipitation (Ekaykin et al., in press). King et al. (2001) have pointed out that, at Halley Station, surface and blowing sublimation make roughly equal contributions to ablation (about 82 kg m^{-2} a^{-1}). Bintanja (1999) shows that there is a clear positive correlation between sublimation and air temperature in blue ice areas. Stearns and Weidner (1993) estimated a net annual surface sublimation of 35 and 80 kg m^{-2} a^{-1}, respectively at the South Pole Station and Ross Ice Shelf, with a SMB of 80 and

100 kg m^{-2} a^{-1}, respectively. Kobayashi et al. (1985) and Takahashi et al. (1988) estimated that the precipitation rate at Mizuho Stations (East Antarctica) was between 140 and >200 kg m^{-2} a^{-1}. The measured surface sublimation was 54 kg m^{-2} a^{-1}, with a SMB of 58 kg m^{-2} a^{-1} determined by strong sublimation in summer (Fujii and Kusunoki 1982). The deficit between precipitation and surface sublimation at Mizuho Stations can be explained by blowing sublimation (Takahashi et al. 1988). We define the blowing sublimation as the sublimation of the snow blown into the air.

Observation of the SMB rate variability and the study of redistribution processes not only provide input for the mass balance estimation but are also essential for better interpreting surface elevation change signals from satellite altimeters (Rémy et al. 2002) and for improving climate and meteorological models (e.g. Cullather et al. 1998; Genthon and Krinner 1998, 2001; Turner et al. 1999; Gallée et al. 2001). Interpretation and dating of palaeoenvironmental records extracted from Antarctic ice cores depend on the knowledge of past accumulation changes (e.g. Bromwich and Weaver 1983; Jouzel et al. 1983). Atmospheric/ocean warming in the coming century may lock up greater volumes of ocean water due to increased precipitation above the Antarctic ice sheet. The SMB trend on a century scale is required to evaluate Antarctica's present mass balance and how it may change over the coming decades and centuries.

As part of the International TransAntarctic Scientific Expedition (ITASE) project (Mayewski and Goodwin 1999) and in the framework of the Franco-Italian Concordia Station collaboration (between 1998 and 2000), a traverse between Terra Nova Bay (TNB) and Dome C (DC) and a research at DC were undertaken (Fig. 1). The study aimed to better understand latitudinal and longitudinal environmental gradients, while documenting climatic, atmospheric and surface conditions over the last 200–1,000 years in the eastern and north-eastern portions of the DC drainage area and in northern Victoria Land.

Frezzotti et al. (in press) combine geophysical surveys (GPR and GPS), field and remote sensing surface observations and firn core analyses to describe spatial and temporal variability of SA along the traverse and at DC. This paper uses these results to provide new information on the SMB process and carries implications for SMB distribution and variability.

2 Methodology

Different methods were used, compared and integrated (stake farms, ice cores, snow radar, surface morphology, remote sensing) at eight sites along a transect from TNB to DC (Fig. 1) to evaluate the SA variability. Three cores were drilled 5–7 km apart at each of the seven main sites along the traverse (23 total cores), while at DC 16 shallow firn cores were drilled within a 25-km radius from

