ACCUMULATION RATE MEASUREMENTS
AT TAYLOR DOME, EAST ANTARCTICA:
TECHNIQUES AND STRATEGIES FOR
MASS BALANCE MEASUREMENTS
IN POLAR ENVIRONMENTS

BY

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ABSTRACT. Accumulation rate measurements on the East Antarctic plateau are challenging due to both spatial and temporal variability. Annual stratigraphy is often not reliably or consistently preserved in the firm, and so accumulation cannot be determined from snow pit stratigraphy. We present a suite of accumulation rate measurements collected over several seasons at Taylor Dome, East Antarctica. We compare net accumulation results from direct burial rate measurements and beta-activity firm cores along a 35 km traverse. The two methods are consistent and show that the net accumulation varies from greater than 10 cm a\(^{-1}\) to about 1 cm a\(^{-1}\) (ice equivalent) southwest to northeast across the dome. We map the depth of shallow radar layering to interpolate and extrapolate these point-location measurements and show that considerable variations occur over kilometer scales resulting from subtle surface topography. We also present accumulation rates estimated from concentration of the cosmogenic isotope \(^{10}\)Be and from activity profiles of \(^{206}\)Pb. Finally, satellite passive microwave data are used to estimate spatially averaged accumulation rates on the regional to continental scale to provide a context for these local observations. We show that robust mass balance measurements in this environment must rely on spatial and/or temporal averaging.

Introduction
The existence of glaciers, ice caps, and ice sheets depends upon a delicate balance of the environmental conditions that govern the processes of accumulation and ablation: any imbalance leads to growth or shrinkage in response to the changing climatic forcing. The study of mass balance of the Earth’s large ice masses is motivated in part for their value as climatic indicators. Yet while mass balance is one of the most fundamental measurements that can be made on an ice body, it can also be one of the most problematic. The differing climates, spatial scales, and response times characteristic of these ice bodies present distinct challenges for measuring the various components of mass balance.

The East Antarctic ice sheet is the world’s largest fresh water reservoir; as a result, it has the greatest long-term potential to affect sea level through changing ice volume. Except in some coastal areas, it is a polar desert. Net annual accumulation rates are typically less than 10 cm a\(^{-1}\) (ice equivalent) (Fortuin and Oerlemans 1990) and the surface is so cold that there is no summer melt season (Zwally and Fiegles 1994). New snow precipitation occurs at any time of the year (Hogan and Gow 1997) and wind plays an important role in affecting net accumulation, both during and after precipitation. Scouring can remobilize a surface layer that represents a large fraction of the annual precipitation. Due to these difficulties, and remoteness, mass balance studies in polar regions are generally reduced to measurement of net accumulation rate and ice thickness change (Hamilton and Whillans 1996).

Net accumulation rates in the interior of East
Antarctica are only sparsely sampled. Remote sensing techniques are desirable due to the physical scale of the problem. With satellite-based remote sensing, accumulation rate measurements are possible over continental-scale areas with high temporal resolution. But these results must be validated with ground observations. In addition, satellite geodesic techniques are approaching the resolution required for continental-scale ice surface elevation change observations (Legrèsy and Rémy 1998; Wingham et al. 1998), yet interpreting these results in terms of ice mass change is complicated by isostatic rebound and transient firn density effects. New total surface mass balance techniques (Hamilton and Whillans 1996) compare vertical velocities measured by Global Positioning Satellites (GPS) to long-term average accumulation rates determined from firn cores. These so-called ‘coffee-can’ experiments promise to give very precise point observations; however, since wind acts to impart spatial variability on the net accumulation rate, it may be difficult to select representative sites at which to measure ‘regional’ mass balance.

