SPATIAL AND TEMPORAL VARIABILITY OF SNOW ACCUMULATION IN EAST ANTARCTICA FROM TRAVERSE DATA

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ABSTRACT Knowledge of recent snow accumulation rate is a primary necessity in ice core and snow accumulation researches. Different methods were used, compared and integrated (stake farm, core, snow radar, surface morphology, remote sensing) at eight sites along a transect from Terra Nova Bay to Dome C (East Antarctica) to provide information about the spatial and temporal variability of snow accumulation. Thirty-nine cores were dated by identifying Tritium/β marker levels (1965-66) and $nssSO_4^{2-}$ spikes of Tambora volcanic event (1816) in order to provide information on temporal variability. Cores were linked by snow radar and GPS surveys to provide detailed information on spatial variability in snow accumulation. Stake farm and ice core accumulation rates are observed to differ significantly, and isochrones (snow radar) correlate well with ice core derived accumulation. The accumulation/ablation pattern using stake farm suggests that the annual local noise (meter scale) in snow accumulation could reach a value of 2 years' ablation and more than 4 times the average annual accumulation, with no accumulation or ablation for a 5 year period in up to 40% of cases. The spatial variability of snow accumulation at the km scale is one order of magnitude higher than temporal variability at the multi-decadal/secular scale. Stake measurements and firn

cores at Dome C confirm an approximate 30% increase in accumulation over the last two centuries, and also in respect to the average over the last 5000 years.

INTRODUCTION

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 and others, 2002a). It has long been known that slope and curvature can play an important role in the snow accumulation; for example, concave depressions accumulate snow at the expense of convex rises (Black and Budd, 1964; Whillans, 1975; Pettré and others, 1986; van den Broeke and others, 1999; Liston and others, 2000; Frezzotti and others, 2002a). A large area of the plateau, where the slope along wind direction is higher than 4 m km^{-1}, has a nil or slightly negative snow accumulation (Frezzotti and others, 2002b). The spatial scale of representativeness for a single ice core record is a critical point in stratigraphic interpretation itself. Micro-relief roughness introduces a high-frequency, quasi-stochastic variability into ice core records of annual layer thickness (Fisher and others, 1985; van der Veen and Bolzan, 1999).

The stratigraphic record is affected by the flow of ice, so the material at depth is slowly moved away from the original deposition site. Richardson and Holmlund (1999) demonstrate the importance of determining the spatial representativeness of cores and making radar surveys prior to drilling. Snow radar is the most useful tool for detecting spatial snow accumulation variability, and firn core time series have the best temporal resolution. Accumulation variability along the radar profile can be surveyed by integrating these two tools.

Ablation processes of snow on short and long spatial scales have a significant impact on post-depositional losses of chemical species by re-emission (Waddington and Cunningham, 1996; Wagnon and others, 1999) and on the interpretation of ice core palaeoclimatic series (Fisher and others, 1985). Interpretation and dating of palaeoenvironmental records extracted from Antarctic ice cores depends on the knowledge of past accumulation changes (e.g. Bromwich and Weaver, 1983; Jouzel and others, 1983).

As part of the ITASE (International TransAntarctic Scientific Expedition) project (Mayewski and Goodwin, 1999) and in the framework of the Franco-Italian Concordia Station collaboration, a traverse between Terra Nova Bay (TNB) and Dome C (DC) and research at Dome C was undertaken (between 1998 and 2000). The study aimed to better understand latitudinal and longitudinal environmental gradients, while documenting climatic, atmospheric and surface conditions over the last 200-1000 years in the eastern and north-eastern portions of the DC drainage area and in northern Victoria Land. The traverse (Frezzotti and Flora, 2002) started from GPS1 (160°48.03'E 74°50.04'S) on November 19, 1998, reaching the Concordia Station at DC (123°23'E 75°06'S, 3232 m) on January 5, 1999; a distance of 1300 km was covered (Fig. 1). The party performed several tasks (drilling, glaciological and geophysical exploration, etc.) during the traverse (Frezzotti and Flora, 2002; Proposito and others, 2002; Gay and others, 2002; Becagli and others, 2003; Traversi and others, in press). GPR (Ground Penetrating Radar or snow radar), GPS (Global Position System) and snow morphology surveys were regularly carried out along the traverse (Frezzotti and others, 2002a; 2002b). In addition, eleven shallow ice cores were drilled within a 25 km radius of the Concordia Station between December 15 and 25, 1999 (Vincent and Pourchet, 2000). In 2000, GPR and GPS surveys, covering a total of about 500 km, were used to link all core sites to DC in order to provide detailed information on the spatial variability of snow accumulation.

This paper combines geophysical surveys (GPR and GPS), field and remote sensing surface observations and firn core analyses to describe the snow accumulation variability along the traverse and at DC. It also provides new information on the snow accumulation process and shows implications on palaeoclimatic series from ice cores.

METHODOLOGY

Twenty-three shallow snow-firn cores, up to 53 m deep, were drilled during the traverse at 8 sites between TNB and DC (at intervals of 90-150 km); five more were drilled at DC using an electromechanical drilling system. In seven areas (Table 1) a 43-53 m deep 'main core' was drilled at the site camp and two 7-15 m deep 'secondary cores' were drilled 5-7 km faraway. The location of the secondary sites was identified in the field after a detailed GPS-GPR profile along 15 km triangular shape (Fig. 1). Snow-radar processing in the field was used to identify variability in the internal layering of the snow pack. Snow temperature profiles, down to the depth of 30 m, were measured at main core sites after a 15-24 hour stabilisation period (Frezzotti and Flora, 2002). Snow/firn density was determined immediately after retrieval by measuring and weighing core sections. Snow was poorly sintered in the uppermost meters; density was thus measured in a pit where stratigraphic studies and snow sampling were also performed (Gay and others, 2002).

Two reference Tritium peak levels resulting from 1965 and 1966 thermonuclear atmospheric bomb tests (Picciotto and others, 1971; Jouzel and others, 1979) and the volcanic sulphate signals of Tambora eruption (1815 AD) and of an Unknown volcanic event (1809 AD) were used to determine the mean accumulation rate at the 'main' cores along the traverse and at DC (Table 1). Tritium depth profiles were transformed into time series by comparing them with the Tritium content of precipitation at the Kaitoke (New Zealand) International Atomic Energy Agency (IAEA) station. In Antarctic ice cores the peaks in SO_4^{2-} signatures of Tambora and Unknown are dated 1816 and 1810, and represent the most reliable volcanic markers for dating the last two centuries (e.g. Legrand and Delmas, 1987; Dai and others, 1991; Cole-Dai and others, 1997; Udisti and others, 2000; Stenni and others, 2002). The volcanic signal was obtained from the sulphate profile. No correction for sea-salt sulphate was made because its contribution is always lower than 15% (Becagli and others, 2003). Analytical procedures for chemical and Tritium measurements are described elsewhere (Udisti and others, 1994; Gragnani and others, 1998; Stenni and others, 2002).

The β–radioactive references of January 1955 and January 1965 were used to determine the mean accumulation rate from 'secondary' cores along the traverse and from 6 m deep cores at DC (Table 1), using methods described in Pourchet and others (1983).

The experimental error ($\pm \sigma_e$) in the calculated snow accumulation rates for the different periods are estimated to be in the order of less than 10% for β–radioactive (1955/65-1998), about 10% for the Unknown-Tambora period (1810-1816) and less than 5% for Tritium (1966-1998) and Tambora-present (1816-1998) periods. These values take into account the different sources of error linked to density determination and the sampling resolution (20-40 cm for β, 3-5 cm for Tritium, 2.5-4 cm for SO_4^{2-}).

At seven sites, 40 stake farms, centred on main core sites, were geometrically positioned about 100 m apart in a cross shape over an area of 4 km^2 (Table 1). At DC, 37 stakes were geometrically positioned at a radius of 3, 6, 12 and 25 km from the culmination of DC. The height of the stakes was measured at Middle Point (MdPt) in 1998, 1999, 2000, 2001 and 2002, at DC in 1996, 1998 and 2000, at others sites (31Dpt, D2, D4, D6) in 1998, 2000 and 2002, at M2 in 1998 and 2002, and at GPS2 in 1993, 1996, 1998 and 2000 (Table 2). The snow accumulation at the stakes was multiplied by the snow density measured in a pit up to 2.5 m deep to obtain water equivalents (we). Snow accumulation variability at stake farms is investigated to determine how representative the results interpreted from a single core might be (Palais and others, 1982).

The integration of GPS and GPR data yields the ellipsoidal height of both the topographic surface and firn stratigraphy (Table 1; Fig. 2). GPS and GPR surveys and analyses are described elsewhere (Frezzotti and others, 2002a; Urbini and others, 2001). For electromagnetic wave speed calculations, the depth-density relation for the snow pack was established using the density profile of three firn cores and one pit in each triangular area along the traverse, and using 17 firn cores and two pits at DC. Density data for each site were input into second order polynomial functions, yielding a determination coefficient of (R^2) >0.9. To facilitate comparison between firn cores and GPR, the eight calculated polynomial functions were used to convert firn depths in the seven triangular areas and at DC into water-equivalent depths. The depth of the snow radar layer was then converted into snow accumulation (GPR_SA) using the depth/age ratio from the 'main' core Tritium snow accumulation. Snow accumulation data derived from core records are in good agreement with data derived from GPR_SA (Table 1). In line with other authors (e.g. Richardson and others, 1997; Vaughan and others, 1999), we assume that the layers producing strong radar reflection are isochronous.