Fig. 1 Schematic map of traverse from TNB to DC and site location (**a**), surface elevation (*solid line*) and micro-relief distribution along TNB–DC traverse. Core sites are indicated with *triangles* (**b**)
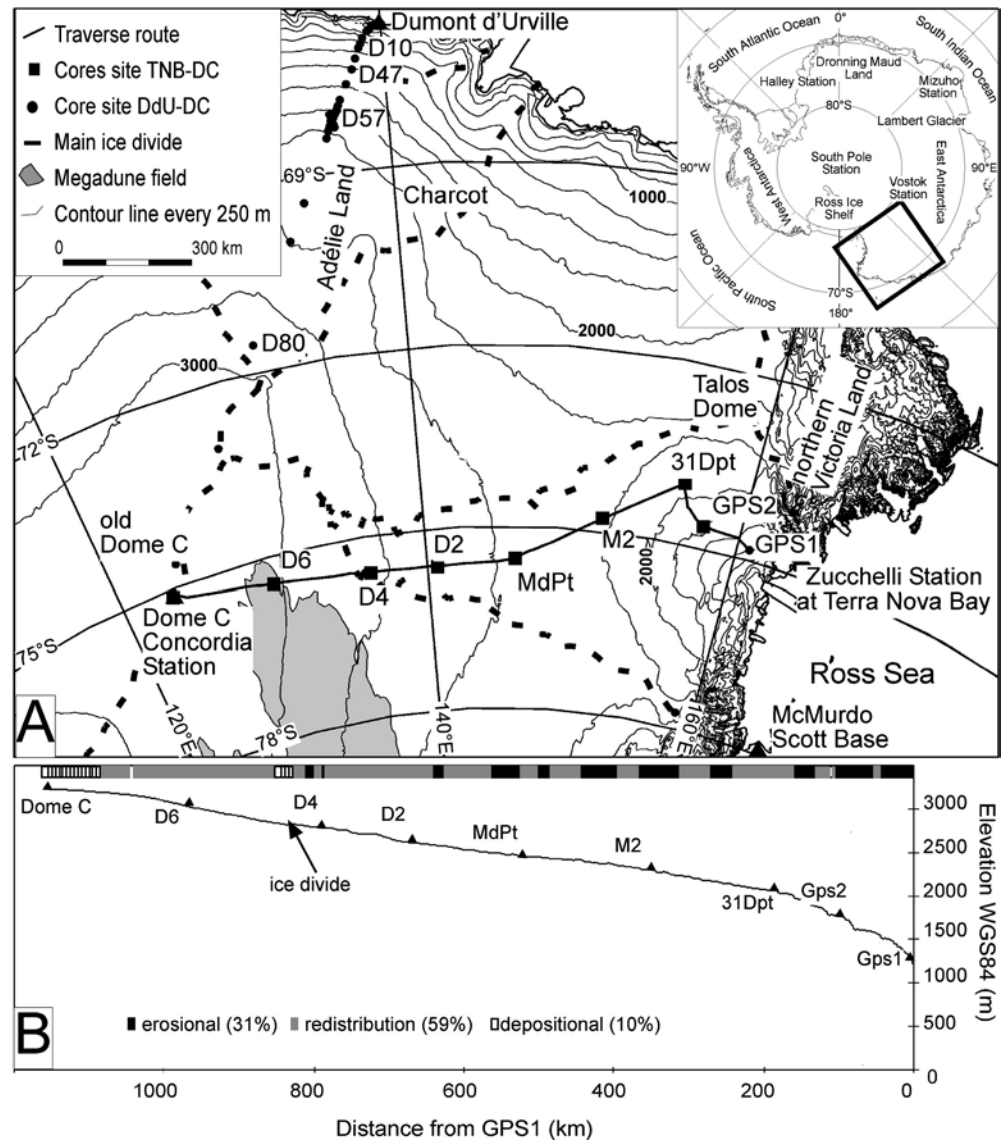
Concordia Station. Cores were linked by snow radar and GPS surveys to provide detailed information on spatial variability in SMB. The 39 cores were dated by identifying Tritium/β marker levels (1965–1966) to provide information on SA at core site. The depth of the snow radar layer was converted into SMB (GPR_SA) using the depth/age ratio from the Tritium SA. The mean depth and the relative standard deviation (standard deviation/average of site) were computed for GPR_SA, and the minimum and maximum accumulations were recorded at each site. SA data derived from core records are in good agreement with data derived from GPR_SA (Table 1). The survey methods and analytical procedures for core and snow radar measurements are described in Frezzotti et al. (in press). We also use in this paper the SA data (41 cores) that were collected along the transect between Dumont d'Urville (DdU) and DC (Pourchet et al. 1997) using the β radioactive horizons in the cores.

The snow temperature measured at a depth of 10 or 15 m gives a fairly close approximation to the mean annual "surface temperature" in the dry snow, and is also close to the mean annual "screen air temperature" (Loewe 1970). This assumption is acceptable in areas where maximum temperature is below 0°C during summer time, as it is the case in the actual sampling sites. Snow temperature profiles were measured at core sites (Frezzotti and Flora 2002; Pourchet et al. 1997), and 15 m depth temperatures were used in this paper.

3 Results and discussion

3.1 Surface mass balance

In this chapter, we first analyse the spatial variability of the SA. Then we discuss the evaluation of the precipitation and ablation values from the SA ones and compare our results to existing accumulation maps.