Ice core paleoclimate studies offer another motivation to determine modern values and patterns of accumulation rate. The depth gradient of age in an ice core gives the current thickness of layers deposited in the past. After correcting for flow-thinning since the time of deposition, the measured depth–age relation can be converted to an accumulation rate history (Cutler et al. 1995; Cuffey and Clow 1997). Because ice deeper in an ice core probably originated as surface accumulation farther from the drill site, the spatial pattern of net accumulation modulates the layer thickness pattern in a core, and must be accounted for in paleo-accumulation rate studies (Reeh 1990). Furthermore, the concentrations of geochemical species measured in ice cores become more useful as paleoclimate indicators when their fluxes to the ice sheet can also be determined (Alley et al. 1995). This requires knowledge of the flux of H2O to the ice sheet surface, i.e. the net accumulation rate.

Because of their large spatial extent and their long response times to climate change, the ice sheets respond only to spatial and temporal averages of net accumulation rate. This presents a challenge for accumulation rate measurements; accumulation rate data are most useful when they are representative of large spatial areas, long time intervals, or both. Wind redistributes snow in most glacial environments. It is not unusual for the uppermost 1–2 decimeters to be scoured (leaving sastrugi) and redeposited elsewhere as drifts, effectively introducing temporal and spatial noise into the thickness of an annual layer of snow. The depth of this reworked surface zone and the height of the wind-generated surface roughness, i.e. the magnitude of the accumulation rate noise, is relatively insensitive to the spatially and temporally averaged net accumulation rate. Where net annual accumulation is much greater than the roughness height, redistribution is a minor source of noise and a short-term observation (e.g. one year) of snow accumulation at a single point can be representative of a larger area. However, on the Antarctic polar plateau, the net accumulation rates are low and comparable to the roughness height, so a short-term measurement of accumulation at a single point is likely to give a poor estimate of the average accumulation rate. This problem can be expressed through the coefficient of variation $C$ given by the ratio of the standard deviation $s$ to the mean value $\bar{x}$. When the coefficient of variation is significant, i.e. of order unity, single measurements cannot be expected to be representative of the average value and some averaging of multiple measurements is required. This averaging is key to extracting meaningful accumulation rate estimates from low accumulation rate areas such as Taylor Dome.

**Taylor Dome**

We present a suite of net accumulation rate measurements obtained over several seasons as part of an ice core paleoclimate program (Grootes et al. 1994; Waddington et al. 1994) at Taylor Dome, South Victoria Land, Antarctica (Fig. 1). In January, 1994, we obtained a 554 m-long ice core to bedrock. The core contains a climate record which extends beyond the previous interglacial period (to >130 ka). Taylor Dome is an approximately 20 by 80 km feature along a ridge that separates two major East Antarctic outlet glaciers. At an elevation of c. 2375 m, the mean annual surface temperature at the dome crest is −43°C (Waddington and Morse 1994).

Situated at the edge of the polar plateau, Taylor Dome is subject to gravity-driven winds that funnel cold interior air masses through valleys of the Transantarctic Mountains (Fig. 1). These winds transport snow from the southwest and are primarily responsible for sastrugi formation and drift transport. Sometimes they result in net deposition, but more commonly they erode the surface. The primary sources of new snow deposition are syn-
Fig. 1. Map showing Taylor Dome (77°47'S, 158°43'E, 2375 m) in relation to McMurdo Sound and South Victoria Land. Sites WS, DS, and TM are shown within a box indicating the region of Fig. 2. Thick and thin arrows (figure and inset) are modern moisture-bearing storm trajectories (st) and gravity wind (gw) directions. The shaded regions distinguish the ice-free Dry Valleys, the Ross Ice Shelf, and the seasonally open Ross Sea. White indicates glacier ice cover.

optic weather systems that enter the Ross Embayment and penetrate the Transantarctic Mountains south of Taylor Dome (Harris 1992). Net annual accumulation at Taylor Dome reflects a balance between these two dominant circulation modes. As a result, the accumulation is both spatially and temporally variable: areas with positive and negative balance are often found in close proximity, and for some sites, years with positive balance are interspersed with years with negative balance.