Differences in snow accumulation between main cores and secondary cores (Figs. 2 and 3) are equally reflected in Tritium/β marker and snow radar data (Table 1). The two methods yield different results because they sample different areas: the core diameter is 10 cm, whilst snow radar works at the meter scale. Major differences between the results of the two methods (core and snow radar) are found where the spatial snow accumulation presents the highest variability (GPS2, M2 and MdPt sites). The maximum difference between core and snow radar (20%), and the highest snow radar relative standard deviation (47%) are found at site GPS2. GPS2A and GPS2C cores were drilled a few tens of meters apart and show a 13% difference for the Tritium/β marker (Figs. 1, 2 and 3; Table 1). In contrast, the lowest standard deviation was detected at DC (3%).

Topographic description of the site (elevation, slope, etc.) was performed using the ERS Radar Altimeter Digital Terrain Model, with a pixel size of 1 km, provided by Rémy and others (1999).

RESULTS AND DISCUSSION

Core sites morphological and climatological characteristics

Meteorological data and continental-scale simulation of the wind field surface (Parish and Bromwich, 1991) show that the area between TNB and DC is characterised by a katabatic wind flow, with speeds ranging from 6 to more than 18 m s^{-1} (Table 1).

The topographic profile of the traverse indicates three sectors (Fig. 4): an area extending about 250 km from GPS1 (slope area), characterised by a large slope of up to 25 m km^{-1}, a second area (plateau area) about up to 900 km, with a slope of up to 4.5 m km^{-1}, and the dome area in the last 250 km, with a slope of less than 2 m km^{-1}. The slope profile shows very high variability in the first and second areas, and a homogenous slope in the third area (Frezzotti and Flora, 2002). The surface morphology in the slope and plateau areas is very irregular and can be mainly linked to undulations in the bedrock morphology and secondarily by environmental conditions (wind and accumulation), whereas in the dome area it is derived mainly from environmental conditions and secondarily from the bedrock morphology. Only the megadune in the dome area, close to D6 (Fig. 1), show regular pattern with regular oscillation of surface (Fahnestock and others, 2000). Megadune formation could be explained by the cyclic variation of the redistribution and erosion process of snow along a slope by an atmospheric wave, the atmospheric wavelength must be the same order of magnitude as the megadune (Frezzotti and others, 2002a).

Analysis of the morphological conditions of the area shows that (Figs. 1 and 2, Table 1):

- GPS2, M2 and MdPt are characterised by relatively complex morphologies with 'abrupt' changes in slope and wind direction;
- 31Dpt area presents relatively "steep" slopes (> 4 m km^{-1}), but wind direction is close to the direction of contour lines, and the slope in the wind direction is therefore very low;
- D2 and D4 present low slopes, with a 50-65° angle between the wind direction and the direction of the general surface slope; the D6 core site is located just a few km leeward of the megadune field (Frezzotti and others, 2002a);
- DC has the lowest slope, and the major axis of the dome is aligned in the direction of the prevailing wind.

The morphological characteristics of core sites, where topography is relatively complex (GPS2, M2 and MdPt), are very different in core sites 5 km apart (Figs. 1 and 2, Table 1): microrelief along the TNB-DC traverse consists (fig. 4) of 31% erosional features (wind crust), 59% redistribution features (sastrugi) and only 10% depositional features (Frezzotti and others, 2002b). The D2 and 31Dpt areas are characterised by the alternation of sastrugi fields with sporadic longitudinal dunes (10-20 m long, a few meters wide and up to one meter high), and by seasonal wind crust with sporadic sastrugi. Depositional microrelief occurs extensively only in the D4 area, a point close to the David Glacier ice divide and near DC (Frezzotti and others, 2002b). The GPS2A, M2A, and MdPtC core sites are characterised by the extensive presence of wind crust, consisting of a single snow-grain layer cemented by thin (0.1 to 2 mm) films of ice, with cracks (up to 2 cm wide) and polygonal pattern. The wind crust forms on the surface following the kinetic heating of saltant drift snow under constant katabatic wind flow (Goodwin, 1990) and the condensation-sublimation process on both sides of the crust (Fujii and Kusunoki, 1982). Larger grain sizes are found at the GPS2A, M2A and MdPtC sites along the traverse, and even on the surface, and smaller grains are found in sastrugi or depositional areas (31DptA, MdPtA, D2A, D4A, D6A and DC). Sites MdPtC and MdPtA are only 5 km apart (Fig. 1) but show quite different grain size profiles, both in terms of mean values and variability (Gay and others, 2002). The MdPtA site shows sastrugi of up to 20 cm in height, but no permanent wind crust. The large difference in grain size at the two sites is due to different local-scale snow accumulation processes.

Local spatial variability of snow accumulation

The accumulation/ablation pattern resulting from the stake farm measurements shows large standard deviations (Table 2), and largely reflects snow surface roughness (sastrugi) and/or an interruption in accumulation (wind crust) at the sites. This variability or "noise" is important, and limits the degree to which a single annual snow accumulation value may be temporally representative (Fisher and others, 1985). Analysis of the MdPt stake farm surveyed each year from 1998 to 2002 reveals local-scale spatial variability, suggesting that the annual local noise (meter scale) in snow accumulation could be from less than 2 times to more than 4 times the mean snow accumulation, with accumulation ≤0 for 32% of annual scale observations. It appears that stake measurements and surface morphology are strongly related for all sites. The lowest standard deviation values are present when the slope along prevalent wind direction (SPWD) is low (31Dpt, D4 and DC). In contrast, site M2, with a high SPWD variability, shows the lowest accumulation value and the highest standard deviation of 144%, with accumulation ≤0 for a five-year period in 40% of cases. We generally obtained the same mean value for a five-year period but a much lower variability between individual points, thus concurring with the observation at the old Dome C (Palais and others, 1982; Petit and others, 1982). Ekaykin and others, in press) pointed out that at Vostok the period of smoothing necessary to suppress the noise, due to micro-relief, was estimated to be 7 years both for snow accumulation and ∂D time series. This is due to the fact that the mean surface roughness tends to be conserved, implying a preferential deposition of snow at low-level points and erosion of sastrugi (Gow, 1965; Petit and others, 1982; Alley, 1988).

Temporal variability in snow accumulation

Snow accumulation temporal variability and ice core interpretations

The analysis of spatial variability in snow accumulation at different sites along the traverse shows an extremely high variability of snow accumulation also at the local (km) scale, particularly for sites with a standard deviation in snow accumulation greater than 10% of the mean value (GPS2, M2, MdPt, D6). Tables 1 and 3 show that snow accumulation variability could reach an order of magnitude in a single core (e.g. M2A and GPS2A). The very high spatial variability in snow accumulation may influence the interpretation of firn/ice core records. This is clear when looking at the GPS2A and GPS2B data. GPS2A-GPS2B cores are located along the ice flow and the distance between the two cores is about 5 km (Figs. 1 and 2). The snow accumulated at GPS2B reaches GPS2A after about 250 years (ice velocity 19 m a^{-1}; Vittuari and others, in press). Snow radar and core analyses reveal (Figs. 2 and 3, Table 1) that snow accumulation at site GPS2B is 3 times higher (137 kg $m^{-2}a^{-1}$) than that at site GPS2A (54-60 kg $m^{-2}a^{-1}$). A simple 1-D model can be used to calculate past accumulation rates at GPS2A using the snow accumulation rate derived from snow radar, the depth-age function of firn cores and ice velocity. The 1-D model allows evaluation of the submergence velocity (or burial rate) of the surface and 'simulation' of the snow accumulation rate at core GPS2A (Fig. 5). We did not take into account layer thinning due to vertical strain, because the ratio of the investigated layer depth (20 m) to the entire ice thickness (more than 3000 m; Testut and others, 2000) is less than 1% and is therefore negligible. The analysis of snow accumulation (stakes, Tritium, Tambora, Unknown) reveals an increase in accumulation with depth (present-1965: ~55 kg $m^{-2}a^{-1}$; Tambora – Unknown: 161 kg $m^{-2}a^{-1}$); 'simulation' and present snow accumulation values at GPS2B are very similar, showing that the core variability is due to a spatial variability and not a temporal variability.