Table 1 Location, morphological and climatological characteristic of drill sites

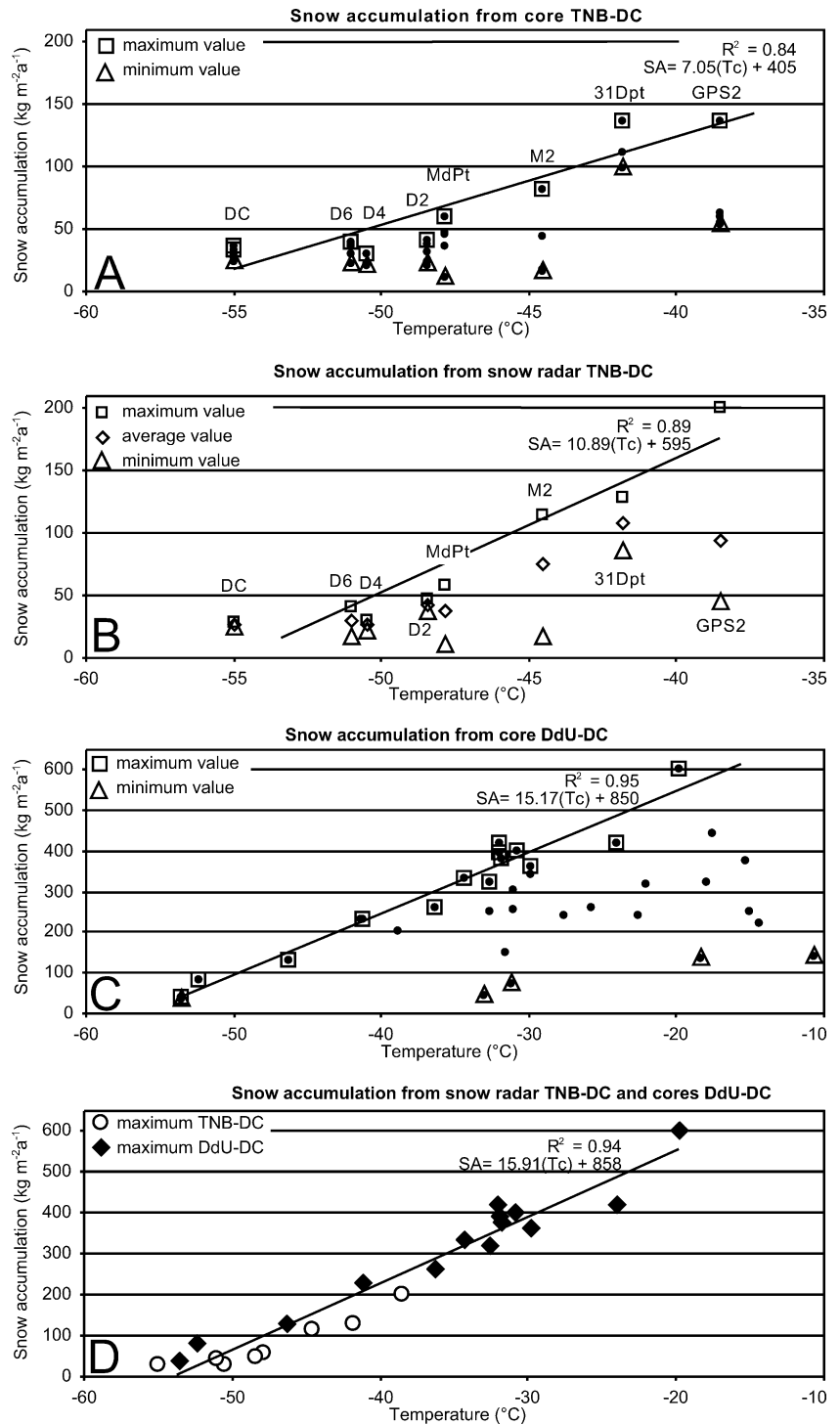| Site	Longitude E	Latitude S	Elevation (m WGS84)	T −15m (°C)	Slope direction	Wind direction	Katabatic wind speed (m s^{-1})	β1966–1998 β1955–1998$^+$ Tritium 1966–1998* (kg m^{-2} a^{-1})	GPR_SA (kg m^{-2} a^{-1})
GPS1							>18		
Core A	160°39.60′	74°48.94′	1,192	−32.7				58 ± 2.9*	–
Core B	160°39.60′	74°48.94′	1,192	–				47 ± 4.7	–
GPS2							>18		
Core A	157°30.13′	74°38.69′	1,804	−38.5	132°	80°		54 ± 2.7*	54 (27%)
Core B	157°22.80′	74°36.81′	1,810	–	132°	80°		137 ± 13.7	200 (100%)
Core C	157°30.13′	74°38.69′	1,804	–	132°	80°		62 ± 6.2	54 (27%)
31Dpt							12–18		
Core A	155°57.6′	74°01.52′	2,069	−41.8	158°	72°		112 ± 5.6*	112 (77%)
Core B	156°0.64′	74°03.51′	2,040	–	158°	72°		137 ± 13.7*	145 (100%)
Core C	155°55.99′	74°03.80′	2,041	–	158°	72°		–	97 (67%)
M2							12–18		
Core A	151°16.17′	74°48.27′	2,278	−44.5	162°	62°		15 ± 7.5*	15 (14%)
Core D	151°09.13′	74°49.86′	2,265	–	111°	62°		82 ± 8.2	107 (100%)
Core C	151°05.91′	74°48.01′	2,272	–	247°	62°		44 ± 4.4	61 (57%)
MdPt							12–18		
Core A	145°51.43′	75°32.16′	2,454	−47.8	78°	32°		45 ± 2.7*	45 (60%)
Core B	145°47.31′	75°33.03′	2,460	–	78°	32°		11 ± 1.1	13 (18%)
Core C	145°55.35′	75°31.74′	2,452	–	78°	32°		60 ± 6.0	75 (100%)
D2							6–12		
Core A	140°37.84′	75°37.33′	2,479	−48.4	89°	37°		31 ± 1.6*	31 (94%)
Core B	140°28.61′	75°38.76′	2,482	–	89°	37°		40 ± 4.0$^+$	33 (100%)
Core C	140°28.54′	75°36.06′	2,483	–	89°	37°		38 ± 3.8	29 (87%)
D4							6–12		
Core A	135°49.89′	75°35.88′	2,793	−50.5	116°	42°		–	22 (79%)
Core B	135°40.43′	75°37.13′	2,795	–	116°	42°		29 ± 2.9	28 (100%)
Core C	135°40.74′	75°34.43′	2,798	–	116°	42°		20 ± 2.0	22 (72%)
D6							6–12		
Core A	129°48.53′	75°26.85′	3,027	−51.0	97°	46°		29 ± 1.4*	29 (85%)
Core B	129°42.41′	75°25.19′	3,035	–	97°	46°		38 ± 3.8	34 (100%)
Core C	129°38.04′	75°26.42′	3,038	–	97°	46°		22 ± 2.2	16 (46%)
Dome C						40°	<6		
EPICA-DC1	123°20.86′	75°06.06′	3,233						
FIRETRAC	123°20.86′	75°06.06′	3,233	−54.5					
DCN	123°18.72′	75°07.32′	3,233	−55.0				29 ± 1.5*	
DC-1 A17	123°36.26′	75°00.53′	3,233					26 ± 1.3*	
DC-2 E16	123°01.95′	75°02.86′	3,230					27 ± 1.4*	
DC-3 D10	123°11.13′	75°11.68′	3,233					26 ± 1.3*	
DC-4 C12	123°45.51′	75°09.29′	3,229					23 ± 1.2*	
DC-A18	123°49.85′	74°54.55′	3,226					32 ± 3.2	
DC-DORIS	123°05.92′	75°09.16′	3,232					33 ± 3.3	
DC-B11	124°15.97′	75°05.99′	3,225					29 ± 2.9	
DC-F10	123°11.30′	75°00.50′	3,230					29 ± 2.9	
DC-DAU	122°57.04′	75°17.72′	3,233					29 ± 2.9	
DC-E19	122°38.83′	74°59.31′	3,223					30 ± 3.0	
DC-A17	123°36.26′	75°00.53′	3,232					29 ± 2.9	
DC-D11	122°57.45′	75°17.71′	3,230					27 ± 2.7	
DC-A15	122°56.66′	75°17.71′	3,233					30 ± 3.0	
DC-C17	123°23.65′	75°12.52′	3,229					26 ± 2.6	
DC-C18	123°49.83′	75°17.51′	3,221					28 ± 2.8	