Fig. 2. Taylor Dome survey region: locations of strain net marker poles are shown by small crosses, β-activity firm core by solid circles, and snow pits of Fig. 3 by squares. The bamboo pole and radar traverse of Fig. 5 are the thin-solid and thick-dashed lines, respectively. The surface elevation contours (25 m spacing) are from an airborne radar survey with 5-km-spaced grid lines (Morse 1997). The drill site (DS), the Taylor Mouth Site (TM) and a sampling site on the western slope of Taylor Dome (WS) are shown. The numbers (10, ...50) and letters (S, C, N) define major intersections of the survey grid. The shaded boxes show areas a, b and c of Table 2.
Local topographic ‘facets’, defined by the surface slope and aspect with respect to these winds (Fig. 2), experience differing climatic conditions (Waddington and Morse 1994). This is evident in varying surface morphology and firn properties. The surface is often relatively soft, smooth, and uniform on the south side of the dome, suggesting comparatively regular, high accumulation. Firn density, hardness and δ^{18}O profiles (Fig. 3a) in a snow pit southeast of the dome suggest approximately 0.5-m-thick annual layers. In contrast, the surface north of the crest is characterized by large sastrugi, by patchy distributions of hard wind-
Table 1. Taylor Dome geochemical accumulation rate results. The sites, each represented by a sub-table, are spatially organized by their relative locations (Fig. 2). In each, the top line is the site identifier, the second line is accumulation rate in cm a⁻¹ (ice equiv.) from ₁⁰Be-cores, and, where available, the third line is accumulation rate estimated from ₁⁰Be concentration by assuming a dry flux of c. 10² atoms cm⁻² a⁻¹ (Steig 1996). Sites 20S and 10C each report results of two separate cores.

<table>
<thead>
<tr>
<th>site</th>
<th>WS</th>
<th>₁⁰Be</th>
</tr>
</thead>
<tbody>
<tr>
<td>40S</td>
<td>7.24</td>
<td>6.16</td>
</tr>
<tr>
<td>30S</td>
<td>6.58</td>
<td></td>
</tr>
<tr>
<td>20S</td>
<td>7.46</td>
<td>8.27</td>
</tr>
</tbody>
</table>

The rate of accumulation measurements were packed snow, and by localized areas of net ablation. Massive wind slab layers are interspersed with thick depth hoar layers and no clear annual isotopic variation is seen in a snow pit northwest of the dome (Fig. 3b). Because of these variations, we were unable to reliably determine the accumulation rate at Taylor Dome purely by snow pit analyses. We explore several techniques that employ spatial and/or temporal averaging to obtain robust mass balance measurements in low net accumulation rate environments. The large spatial variability at Taylor Dome allows us to investigate the effectiveness of these techniques for net accumulation rates that range from decimeters per year (ice equiv.) to nearly zero, and thus to assess these methods for estimating mass balance over large areas of the polar ice caps.

Direct burial-rate measurements
The benchmark technique for net accumulation rate measurements in polar environments relies on the detection of radioactive fallout from atmospheric nuclear bomb testing. The gross-β activity measurements (Picciotto and Wilgain 1963) give the depth to these horizons, and hence, the average accumulation since deposition. Because it averages accumulation rate over several decades, this method is insensitive to high variance due to interannual variability in deposition. We collected firm cores from 15 locations distributed over the survey area (Fig. 2). Firn samples of 0.5 kg were melted and filtered with cation filters. Gross-β activity of these filters (e.g. Fig. 4a) reveals the fallout maxima which were deposited in Antarctica c. 1955 and 1964 (after accounting for an approximately 1 year interhemispheric transport time). This approach is best suited for accumulation rates near 10 cm a⁻¹; if the accumulation rate is much higher, the horizon may be too deeply buried for ready access, and for much lower rates, the atmospheric test decade may be poorly resolved due to undersampling or postdepositional wind scouring and redistribution of the snow (e.g. the TM site profile of Fig. 4b).