Core site D6 is downstream of a megadune with a wavelength of about 2-3 km and amplitude of 2 to 4 m (Frezzotti and others, 2002a). Accumulation in this area is about 30 kg $m^{-2}a^{-1}$, the low accumulation value do not allow to identify any seasonal signal in isotope-chemistry stratigraphy of the firn core. A snow radar profile 45 km West of D6 follows the ice flow direction in the megadune area and reveals the presence of buried megadunes. As for GPS2, the 1-D model (snow radar and ice velocity: 1.5 m a^{-1}; Vittuari and others, in press) can be used to estimate the snow accumulation rate in the megadune area (Fig. 6A). The 'simulated' accumulation rate variability show a periodical variation of about 1500 years, with a standard deviation in the snow accumulation rate of about 28%. At Vostok Station, significant oscillations of snow accumulation and snow isotope composition up to 20 years and, possibly, ~10^2 years has been observed. These are interpreted in terms of drift of snow accumulation waves of various scales on the surface of the ice sheet (Ekaykin and others, 2002). The authors have observed similar periodicity in the δD profile of Vostok ice core at different depths. The reconstruction of past climates based on ice core data from areas with high accumulation spatial variability could then be distorted, snow accumulation variations could be the source of misinterpretation due to mass exchange by diffusion (isotope), gaseous re-emission (NO_3^-, $nssCl^-$ and MSA), densification, and metamorphic processes (highly recrystallised firn, permeability, grain size, snow porosity, close off, total air content, etc.), that occur mainly in porous snow (depth hoar layer) under wind crust (e.g. Alley, 1988; Wolff, 1996; Mulvaney and others, 1998; Delmotte and others, 1999; Wagnon and others, 1999; Traversi and others, 2000; Proposito and others, 2002). A future intermedia ice core (about 500 m in depth) downward the megadune area could provide information about the periodicity (about 7 oscillations) induced by megadune in deep ice core. The length of periodical variations due to meso relief or/and megadunes is strictly correlated with ice velocity and snow accumulation, and can therefore vary in space and time. Nevertheless the length of these periodical variations should be compatible with the use of ice core accumulation records to study temporal changes in snow accumulation (decade scale). As a consequence, a lack of information on local conditions (snow radar) can lead to the incorrect definition of climatic conditions based on the interpretation of core stratigraphy alone. Based on this observation the sites with high standard deviation (GPS2, M2, MdPt, and D6) are not useful in providing information on temporal variations in snow accumulation, because interpretation is very difficult or impossible when the snow originates from different snow accumulation conditions.

The Dome C area was chosen for the EPICA (European Project for Ice Coring in Antarctica) deep drilling project. Deep drilling recently (January 2003) recovered a 3201 m ice core that provides a climate record extending more than 750 kyr back in time. The drill site is about 1.4 km East from present culmination. Our measurements show that the snow accumulation variability is negligible at DC, and that, over an area of 50 km in diameter, the standard deviation in spatial variability (3%) and the difference between maximum and minimum value (12%) are the smallest. On the base of this observation, if the dome has migrated within 50 km in the past, the distortion in the core due to change in accumulation should be negligible. The Vostok core may contain sections of ice that were once megadunes, and accumulated snow may have also been affected by spatial variability. Based on present ice flow at Vostok ice core (Bell and others, 2002), the slope change from the flat surface (due to the subglacial lake) to the slope coming from Ridge B occurred at the end of the Last Glacial Maximum (about 20,000 years ago). Variations in snow accumulation at that time could then be due to the change in slope and increase in wind-driven sublimation. Siegert (2003) pointed out that upstream of Lake Vostok there is a significant difference between the accumulation rate during the glacial period and the more recent period. However, Udisti and others (in press) used stratigraphic correlation between the EPICA-Dome C and Vostok ice cores to show the relative variations of snow accumulation over the past 45kyr. The correlation shows higher variability between glacial and Holocene conditions at DC than at Vostok. They interpret the difference due to regional changes in atmospheric circulation with either a negative anomaly in DC or a positive accumulation anomaly in Vostok, or a combination of both during glacial climate.

It is important to consider aeolian processes when selecting optimum sites for firn/ice coring, because slope variations of even a few meters per kilometre have a significant impact on winds and the snow accumulation process.

Temporal trends of snow accumulation

The sites with less than 10% standard deviation (31Dpt, D2 and D4) show an increase in accumulation (from 14 to 55%) between Tambora-present (1816-1998) and Tritium/β-present (1966-1998; Table 1). Also DC shows a clear increase in accumulation from the Tambora marker (average 25.3±1 kg $m^{-2}a^{-1}$) to Tritium/β (average 28.3±2.4 kg $m^{-2}a^{-1}$) and the stake farm (average 39 kg $m^{-2}a^{-1}$ with a standard deviation of 12 kg $m^{-2}a^{-1}$). With respect to the Tambora marker, stake farm values show a 30% increase in accumulation during the 1996-1999 period. The Tritium/β marker also records about a 10% increase. Based on numerical data for age scale from the EPICA DC1 core, the snow accumulation average over the last 5000 years is 26.6 kg $m^{-2}a^{-1}$, with a standard deviation of 1.0 kg $m^{-2}a^{-1}$ (Schwander and others, 2001). An average value of snow accumulation of 34 cm, with a standard deviation of 12 cm, was measured during the 1996-1999 period using stake farm. The absolute elevations of the stakes (measured by GPS) change homogeneously, with an average value of 0.9 cm and standard deviation of 0.9 cm during the 1996-1999 period (Vittuari and others, in press). The homogeneous decrease in elevation (0.9±0.9 cm) with respect to the snow accumulation value (34 cm) indicates that the stakes are anchored to the bottom. Although snow compaction was not taken into account when calculating snow accumulation, it has often been found to be negligible to accumulation estimation, for low accumulation sites (Lorius, 1983).

Recent increases in accumulation have been reported at other Antarctic locations. Pourchet and others (1983) have observed a general 30% increase in accumulation at 14 Antarctic sites (including Vostok, old Dome C, the South Pole and Ross Ice Shelf) in the period 1960-1975. This general trend was also observed some years later in the South Pole area, with a 32% increase between 1960 and 1990 (Mosley-Thompson and others, 1995; 1999). At Vostok station a stack of 8 sites (6 cores and 3 deep pits) for Tambora marker and 9 sites for β marker suggests slight increase of snow accumulation during the last two centuries (Ekaykin and others, in press). Staked records of δD and snow accumulation for the period 1774-1999 show a 50 year cycle, these variations seem to be in phase with Pacific Deacadal Oscillation index, which suggests a teleconnections between central Antarctica and tropical Pacific (Ekaykin and others, in press). An 11%

increase in accumulation was observed at Talos Dome during the 20th century, while from the Tritium marker to the present there has been a 7% increase with respect to the 800 year average (Stenni and others, 2002). In Wilkes Land, Morgan and others (1991) found a decrease in the accumulation rate from 1955 to 1960, with an increase during the subsequent period. Stenni and others (1999) reported no significant accumulation change from a 200 year ice core record in northern Victoria Land. In Dronning Maud Land, Isaksson and others (1996) have observed a decrease in accumulation from 1932 to 1991. Oerter and others (1999) have produced a composite record of accumulation rates in Dronning Maud Land for the last 200 years by stacking 12 annually-resolved records. The authors have reported a decrease during the 19th century followed by an increase during the 20th century, and these trends were linked to temperature variations derived from stable isotope records. Mosley-Thompson and others (1999) suggest that South Pole recent accumulation increase may be characteristic of the high East Antarctic Plateau.

Domes are the preferred sites for studying the temporal variability of climate using firn/ice cores, because interpretation is easier when all the snow originates from the same point on the surface. Stenni and others (2002) found a good level of concordance between Talos Dome and DC isotopic profiles, which were attributed to similarly distributed storm tracks and sources of moisture. The clear increase in accumulation at DC and Talos Dome, as indicated by stakes and Tritium/β, is consistent with that observed at other sites in East Antarctica, suggesting a regional scale phenomenon. We observed that the increase in accumulation is mainly evident in the inner part of the plateau or at dome sites, where snow accumulation, spatial variability and the ablation process have less impact. Mosley-Thompson and others (1995) have pointed out that the observed increase is not well understood, but that there are several dominant processes that affect snow precipitation, e.g. variability in the sea-ice extent, changes in moisture source regions, frequency, duration and seasonality of cyclonic storms. Bromwich and others (2004), using an atmospheric model (ECT/ERA-15, NCEP2 and DRM) to calculate precipitation over Antarctica from 1979 to 1999, found a significant upward trend of +1.3 to 1.7 kg m^{-2}a^{-1}, which is significant at 95% confidence level for all three datasets. They pointed out that the precipitation increase is consistent with the positive trend in sea surface temperatures observed in midlatitudes of Southern Ocean over the later half of 20th century (Casey and Cornillon, 2001) that encompass the source region of Antarctic precipitation (Delaygue and others, 1999). In addition to these effects, wind-driven sublimation processes must be considered as factors affecting the snow accumulation. The analysis of a 20 year (1980-2000) surface temperature record shows a general cooling of the Antarctic continent, warming of the sea ice zone, and moderate changes over the ocean (Kwok and Comiso, 2002; Doran and others, 2002; Torinesi and others 2003). Wind speeds over sloped terrain increase with decreasing temperature; cold temperatures are associated with strong inversions and hence strong gravitational flows (Wendler and others, 1993). Cooler temperatures over East Antarctica and warming in sea ice areas increase the temperature gradient, and the persistence of katabatic winds and associated wind-driven sublimation. An increase in snow precipitation coupled with an increase in temperature and/or wind could increase the surface mass balance in the inner part of East Antarctica alone, whereas it could determine a decrease in surface mass balance in the windy areas that represent 90% of the Antarctic surface.