Snow accumulation from stake farms and core analyses, and comparison with snow radar depths (water equivalent) in percent. GPR_SA depth of the snow radar layer converted into SMB using the depth/age ratio from the 'core A' core Tritium snow accumulation. Katabatic wind speed from Parish and Bromwich (1991)

3.2 Snow precipitation vs. surface mass balance

The maximum value of SMB revealed by snow radar (along 15 km) and core data along the traverse presents an excellent correlation with firn temperature (Fig. 2a; $R^2 = 0.84$ core data; Fig. 2b: $R^2 = 0.89$ snow radar). The average core data value is not well correlated with firn temperature ($R^2 = 0.37$). Core and snow radar minimum values show no significant correlation with elevation or temperature ($R^2 = 0.31$ and 0.32), but values are linked with the change of Slope along the Prevalent Wind Direction (SPWD). Noone et al. (1999) found that precipitation reflects large (synoptic) scale phenomena connected to circulation on a global scale.

Fig. 2 Snow accumulation vs. temperature from core analysis (**a**) ($R^2 = 0.84$, significance level > 95%) and from snow radar (**b**) ($R^2 = 0.89$, significance level > 95%) along the TNB–DC traverse. SA vs. temperature from core analysis (**c**) ($R^2 = 0.95$, significance level > 99%) along DdU–DC traverse (Pourchet et al. 1997) and maximum SA vs. temperature along TNB–DC and DdU–DC (**d**) ($R^2 = 0.94$, significance level > 99%). Snow temperature was measured at 15 m depth

Snow accumulation from core TNB-DC

Snow accumulation from snow radar TNB-DC

Snow accumulation from core DdU-DC

Snow accumulation from snow radar TNB-DC and cores DdU-DC

Several authors (e.g. Robin 1977; Muszynski and Birchfield 1985; Fortuin and Oerlemans 1990; Giovinetto et al. 1990) suggest that the SMB is directly correlated with temperature, elevation, saturation vapour pressure and distance from the open ocean (continentality). However, the firn temperature is the only ground field climatological datum available in most of Antarctica. Young et al. (1982) pointed out that for the Pionerskaya DC line the smooth accumu-

lation trend followed temperature very closely. Frezzotti and Flora (2002) and Magand et al. (in press) show that the spatial variability in the local climate in this area may be unrelated to altitude and distance from the open ocean. No single process accounts entirely for the firn temperature patterns. However, we suspect that the interplay of katabatic winds, storm intrusions and the winter temperature inversion driven by the local and regional morphology (slope, convex or

concave surface) and surface albedo (wind crust, sa-strugi, etc.) may be very important.