New down-hole sensors can detect bomb layers by gamma-ray counting (Dunphy and Dibb 1994). By providing on-site information about the bomb-layer depths, these techniques make it possible to adjust geochemical sampling strategies in the field. This may be particularly valuable at low accumulation rate sites, by identifying the appropriate depth zones for higher resolution  β-sampling. The down-hole Gamma counting technique requires lengthy integration times at discrete depths. However, the bomb layers can be identified without need to sample and transport firm core or snow samples.

Net accumulation rate results from (Table 1) show higher values to the south of the divide crest.
than to the north, consistent with our expectation from snow pit analyses. In addition, substantial site-to-site variations are seen over distances of only kilometers. Because gross-$\beta$ measurements have temporally averaged out effects of year-to-year drifting and sastrugi, these differences reflect real spatial differences. Apparently the spatial pattern of net accumulation is still undersampled, in spite of the large number of $\beta$-cores analyzed.

Our spatial sampling strategy for the $\beta$-activity accumulation rate measurements emphasized a roughly north–south transect that follows the flow of ice passing through the core site (Fig. 2). We focus on this transect for comparison of results from other accumulation rate measurement techniques (Fig. 5).

Since the $\beta$-core observations undersampled the spatial variability, we employed traditional marker burial rate measurements in an attempt to capture more spatial detail. A network of 250 metal poles, spaced at 600 m to 2.5 km intervals, was emplaced to measure ice motion over successive seasons. As a by-product of this survey, height measurements from these survey poles form a spatially detailed burial rate data set (Fig. 2). Year-to-year, and pole-to-pole, burial rates of these poles show substantial variability. We estimate the precision of each accumulation measurement to be better than 1 cm; although these errors are not large in an absolute sense, the coefficients of variation for the resulting accumulation rate data sets (Table 2) are of order unity. This presumably reflects real spatial

![Image of graph showing accumulation rate measurements](Image)

**Fig. 5.** Comparison of accumulation rate measurements along southwest–northeast line shown in Fig. 2. Gross-$\beta$ (circles) measures net accumulation since initial atmospheric bomb tests in 1955. Crosses show burial rates of undisturbed marker poles spaced every 300 m over the period Jan. 1992 to Jan. 1996. A density of 0.4 g cm$^{-3}$ was used to convert pole depths to ice equivalent. The solid curve is the relative depth to a radar layer normalized to 5.5 cm a$^{-1}$ at site DS. The lower panel shows surface topography along profile (The path of radar profile, shown dashed, diverges from the pole line left of approximately km 10, with the profile following the surface slope).

**Table 2.** Statistical summary of marker pole burial rate and $\beta$-core observations for sites shown in Fig. 2. $\bar{x}$ is average accumulation rate in cm a$^{-1}$ (ice equivalent), $s$ is the standard deviation, $C = s/\bar{x}$ is the coefficient of variation, and $n$ is the number of samples. We have assumed a constant firm density of 0.4 Mg m$^{-3}$ to convert burial depths to ice equivalent. Densities for firm ($\beta$) core samples were calculated from weight and volume measurements.

| method | area a | | | area b | | | area c | |
|--------|--------|--------|--------|--------|--------|--------|--------|
|        | $\bar{x}$ | $s$ | $C$ | $n$ | $\bar{x}$ | $s$ | $C$ | $n$ | $\bar{x}$ | $s$ | $C$ | $n$ |
| poles  |        |     |     |     |        |     |     |     |        |     |     |     |
| 1992–93| 0      | 8.7 | 7.5 | 0.86 | 8      | -1.5 | 5.0 | 0.4 | 1        |     |     |     |
| 1993–94| 0      | 9.3 | 7.6 | 0.82 | 21     | 5.6  | 9.1 | 1.63| 2       |     |     |     |
| 1994–95| 12.2   | 11.4| 0.93| 44    | 6.7    | 4.5  | 0.67| 33   | 1.4 | 5.9 | 4.21| 11 |
| 1995–96| 11.8   | 14.6| 1.24| 32    | 11.7   | 4.4  | 0.38| 38   | 5.6 | 3.5 | 0.63| 2  |
| bamboo | 8.7    | 1.3 | 0.15| 21    | 7.5    | 1.4  | 0.19| 31   | 3.4 | 2.2 | 0.65| 39 |
| $\beta$-core| 7.2 | 0.8 | 0.11| 4     | 5.9    | 2.3  | 0.43| 3    | 3.1 | 2.6 | 0.84| 2  |
variability in wind scouring and drifting, accompa-
nied by low net deposition, and indicates that indi-
vidual measurements are not reliable estimators of
the average accumulation rate. Two-dimensional
polynomial approximations following the method
of Lliboutry (1974) did not convincingly resolve
either spatial or temporal trends in these noisy data.
However, spatially averaging in three bins (Fig. 2)
resolves the north–south trend (Table 2) indicated
by the snow pit and β-core observations.