CONCLUSIONS

The accumulation/ablation pattern resulting from the stake farm measurements, wind crust and sastrugi heights, suggests that the annual local noise (meter scale) in snow accumulation could be from less than 2 times to more than 4 times the average annual accumulation, with no accumulation or ablation for a 5 year period in up to 40% of cases. The lowest standard deviation value is present where the snow accumulation is the highest and/or where the slope along the prevalent wind direction is low. If we compare the stake farms measurements to the series derived from stratigraphic core analysis, we observe that the number of gaps of one or more years increase with spatial variability and wind scouring.

At many sites stake farms and ice cores accumulation rates are observed to differ significantly. The spatial variability of snow accumulation at the km scale is one order of magnitude higher than temporal variability (20-30%) at the multi-decade/secular scale. The reconstruction of past climates based on firn/ice core drilled in areas with high snow accumulation spatial variability (>10%) is distorted. In megadune areas the distortion of recordings is characterised by a snow accumulation periodicity of about 1500 years. Ice/firn cores require surveys (snow-radar, GPS etc) to characterise the site and its geographical/environmental representativeness. Our results show that snow precipitation is well correlated with firn temperature, but that wind-driven sublimation has a strong impact on snow accumulation.

Stake measurements and firn cores at Dome C confirm an approximate 30% increase in accumulation over the last two centuries, and also with respect to the average over the last 5000 years. An increase in accumulation was also observed at other sites where the slope along the prevalent wind direction is small (31Dpt, D2 and D4). Data from relative "high" slope (>2 m km $^{-1}$) areas should be avoided because wind processes very likely affect it; short-term changes in snow precipitation should thus be studied only in dome areas or at sites undisturbed by winds. Wind-driven ablation greatly affects the snow accumulation, and one of the largest area of uncertainty regarding present and future surface mass balance 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 surface mass balance in the inner part of East Antarctica alone, whereas it could determine a decrease in surface mass balance in the windy areas that represent 90% of the Antarctic surface.

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References

Alley, R.B. 1988. Concerning the deposition and diagenesis of strata in polar firn. *J. Glaciol.*, **34 (118),** 283-290.

Becagli, S., *and 6 others*. 2003. Variability of snow depositions along the 1998/99 ITASE traverse. *Terra Antartica Report*, **8**, 43-48.

Bell, R.E., M. Studinger, A.A. Tikku, G.K.C. Clarke, M.M. Gutner, and C. Meertens. 2002. Origin and fate of Lake Vostok water frozen to the base of the East Antarctic ice sheet. *Nature*, 416, 307-310.

Black, H.P., W. Budd. 1964. Accumulation in the region of Wilkes, Wilkes Land, Antarctica. *J. Glaciol.,* **5 (37),** 3-15.

Bromwich, D.H., and C.J. Weaver 1983. Latitudinal displacement from main moisture source controls $\delta^{18}O$ of snow in coastal Antarctica. *Nature*, **30,** 145-147.

Bromwich, D.H., Z. Guo, L. Bai and Q. Chen. 2004. Modelled Antarctic Precipitation. Part I: Spatial and Temporal Variability. *J. Climate,* **17 (3)**, 427–447.

Casey, K.S. and P. Cornillon. 2001. Global and regional sea surface temperature trends. *J. Climate,* **14**, 3801-3818.

Cole-Dai,, J., E. Mosley-Thompson, and L.G. Thompson. 1997. Annually resolved southern hemisphere volcanic history from two Antarctic ice cores. *J. Geophys. Res.,* **102(D14),** 16761-16771.

Dai, J., E. Mosley-Thompson, and L.G. Thompson. 1991. Ice core evidence for an explosive tropical volcanic eruption 6 years preceding Tambora. *J. Geophys. Res.,* **96(D9)**, 17361-17366.

Delaygue, G., V. Masson, and J. Jouzel. 1999. Climatic stability of the geographic origin of Antarctic precipitation simulated by an atmospheric general circulation model. *Ann. Glaciol.,* **29,** 45-48.

Delmotte, M., D. Raynaud, V. Morgan, and J. Jouzel. 1999. Climatic and glaciological information inferred from air-content measurements of a Law Dome (East Antarctica) ice core. *J. Glaciol.,* **45(150),** 255-263.

Doran, P.T. *and 7 others*. 2002. Antarctic climate cooling and terrestrial ecosystem response. *Nature***, 415**, 517-519.

Ekaykin, A.A., V. Ya. Lipenkov, N.I. Barkov, J.R. Petit, and V. Masson-Delmotte. 2002. Spatial and temporal variability in isotope composition of recent snow in the vicinity of Vostok Station: Implications for ice-core interpretation. *Ann. Glaciol.,* **35,** 181–186.

Ekaykin, A.A., V. Ya. Lipenkov, I.N. Kuzmina, J.R. Petit, V. Masson-Delmotte, and S.J. Johnsen. In press. The changes in isotope composition and accumulation of snow at Vostok Station over the past 200 years. *Ann. Glaciol.,* **39.** In Press.

Fahnestock M.A., T.A. Scambos, C. Shuman, R.J. Arthern, D.P. Winebrenner, and R. Kwok. 2000. Snow megadune fields on the East Antarctic Plateau: extreme atmosphere-ice interaction, *Geoph. Res. Lett.* **27 (20),** 3719-3722.

Fisher, D.A., N. Reeh, and H.B. Clausen. 1985. Stratigraphic noise in time series derived from ice cores. *Ann. Glaciol.,* **7**, 76-83.

Frezzotti, M., S. Gandolfi, and S. Urbini. 2002a. Snow megadune in Antarctica: sedimentary structure and genesis. *J. Geophys. Res.,* **107 (D18)**, 4344, 10.1029/2001JD000673, 1-12.

Frezzotti, M., S. Gandolfi, F. La Marca, and S. Urbini. 2002b. Snow dune and glazed surface in Antarctica: new field and remote sensing data. *Ann. Glaciol*., **34**, 81-88.

Frezzotti, M., and O. Flora. 2002. Ice dynamics and climatic surface parameters in East Antarctica from Terra Nova Bay to Talos Dome and Dome C: ITASE Italian Traverses. *Terra Antartica*, **9(1),** 47-54.

Fujii, Y., and K. Kusunoki. 1982. The role of sublimation and condensation in the formation of ice sheet surface at Mizuho Station, Antarctica. *J. Geophys. Res.,* 4293-4300.

Gay, M., M. Fily, C. Genthon, M. Frezzotti, H. Oerter, and J.G. Winther. 2002. Snow grain size measurements in Antarctica. *J. Glaciol.,* **48(163),** 527-535.

Goodwin, I.D. 1990 Snow accumulation and surface topography in the katabatic zone of Eastern Wilkes Land, Antarctica. *Antarctic Science*, 2(3), 235-242.

Gow, A.J. 1965. On the accumulation and seasonal stratification of snow at the South Pole. *J. Glaciol.,* **5**, 467-477.

Gragnani, R., C. Smiraglia , B. Stenni, and S. Torcini. 1998. Chemical and isotopic profiles from snow pits and shallow firn cores on Campbell Glacier, northern Victoria Land, Antarctica. *Ann. Glaciol.,* **27**, 679-684.

Isaksson, E., W. Karlén, N. Gundestrup, P. Mayewski, S. Whitlow, and M. Twickler. 1996. A century of accumulation and temperature changes in Dronning Maud Land, Antarctica. *J. Geophys. Res.,* **101(D3)**, 7085-7094.

Jouzel, J., L. Merlivat, M. Pourchet, and C. Lorius. 1979. A continuous record of artificial tritium fallout at the South Pole (1954-1978). *Earth Planet. Sci. Lett.,* **45(1)**, 188-200.

Jouzel, J., L. Merlivat, J.R. Petit, and C. Lorius. 1983. Climatic information over the last century deduced from a detailed isotopic record in the South Pole snow. *J. Geophys. Res.*, **88(C4)**, 2693-2703.

Kwok, R., and J.C. Comiso. 2002. Spatial patterns of variability in Antarctic surface temperature: Connections to the Southern Hemisphere Annular Mode and the Southern Oscillation. *Geophys. Res. Lett.*, **29(14)**, 10.1029/2002GL015415, 50, 1-4.

Legrand, M., and R.J. Delmas. 1987. A 220-year continuous record of Volcanic H_2SO_4 in the Antarctic ice sheet. *Nature*, **327,** 671-676.

Liston, G.E., J.G. Winther, O. Bruland, H. Elvehoy, K. Sand, and L. Karlof. 2000. Snow and blue-ice distribution patterns on the coastal Antarctic Ice Sheet. *Antarctic Science*, **12(1)**, 69-79.

Lorius, C. 1983. Accumulation rate measurements on cold polar glaciers. In Robin, G. de Q., ed. *The climatic record in polar ice sheets.* Cambridge University Press, Cambridge, 65-70.