On the basis of our observations, we suggest that the maximum SA rate value at each site, which is well correlated with the mean annual temperature ($R^2 = 0.89$, $n = 8$), represents a value close to snow precipitation, according to the following function:

$$SA = 10.89(Tc) + 595 \tag{1}$$

where SA is the maximum SA (kg m^{-2} a^{-1}) and Tc is the firn temperature at 15 m (°C).

This hypothesis is confirmed by the fact that snow radar and satellite image analysis along the traverse reveal that only megadune and some occasional transversal dunes represent depositional features. Frezzotti et al. (2002a) pointed out that megadunes have a wave formed by a variable net accumulation, ranging from 25% (lee-ward slopes) to 120% (windward slopes) of the snow accumulated in adjacent non-megadune areas. All other sites show no clear accumulation morphology at the km scale, but a surface morphology prevalently determined by ablation processes (Frezzotti et al. 2002a, b). The analysis of $\delta^{18}O$, performed every 5 km along the traverse (1 m core), shows a good correlation ($R^2 = 0.83$, $n = 227$) between $\delta^{18}O$ values and site temperatures with a slope of 0.99‰°C^{-1} (Proposito et al. 2002). It follows that the snow sampled every 5 km was not blown from far away, but fell close to the sample site.

Along the transect between DdU and DC, the SA vs. firn temperature plot shows a very scattered distribution of data points, but with a clear trend in maximum values (Fig. 2c). On the basis of observations along the TNB–DC transect, we suggest that, also for DdU–DC transect, the maximum accumulation value represents a value close to snow precipitation, which is in turn very well correlated ($R^2 = 0.95$, $n = 14$) with firn temperature at each site, according to the following function:

$$SA = 15.17(Tc) + 850 \tag{2}$$

this equation is very similar to the one calculated along the TNB–DC transect (Eq. 1).

Using the maximum data from both transects (TNB–DdU–DC), the accumulation is highly correlated ($R^2 = 0.94$, significance level $> 99\%$, $n = 22$) with firn temperature at each site, according to the following function (Fig. 2d):

$$SA = 15.91(Tc) + 858 \tag{3}$$

3.3 Snow ablation and sublimation vs. surface mass balance measurements

Snow redistribution is a local process and has a strong impact on the variability of accumulation at the annual/metre scale (i.e. noise in ice cores). The high variability of SMB is linked to the ablation process. Ablation is determined by the surface sublimation process (wind scouring and, subsequently, sublimation of blowing snow "blowing sublimation") on the plateau and by the snow blown into the sea in coastal areas. Based on these observations, we suggest that the difference between the maximum and minimum SMB value at each site represents the ablation value (Table 2, Fig. 3): GPS2, M2 and MdPt ablation exceeds about 80% of maximum SA, whereas D6 ablation represents 58%, 31Dpt, D2 and D4 range between 20 and 33% and DC represents 12% (the lowest value) of maximum SA. A clear decrease is observed from the coast to the plateau (GPS2-DC; Maximum-Minimum absolute value in Table 2).

The sites where wind action has less impact (site with a small SPWD) present a standard deviation of less than 10% (31Dpt, D2, D4 and DC, Table 2, Fig. 4a) and a good correlation ($R^2 = 0.86$, $n = 4$) between ablation (maximum–minimum accumulation value) and firn temperature according to the following linear equation:

$$Ab = 3.0(Tc) + 162 \tag{4}$$

where Ab is snow ablation (kg m^{-2} a^{-1}) and Tc is the firn temperature at 15 m (°C).

Equation 4 could be mainly related with a small amount of blowing sublimation in areas with no

**Table 2** Spatial variability in SMB from snow radar calibrated using accumulation at main core

| Site | Spatial variability of SMB from GPR[a] | | | | | | SMB compilation | |
	Average (kg m^{-2} a^{-1})	Standard deviation (%)	Minimum (kg m^{-2} a^{-1})	Maximum (kg m^{-2} a^{-1})	Maximum-Minimum (kg m^{-2} a^{-1})	Maximum-Minimum (%)	Vaughan et al. (1999) (kg m^{-2} a^{-1})	Giovinetto and Zwally (2000) (kg m^{-2} a^{-1})
GPS2	93	47	45	201	156	78	121	135
31Dpt	108	10	86	128	42	33	131	110
M2	76	34	17	115	98	85	97	64
MdPt	38	37	10	59	48	82	107	45
D2	43	5	37	46	9	20	67	39
D4	26	9	22	30	8	26	47	37
D6	29	24	17	41	24	58	41	34
Dome C[a]	26	3	25	28	3	12	49	38