Mayewski, P.A., and I.D. Goodwin. 1999. Antarctic's role pursued in global climate change. *Eos, Transactions*, **80**, 398-400.

Morgan, V.I., I.D. Goodwin, D.M. Etheridge, and C.W. Wookey. 1991. Evidence from Antarctic ice cores for recent increase in snow accumulation. *Nature*, **354,** 58-60.

Mosley-Thompson, E., J.F. Paskievitch, A.J. Gow, and L.G. Thompson. 1999. Late 20th Century increase in South Pole snow accumulation. *J. Geophys. Res.,* **104(D4),** 3877-3886.

Mosley-Thompson, E., *and 6 others*. 1995. Recent increase in South Pole snow accumulation. *Ann. Glaciol.*, **21**, 131-138.

Mulvaney, R., D. Wagenbach, and E. Wolff. 1998. Postdepositional change in snowpack nitrate from observation of year-round near-surface snow in coastal Antarctica. *J. Geophys. Res.,* **103 (D9)**, 11,021-11,031.

Oerter, H., W. Graf, F. Wilhelms, A. Minikin, and H. Miller. 1999. Accumulation studies on Amundsenisen, Dronning Maud Land, Antarctica, by means of tritium, dielectric profiling and stable-isotope measurements: first results from the 1995-96 and 1996-97 field seasons. *Ann. Glaciol.,* **29**, 1-9.

Palais, J.M., I.M. Whillans, and C. Bull. 1982. Snow stratigraphic studies at Dome C, East Antarctica: an investigation of depositional and diagenetic processes. *Ann. Glaciol.,* **3,** 239-242.

Parish, T.R., and D.H. Bromwich. 1991. Continental scale of the Antarctic katabatic wind regime. *J. Climate*, **4(2)**, 135-146.

Petit, J.R., J. Jouzel, M. Pourchet, and L. Merlivant. 1982. A detailed study of snow accumulation and stable isotope content in Dome C (Antarctica). *J. Geophys. Res.,* **87(C6),** 4301-4308.

Pettré, P., J.F. Pinglot, M. Pourchet, and L. Reynaud. 1986. Accumulation in Terre Adélie, Antarctica: effect of meteorological parameters. *J. Glaciol.,* **32,** 486-500.

Picciotto, E., G. Crozaz, and W. De Breuk. 1971. Accumulation on the south Pole-Queen Maud Land Traverse, 1964-1968. In Crary A.P. ed., *Antarctic snow and ice studies II.* Washington, DC, American Geophysical Union, 257-291, (Antarctic Research Series **16**).

Pourchet, M., J.F. Pinglot, and C. Lorius. 1983. Some meteorological applications of radioactive fallout measurements in Antarctic snows. *J. Geophys. Res.*, **88(C10)**, 6013-6020.

Proposito, M., *and 9 others*. 2002. Chemical and isotopic snow variability along the 1998 ITASE traverse from Terra Nova Bay to Dome C (East-Antarctica). *Ann. Glaciol.,* **35,** 187-194.

Rémy, F., P. Shaeffer, and B. Legresy. 1999. Ice flow processes derived from the ERS-1 high resolution map of the Antarctica and Greenland ice sheets. *Geophys. J. Int.*, **139**, 645-656.

Richardson, C., E. Aarholt, S.E. Hamran, P. Holmlund, and E. Isaksson. 1997. Spatial snow distribution mapped by radar. *J. Geophys. Res.,* **102 (B9)**, 20,343-20,353.

Richardson, C., and P. Holmlund. 1999. Regional and local variability in shallow snow-layer depth from a 500 km continuous radar traverse on the polar plateau, central Dronning Maud Land, East Antarctica. *Ann. Glaciol.*, **29**, 10-16.

Schwander, J., J. Jouzel, C.U. Hammer, J.R. Petit, R. Udisti, and E. Wolff. 2001. A tentative chronology for the EPICA Dome Concordia ice core. *Geophysic. Res. Lett.*, **28(22)**, 4243-4246.

Siegert M.J. 2003. Glacial-interglacial variations in central East Antarctica ice accumulation rates. *Quater. Res. Rev*., 22, 741-750.

Stenni, B., *and 8 others*. 1999. 200 years of isotope and chemical records in a firn core from Hercules Névé, northern Victoria Land, Antarctica. *Ann. Glaciol.,* **29**, 106-112.

Stenni, B., M. *and 6 others*. 2002. Eight centuries of volcanic signal and climate change at Talos Dome (East Antarctica). *J. Geophys. Res.,* **D9(107)**, 10.1029/2000JD000317, 1-14.

Testut, L., I.E. Tabacco, C. Bianchi, and F. Rémy. 2000. Influence of geometrical boundary conditions on the estimation of rheological parameters. *Ann. Glaciol.*, **20**, 102-106.

Torinesi, O., M. Fily, and C. Genthon. 2003. Variability and trends of summer melt period of Antarctic ice margins since 1980 from microwave sensors. *J. Climate,* **16(7)**, 1047-1060.

Traversi, R., S. Becagli, E. Castellano, O. Largiuni, and R. Udisti. 2000. Stability of chemical species in firn layers from Antarctica. In: *Conference Proceedings Vol. 69 'Italian Research on Antarctic Atmosphere'.* Colacino M. and Giovannelli G. eds. SIF Bologna, 2000. 421-443.

Traversi, R., *and 7 others.* In press. Spatial and temporal distribution of environmental markers from coastal to plateau areas in Antarctica by firn core chemical analysis. *Int. J. Env. Analit. Chem.in press.*

Udisti, R., S. Bellandi and G. Piccardi. 1994. Analysis of snow from Antarctica: a critical approach to ion-chromatographic methods. *Fresenius J. Anal. Chem*. **349**, 289-294.

Udisti, R., *and 6 others*. 2000. Holocene electrical and chemical measurements from the EPICA-Dome C ice core, *Ann. Glaciol*., **30,** 20-26.

Udisti R., *and 8 others*. In press. Statigraphic correlation between the EPICA-Dome C and Vostok ice cores showing the relative variations of snow accumulation over the past 45 kyr. *J. Geophys. Res. In press*

Urbini, S., S. Gandolfi, and L. Vittuari. 2001. GPR and GPS data integration: examples of application in Antarctica. *Annali di Geofisica,* **44 (4),** 687-702.

van den Broeke, M.R., *and 6 others*. 1999. Climate variables along a traverse line in Dronning Maud Land, East Antarctica. *J. Glaciol.* **45(150),** 295-302.

Van der Veen C.J.,and J.F. Bolzan. 1999. Interannual variability in the net accumulation on the Greenland Ice Sheet: observations and implications for the mass-balance measurements. *J. Geophys. Res*.**,** **104**, 2009-2014.

Vaughan, D.G., H.J.F. Corr, C.S.M. Doake, and E.D. Waddington 1999. Distortion of isochronous layers in ice revealed by ground-penetrating radar. *Nature,* **398(6725),** 323–326.

Vincent, C., and M. Pourchet. 2000. Geodetic measurements and accumulation rate at Dome Concordia, December 1999 and January 2000. *Report for Institut Francais pour la Recherche et la Technologie Polaire and for Ente Per La Nuove Tecnologie l'Energia e l'Ambiente,* pp 16.

Vittuari, L., *and 6 others*. In press. Space geodesy as a tool for measuring ice surface velocity in the Dome C region and along ITASE traverse (East Antarctica). *Ann. Glaciol*., **39,** In press.

Waddington, E.D., and J. Cunningham. 1996. The effects of snow ventilation on chemical concentrations. In Wolff E.W. and Bales R.C., eds. *Chemical exchange between the atmosphere and polar snow.* Berlin, etc., Springer-Verlag, 403-451. (NATO ASI Series I: Global Environmental Change, 43).

Wagnon, P., R.J., Delmas, and M. Legrand. 1999. Loss of volatile acid species from upper firn layers at Vostok, Antarctica. *J. Geophys. Res.,* **104(D3)**, 3423-3431.

Wendler, G., J.C. André, P. Pettré, J. Gosink, and T. Parish. 1993. Katabatic winds in Adélie Coast. In Bromwich, D.H. and C.R. Stearns (eds.), *Antarctic meteorology and climatology: studies based on automatic weather stations*. Washington, DC, American Geophysical Union, 23-46, (Antarctic Research Series 61).

Whillans, I.M. 1975. Effect of inversion winds on topographic detail and mass balance on inland ice sheets. *J. Glaciol.,* **14(70)**, 85-90.

Wolff, E.W. 1996. Location, movement and reactions of impurities in solid ice. In Wolff E.W. and Bales R.C. eds. *Chemical exchange between the atmosphere and polar snow.* Berlin, etc., Springer-Verlag, 541-560. (NATO ASI Series I: Global Environmental Change, 43).

Table 1. Location, morphological and climatological characteristic of drill sites; snow accumulation from stake farms, core analyses, and comparison with snow radar data (GPR_SA). GPR_SA value are reported as snow accumulation using the depth/age ratio from the ‘main’ core Tritium snow accumulation (“Tambora for D4) and depth in % respect to the maximum depth of layering used for the statistical analysis. Katabatic wind speed from Parish and Bromwich (1991).

Site	Long E	Lat S	Elevation (m WGS84)	Depth (m)	T –15 m (°C)	Slope direction (°)	Wind direction (°)	Katabatic wind speed (m s^{-1})	Stakes 1998-2002 (kg m^{-2} a^{-1})	β 1966-98 β 1955-98^{+} Tritium 1966-98* (kg $m^{-2}a^{-1}$)	$nssSO_4^{2-}$ Tambora 1816-1998 (kg $m^{-2}a^{-1}$)	$nssSO_4^{2-}$ Tambora Unknown 1810-1816 (kg $m^{-2}a^{-1}$)	GPR_SA (kg $m^{-2}a^{-1}$) (%)
GPS1								>18					
Core A	160°39.60’	74°48.94’	1192	12.0	-32.7				-----	58±2.9*	-----	-----	-----
Core B	160°39.60’	74°48.94’	1192	8.0	-----				-----	47±4.7	-----	-----	-----
GPS2								>18					
Core A	157°30.13’	74°38.69’	1804	43.6	-38.5	132°	80°		55 (1993-2000)	54±2.7*	60±3.0	161±16.1	54 (27%)
Core B	157°22.80’	74°36.81’	1804	11.4	-----	132°	80°		-----	137±13.7	-----	-----	200 (100%)
Core C	157°30.13’	74°38.69’	1810	8.0	-----	132°	80°		-----	62±6.2	-----	-----	54 (27%)
31Dpt								12-18					
Core A	155°57.6’	74°01.52’	2069	48.0	-41.8	158°	72°		98	112±5.6*	98±4.9	86±8.6	112 (77%)
Core B	156° 0.64’	74°03.51'	2040	16.2	-----	158°	72°		-----	137±13.7*	-----	-----	145 (100%)
Core C	155°55.99'	74°03.80'	2041	7.4	-----	158°	72°		-----	-----	-----	-----	97 (67%)
M2								12-18					
Core A	151°16.17’	74°48.27’	2278	49.5	-44.5	162°	62		8.5	15±7.5*	17±0.8	19±1.9	15 (14%)
Core D	151°09.13’	74°49.86’	2265	12.0	-----	111°	62°		-----	82±8.2	-----	-----	107 (100%)
Core C	151°05.91’	74°48.01’	2272	8.0	-----	247°	62°		-----	44±4.4	-----	-----	61 (57%)
MdPt								12-18					
Core A	145°51.43’	75°32.16’	2454	44.5	-47.8	78°	32°		47	45±2.7*	36±1.8	42±4.2	45 (60%)
Core B	145°47.31’	75°33.03’	2460	12.5		78°	32°		-----	11±1.1	-----	-----	13 (18%)
Core C	145°55.35'	75°31.74’	2452	7.0		78°	32°		-----	60±6.0	-----	-----	75 (100%)
D2								6-12					
Core A	140°37.84’	75°37.33’	2479	49.0	-48.4	89°	37°		30	31±1.6*	20±1.0	23±2.3	31 (94%)
Core B	140°28.61’	75°38.76’	2482	12.5		89°	37°		-----	40±4.0^{+}	24±1.4	-----	33 (100%)
Core C	140°28.54’	75°36.06’	2483	8.5		89°	37°		-----	38±3.8	-----	-----	29 (87%)
D4								6-12					
Core A	135°49.89’	75°35.88’	2793	43.0	-50.5	116°	42°		29	-----	22±1.1	77±7.7	22” (79%)
Core B	135°40.43’	75°37.13’	2795	12.3		116°	42°		-----	29±2.9	-----	-----	28 (100%)
Core C	135°40.74’	75°34.43’	2798	9.0		116°	42°		-----	20±2.0	-----	-----	22 (72%)
D6								6- 12					
Core A	129°48.53’	75°26.85’	3027	52.7	-51.0	97°	46°		39	29±1.4*	36±1.8	37±3.7	29 (85%)
Core B	129°42.41’	75°25.19’	3035	12.3		97°	46°		-----	38±3.8	-----	-----	34 (100%)
Core C	129°38.04’	75°26.42’	3038	9.0		97°	46°		-----	22±2.2	-----	-----	16 (46%)
Dome C							40°	< 6					
EPICA-DC1	123°20.86’	75°06.06’	3233										
FIRETRAC	123°20.86’	75°06.06’	3233		-54.5				-----		24±1.2	32±3.2	
DCN	123°18.72’	75°07.32’	3233	42.3	-55.0				-----	29±1.5*	27±1.4	27±2.7	
DC-1 A17	123°36.26’	75°00.53’	3233	18.0					-----	26±1.3*	25±1.3	30±3.0	
DC-2 E16	123°01.95’	75°02.86’	3230	18.0					-----	27±1.4*	25±1.3	28±2.8	
DC-3 D10	123°11.13’	75°11.68’	3233	20.0					-----	26±1.3*	26±1.3	32±3.2	
DC-4 C12	123°45.51’	75°09.29’	3229	18.0					-----	23±1.2*	25±1.3	25±2.5	
DC-A18	123°49.85’	74°54.55’	3226	6.61					-----	32±3.2	-----	-----	
DC-DORIS	123°05.92’	75°09.16’	3232	6.59					-----	33±3.3	-----	-----	
DC-B11	124°15.97’	75°05.99’	3225	6.40					-----	29±2.9	-----	-----	
DC-F10	123°11.30’	75°00.50’	3230	6.38					-----	29±2.9	-----	-----	
DC-DAU	122°57.04’	75°17.72’	3233	5.85					-----	29±2.9	-----	-----	
DC-E19	122°38.83’	74°59.31’	3223	6.60					-----	30±3.0	-----	-----	
DC-A17	123°36.26’	75°00.53’	3232	6.44					-----	29±2.9	-----	-----	
DC-D11	122°57.45’	75°17.71’	3230	6.56					-----	27±2.7	-----	-----	
DC-A15	122°56.66’	75°17.71’	3233	6.62					-----	30±3.0	-----	-----	
DC-C17	123°23.65’	75°12.52’	3229	6.60					-----	26±2.6	-----	-----	
DC-C18	123°49.83’	75°17.51’	3221	6.59					-----	28±2.8	-----	-----	

Table 2. Stake farm results and surface morphology conditions at main core sites.

Site	*Average snow accumulation 1998-2000 (kg m^{-2} a^{-1})*	*St. dev (%)* 1998-2000	Number of stakes without accumulation or with ablation (%) 1998-2000	*Average snow accumulation 1998-2002 (kg m^{-2} a^{-1})*	*St. dev (%)* 1998-2002	Number of stakes without accumulation or with ablation (%) 1998-2002	*Surface morphology*	*Sastrugi average height (cm)*	*Sastrugi maximum height (cm)*
31Dpt	105.0	22	0	98.4	13	0	sastrugi	5	10
M2	-------	-------	-------	8.5	144	40	Wind crust sastrugi	15	30
MdPt	47.4	65	11	46.7	40	3	sastrugi	30	70
D2	31.8	99	19	30.4	67	14	Wind crust sastrugi	10	70
D4	38.7	48	6	29.3	36	0	deposition form	20	40
D6	38.0	72	9	39.3	78	9	deposition form and sastrugi	30	150
Dome C 1996-2000	-------	-------	-------	39.0	36	0	deposition form	10	30

Table 3: Spatial variability in surface mass balance from snow radar calibrated using accumulation at main core.

	*Spatial variability of surface mass balance from GPR**						
Site	*Average* (kg $m^{-2}a^{-1}$)	*St. dev*	*St. dev* *%*	*Min* (kg $m^{-2}a^{-1}$)	*Max* (kg $m^{-2}a^{-1}$)	*Max-Min* (kg $m^{-2}a^{-1}$)	*Max-Min* %
GPS2	93	44	47	45	201	156	78
31Dpt	108	11	10	86	128	42	33
M2	76	26	34	17	115	98	85
MdPt	38	14	37	10	59	48	82
D2	43	2	5	37	46	9	20
D4	26	2.5	9	22	30	8	26
D6	29	7	24	17	41	24	58
Dome C*	26	0.8	3	25	28	3	12

* Spatial variability has been evaluated along 15 km in each site along the traverse and along 100 km at Dome C