[a]Spatial variability has been evaluated along 15 km in each site along the traverse and along 100 km at DC

Fig. 3 Maximum and average SA value, ablation value and map compilation value along the TNB–DC transect
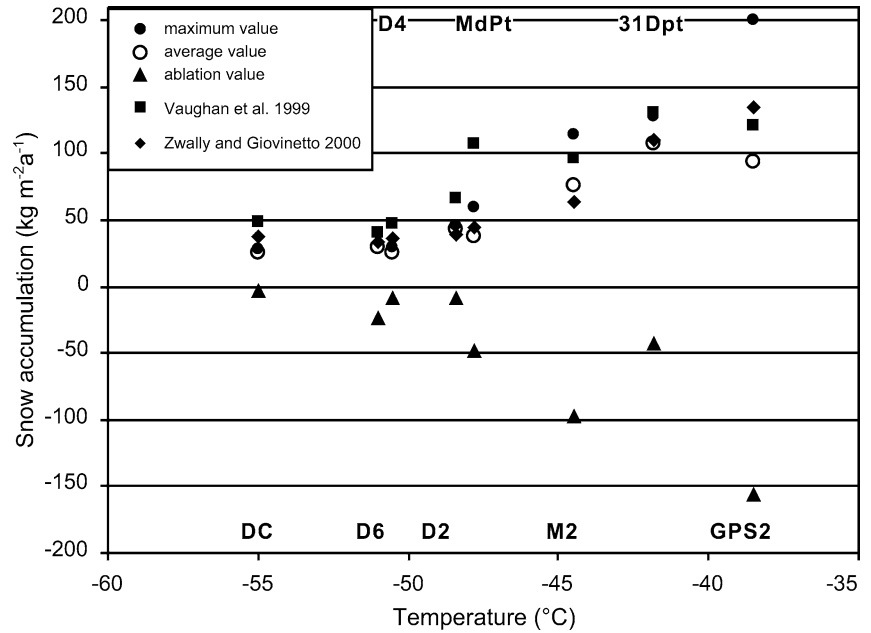
Fig. 4 Ablation vs. temperature along TNB–DC (**a**) and DdU–DC (**b**) traverses. Values along TNB–DC were calculated from maximum and minimum snow radar values; values along DdU–DC represent the difference between firn core maximum and minimum values, and the result of AWS calculations by Bintanja (1998). Confidence level is >95% for correlations
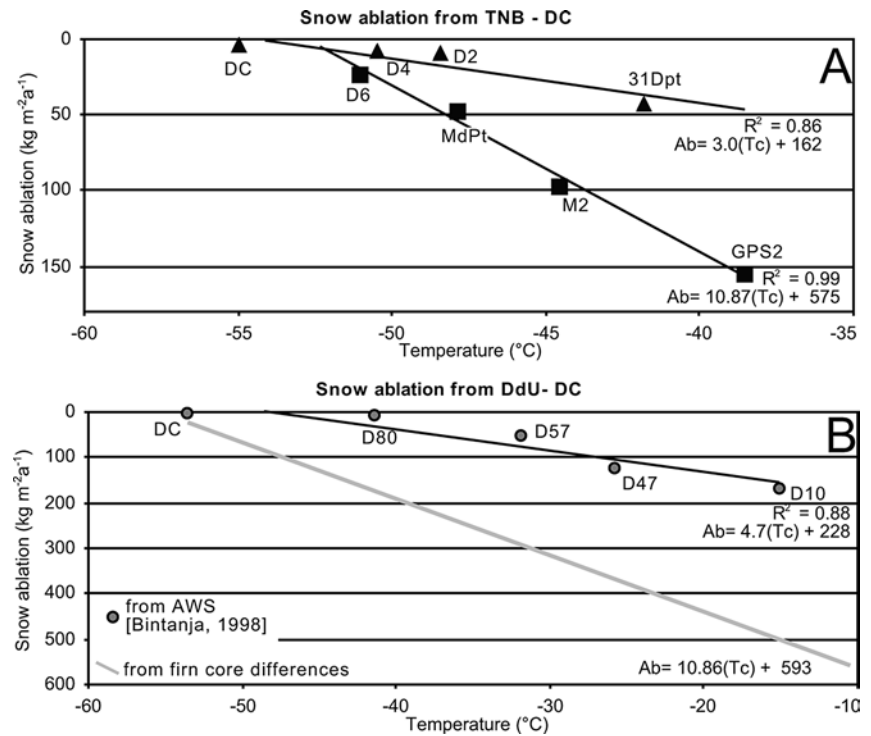
significant change in SPWD (DC, D4, D2 and 31Dpt). Blowing sublimation is negligible at DC, where the mean wind speed is 2.8 m s^{-1} and maximum wind speed is 17 m s^{-1}, but it is still an important factor in SMB at D4 (26%). From the values obtained by Bintanja (1998) along the transect DdU–DC using a model to calculate snow sublimation rates from year-round automatic weather stations data, a linear function is found which is very similar to Eq. 4:

$$Ab = 4.7(Tc) + 228 \tag{5}$$

The sites where the wind action has a greater impact (GPS2, M2, MdPt, D6) show ablation values (from 48 to 156 kg m^{-2} a^{-1}) ranging from 58 to 85% of SA (maximum SA value; Table 2). The standard deviation in SMB at these sites is higher than 10% (Table 2); they have a high variability of SPWD (GPS2, M2, MdPt and D6) and show a very good correlation ($R^2 = 0.99$, $n = 4$)

between ablation (maximum–minimum accumulation value) and firn temperature, according to the following function (Fig. 4a):

$$Ab = 10.87(Tc) + 575 \qquad (6)$$

Using the maximum and minimum values for SA along the DdU–DC traverse (Fig. 4b), it is possible to calculate the analogous ablation function (Eq. 6) along this transect:

$$Ab = 10.86(Tc) + 593 \qquad (7)$$

The ablation functions (Eqs. 6 and 7) along the two transects are very similar. For functions in Eqs. 4 and 5 the constant parameters range from 30 to 45% of functions in Eqs. 6 and 7 constant values, with higher values inland (reduction in wind speed and turbulence), making both surface and blowing sublimation negligible at DC; nevertheless, blowing sublimation is still an important factor in SMB at D6 (Figs. 3 and 4a).

Along the TNB–DC transect, the sites with high standard deviation and high ablation values are extensively covered by permanent wind crust with cracks in polygonal patterns (GPS2, M2, MdPt and D6). Depth hoar layer under well-developed wind crust with cracks clearly indicates prolonged sublimation due to an interruption in accumulation and therefore long, multi-annual, steep temperature-gradient metamorphism (Gow 1965; Watanabe 1978; Goodwin et al. 1994). Wind crusts and snow surfaces have distinctly different albedos, with wind crusts absorbing more solar energy (Frezzotti et al. 2002b; Cagnati et al. 2003). Solar radiation penetrates below the wind crust and warms the subsurface snow layer; this causes an upward transport of water vapour from the subsurface snow layer, condensation of vapour and the growth of loose depth hoar below the crust (Fujii and Kusunoki 1982).

The roughness parameter of wind crust is smaller than that of snow surfaces with sastrugi, dunes and barchans. The lower roughness promotes wind acceleration and divergence of drifting snow (Takahashi et al. 1988). Measurements at the Charcot camp show that snow transportation by saltation starts at wind speeds less than 5 m s^{-1} (Pettré et al. 1986). King and Turner (1997) used Ball's (1960) formula to calculate wind speed as a function of slope gradient and inter-layer potential temperature difference (20 K); wind speeds increase rapidly from 2 to 5 m s^{-1} for slopes between 1 and 2 m km^{-1}. Changes in slope and roughness in the wind direction increase turbulence, thereby increasing the diffusion coefficient and breaking the stable profile of absolute humidity through air mass mixing (Takahashi et al. 1988), which also promotes sublimation of blowing snow.

Eleven metre per second appears to be the threshold wind speed at which the sublimation of blowing snow starts to contribute substantially to katabatic flows in a feedback mechanism (Kodama et al. 1985; Wendler et al. 1993). An increase in drifting snow leads to an increase in air density due to cooling from drifting snow sublimation and to particle incorporation, thus increasing katabatic flow speeds by another 20–30% (Kodama et al. 1985). The threshold wind velocity which differentiates drifting snow (with snow particles moving at low levels) from blowing snow (with snow particles moving at high levels) is 13–14 m s^{-1}. Wind speeds less than 15 m s^{-1} produce transverse features such as ripples, waves and barchans (depositional features), whereas those greater than 15 m s^{-1} produce longitudinal features such as dunes and sastrugi (Kobayashi and Ishida 1979). The extensive presence of erosional (wind crust) and redistribution (sastrugi) features (up to 90%) along the TNB–DC traverse (Frezzotti et al. 2002b) indicates that the threshold wind velocity of 15 m s^{-1} is commonly exceeded over most of the traverse (Fig. 1b).

Surface and blowing sublimation increase with temperature, and blowing sublimation also increases greatly with wind speed. Additional parameters significantly influencing snowdrift sublimation are: turbulence, humidity and friction velocity (Bintanja 1998; Bintanja and Reijmer 2001; Pomeroy 1989). Moreover, high variability of slopes at local scales could drive higher wind turbulence due to changes in SPWD at the kilometre scale. The blowing snow particles are more continuously ventilated on their entire surface in turbulent conditions, and this leads to steeper temperature and humidity profiles, thereby causing higher wind-driven sublimation.