FIGURES

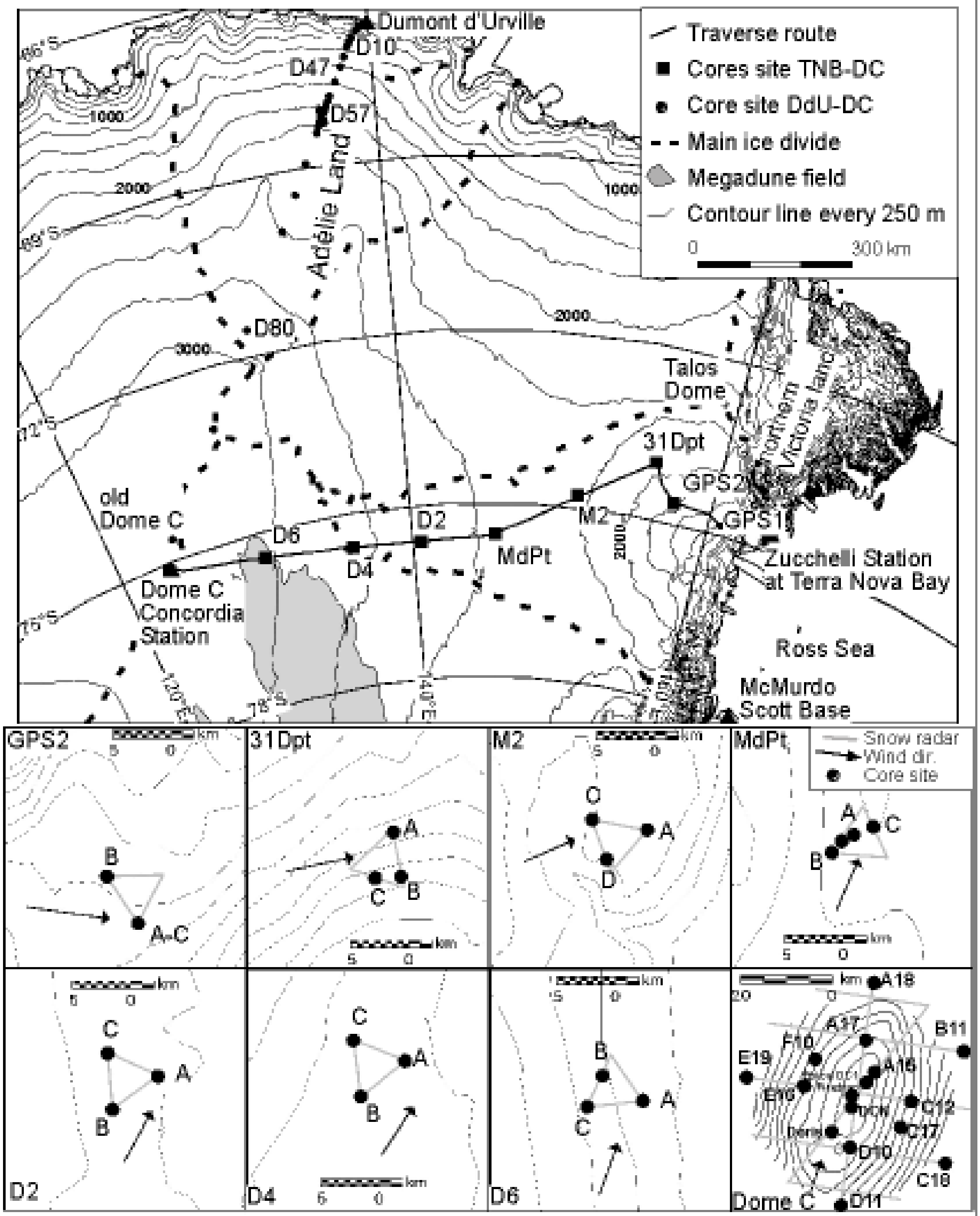

Fig.1 Schematic map of traverse from Terra Nova Bay to Dome C and site location, and detailed map of drill sites (contour every 10 m for GPS2, 31Dpt, M2, MdPt, D2, D4, D6, from Rémy and others, 1999); contour every 1 m for Dome C, from Capra and others (2000); prevalent wind direction from Frezzotti and others (2002b).

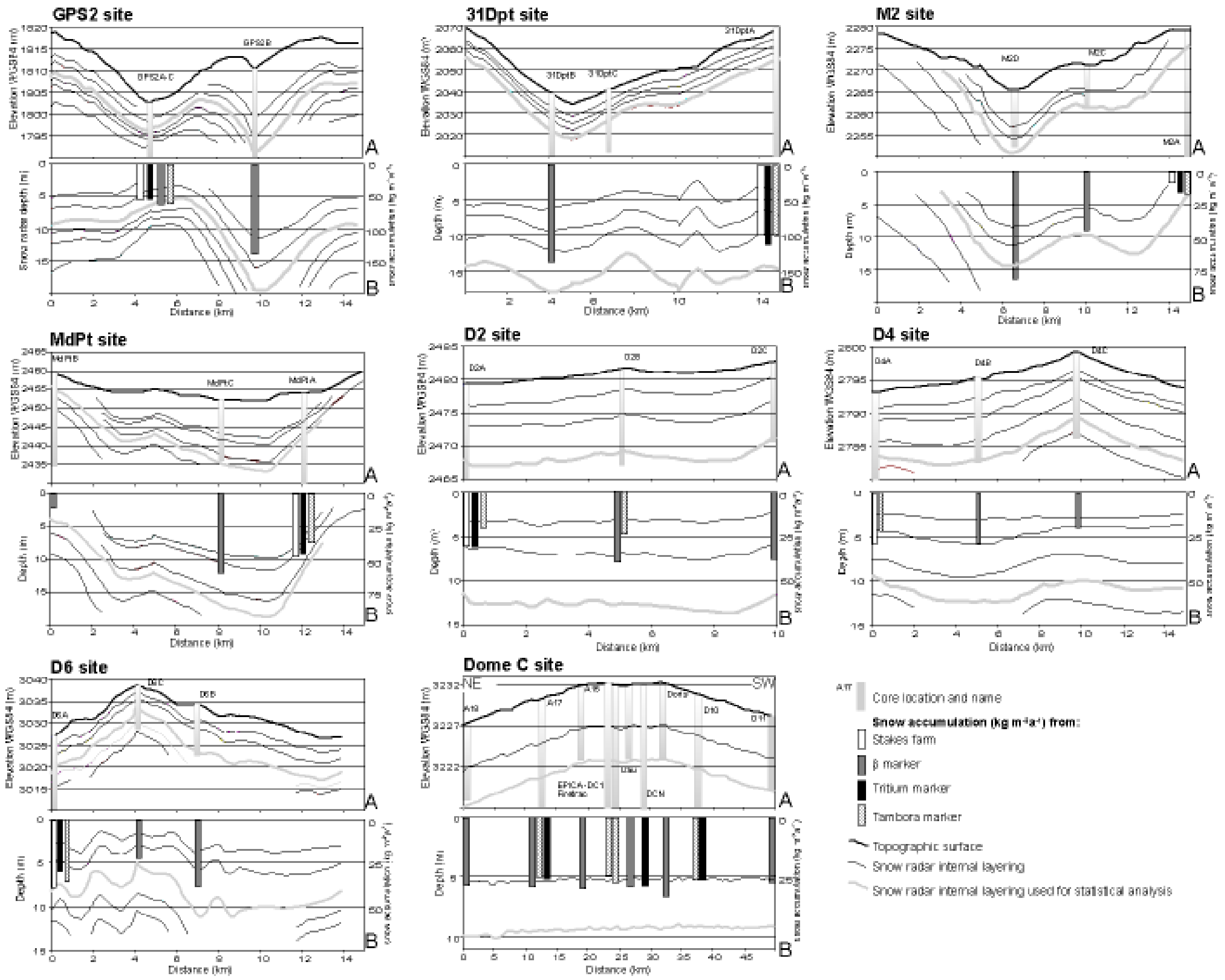

Fig. 2 Core location and elevation of snow radar and surface profiles (A), depth snow radar profile and snow accumulation of firn core using different methods (B).

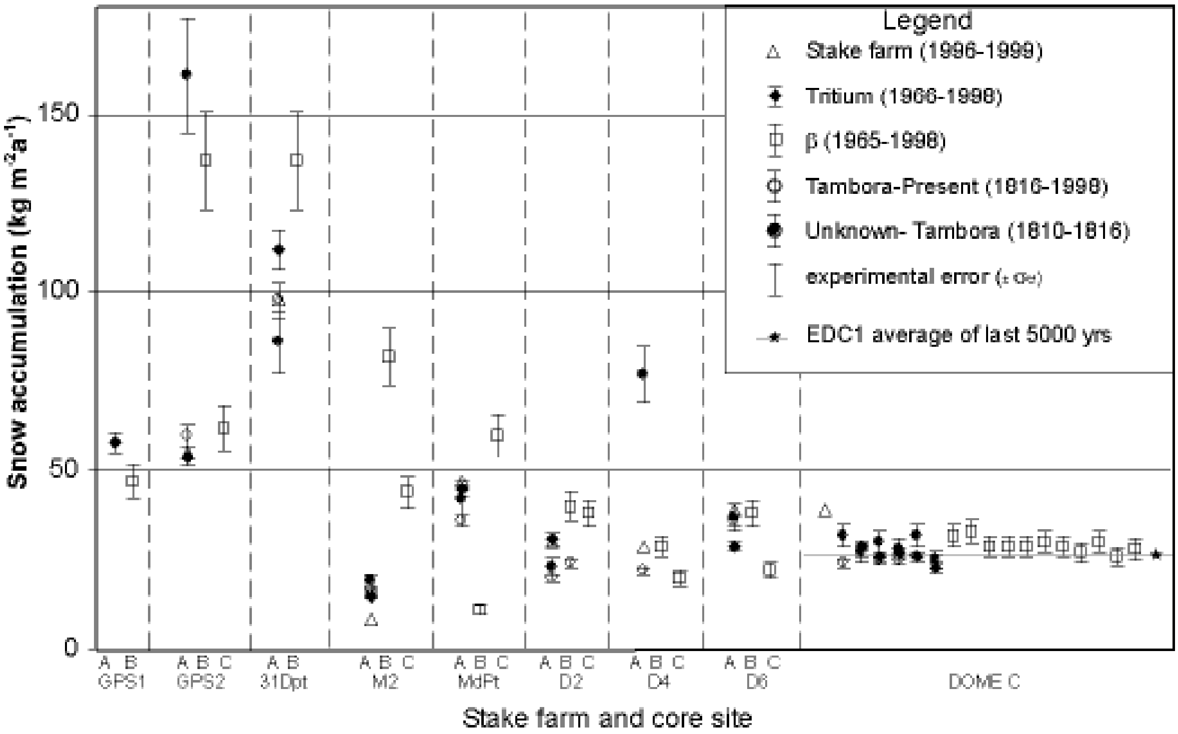

Fig. 3 Snow accumulation rates at core sites using stake farms and different stratigraphic markers.

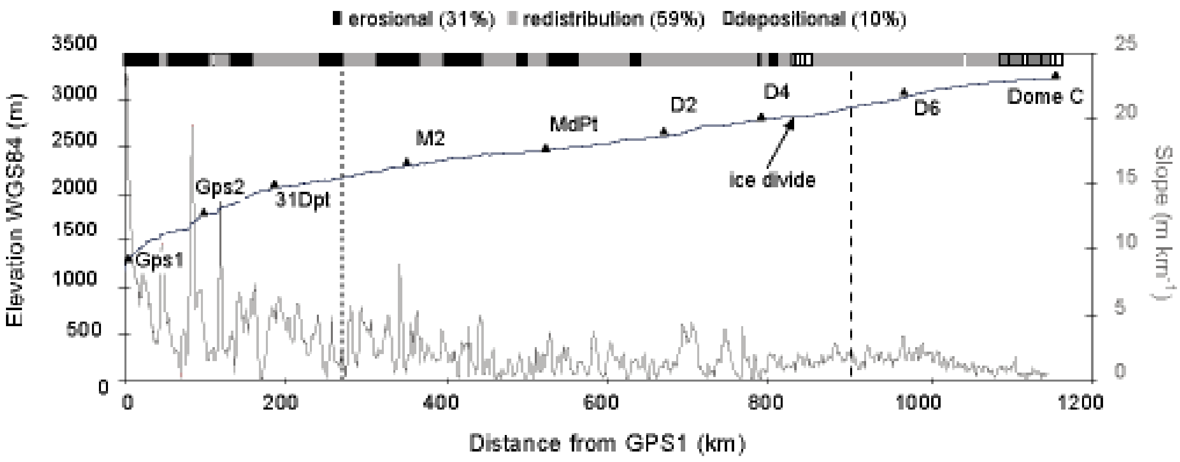

Fig. 4 Surface elevation (solid line), terrain slope (grey line) and micro-relief distribution along Terra Nova Bay – Dome C traverse. Core sites are indicated with triangles.

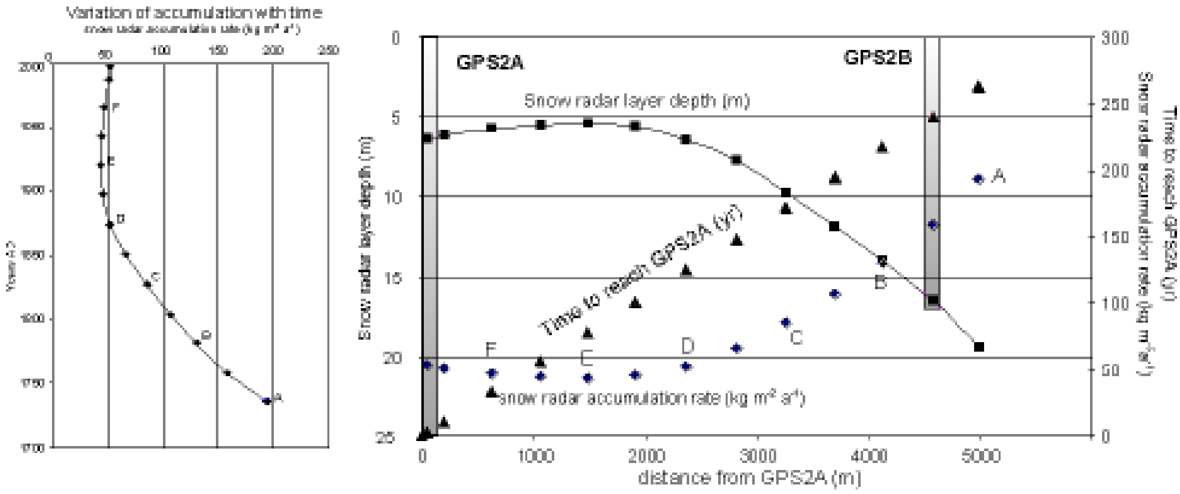

Fig. 5 "Simulation" of snow accumulation rate at GPS2A core using snow-radar, ice velocity and core analyses.

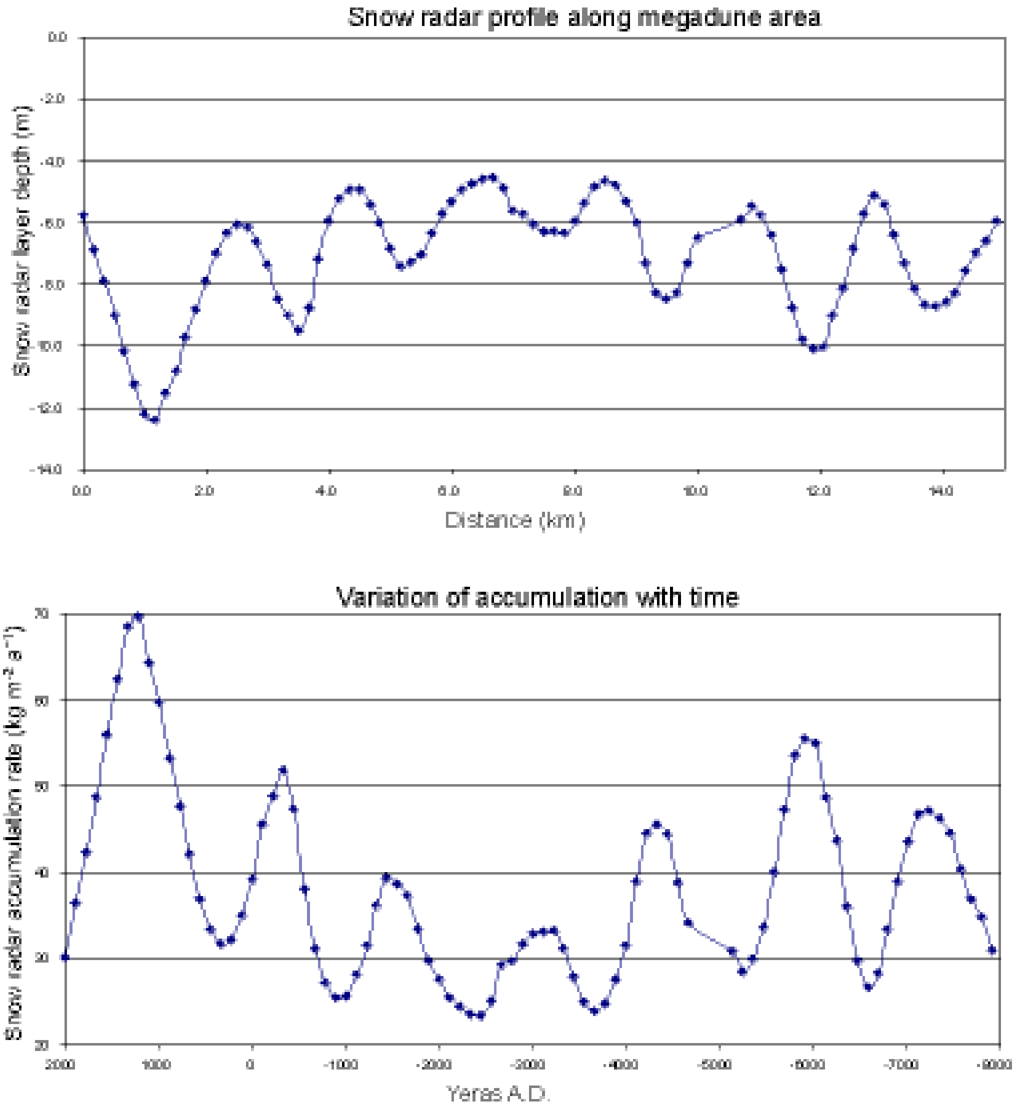

Fig. 6 "Simulation" of snow accumulation rate at D6A core using snow-radar and core analyses.