Observations show that the erosion and snowdrift threshold depends on the properties of the snow pack and that it changes as a function of snow metamorphism: the crystal shape of freshly fallen snow does not allow large grain cohesion in snow pack, and this is reflected in relatively high snow mobility values (Gallée et al. 2001). Satellite analysis at TNB (Zibordi and Frezzotti 1996) and field observations along the traverse reveal that the largest blowing snow and sublimation phenomena occur just after snow precipitation, before sintering begins. Very few strong wind events (wind speed higher than 15 m s^{-1}) can greatly increase sublimation through snowdrift, especially during summer.

The maximum value of SA at one site (15 km survey) is very highly correlated with firn temperature and represents the snow precipitation minus sublimation induced by solar radiation during summer.

3.4 Comparison with existing surface mass balance map compilations

The average values of SMB calculated (along 15 km GPR_SA) on the transects at the eight sites (Table 2) were compared with previous SMB map compilations, revealing a general over-estimation of SMB map compilations (Fig. 3). The difference between our data and previous compilations varies on average between 18% (Giovinetto and Zwally 2000) and 65% (Vaughan et al.

1999). The greatest differences are found at the sites from MdPt to DC with lower accumulation (Table 2). Differences between the two map compilations, with a concentration of positive residuals centred at 78°S–140°E, were already observed (Giovinetto and Zwally 2000; Genthon and Krinner 2001). The difference between the two compilations and data collected between DdU and DC is less marked (data are used in the compilations), with values close to survey maximum values (SA). However, values were over-estimated also in this case, with the greatest differences in the Vaughan et al. (1999) compilation. Magand et al. (in press) pointed out analogous overestimation (with an average of about 33%) for the transects along the Wilkes Land. The explanation for the large difference between our results and those reported in the SMB map compilations could be in relation to the following:

– The only accumulation values available in the TNB–DC area before our study were surveyed by the 1959–1960 US Traverse using snow-pit stratigraphy (Stuart and Heine 1961). Some authors (Koerner 1971; Picciotto et al. 1971) compared the accumulation determined by the variation of β-radioactivity with depth with the accumulation rate evaluated from firn stratigraphy in East Antarctica and concluded that the stratigraphic method was basically unreliable in regions of low accumulation and that it overestimated the snow accumulation rate.
– Snow accumulation is assumed close to snow precipitation in the interior of plateau, but our result shows that the sublimation process is underestimated or not even taken into account (e.g. Giovinetto et al. 1992; Genthon and Krinner 2001).

The SMB from European Centre for Medium-Range Weather Forecasts (ECMWF) re-analysis also shows relatively low accumulation rates (e.g. Cullather et al. 1998; Turner et al. 1999; van Lipzig et al. 2002) for this region.

4 Conclusions

This paper reports on a SMB study along the TNB–DC traverse, one of the most unknown areas of East Antarctica. The results provide a new quantification of the impact of surface and blowing sublimation processes on the SMB, its temporal and spatial variability. New data are compared with those of previous SMB studies. The main findings of the paper may be summarised as follows.

Snow precipitation is homogeneous at a large scale (hundreds of km²), but the wind-driven sublimation phenomena controlled by SPWD have considerable impact on the spatial distribution of snow at short (tens of metre) and medium (kilometre) spatial scales. The maximum value of SA at one site (15 km survey) is very highly correlated with firn temperature and represents the snow precipitation minus ablation not induced by wind. The high variability of SMB is mainly due to ablation processes driven by katabatic winds (wind-driven sublimation); a few strong wind events can greatly decrease the mass through snowdrift sublimation, especially during summer.

Compilation and interpolation maps as done are essential contributions to evaluating the mass balance of the Antarctic ice sheet, however compared to our surface mass balance estimations, previous compilations generally over-estimate SMB, up to 65%. In these previous compilations there are large gaps in the SA observation coverage; this is due to the very sparse distribution and representativeness of in situ SMB measurements (1,860 accumulation data points for about 12 million km², i.e. one data point every 6,500 km² on average), particularly in East Antarctica. We have shown that single cores are not always representative of an area. An over-estimation of only 10% to entire Antarctica in the SMB is equivalent to 0.5 mm of annual sea level variation. Despite the fact that very large regions are without accumulation at times, the snow radar layers are traceable everywhere and could represent the tools to provide the spatial variability of the entire continent.

Wind-driven ablation greatly affects the SMB, and one of the largest areas of uncertainty regarding present and future SMB calculations is the role of wind-driven sublimation. An increase in snow precipitation coupled with an increase in temperature and/or wind could increase the SMB in the inner part of East Antarctica alone, whereas it could determine a decrease in SMB in the windy areas that represent 90% of the Antarctic surface.

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