We also measured the burial rate of bamboo
marker poles spaced at c. 300 m intervals along
several of our primary travel lines to augment the
burial rate measurements from the ice motion
marker survey. After spatial averaging using a five-
point running mean (c. 1.5 km boxcar), measure-
ments spanning the 1990–94 period clearly resolve
the south–north accumulation rate gradient indi-
cated by the β-core analysis (Fig. 5). By using
denser spatial sampling, and by treating spatial av-
ergates of groups of poles as primary data to mini-
mize scouring and drifting effects, we obtained
data sets with lower coefficients of variation (of or-
der 0.2; see ‘bamboo’ in Table 2), reflecting prima-
rily real spatial differences in long-term net accu-
mulation. The accumulation pattern from these ob-
servations reveals spatial structure along the transect that was not resolved by the β-core results
due to their poorer spatial coverage. Evidently the
accumulation pattern at Taylor Dome varies over
distances of kilometers. The bamboo pole burial
rate observations were able to resolve the rich spa-
tial pattern. Still, comparing this pattern with the β-
core results must be done with caution since they
sample very different time intervals.

Radar stratigraphy
Direct mass balance measurements are inherently
laborious and prone to spatial undersampling.
Therefore we explore remote sensing techniques
for spatially continuous observations. Ice-pene-
trating radar surveys of polar ice sheets commonly
reveal spatially coherent internal layering. The
‘layers’ are widely recognized as isochrons since
they likely result from depositional processes,
shallow layers (<500–1000 m) being primarily due
to ice density variations while at greater depths
they are likely due to acidic fallout from volcanic
eruptions and/or changes in impurity concentration
associated with climatic transitions (Hammer
1980; Fujita and Mae 1994). The layer pattern deep
in the ice column is often complicated by flow over
basal topography; but near the surface, the relative
depths and thickness of the layers are controlled
primarily by burial rate. We used a 7 MHz center-
frequency, monopulse radar system developed at
the University of Washington (Weertman 1993) to
detect bed topography and internal layering in the
vicinity of the Taylor Dome ice core (Morse 1997).
A profile of relative ice equivalent depth to the shal-
lowest resolvable layer yields a spatial pattern that
agrees with the burial rate measurements along the
north–south line (Fig. 5). In particular, the radar
data confirm the high amplitude, kilometer-scale
variations indicated by the bamboo pole burial rate
observations. From vertical strain rate measure-
ments near the core site, we estimate that the age of
this layer is nearly 700 years. The radar stratigra-
phy shows that this spatial pattern of accumulation
rate has been present through the Holocene (Morse
1997). These spatial variations are strongly corre-
lated with changes in surface slope (Fig. 5), sug-
uggesting an interplay of prevailing weather direc-
tions and surface topography.

Relative accumulation rate patterns mapped by
radar stratigraphy may be converted to absolute
values if the layers can be dated in at least one lo-
cation, e.g. by an ice core, and when ice flow mod-
els are used to convert the observed layer thick-
nesses to initial layer thicknesses by calculating
the total vertical strain experienced by the ice. After ac-
counting for the density profile and dynamic layer
thinning due to ice flow, the spatial variations of
layer depths and thickness can give continuous
profiles of modern and past accumulation rate pat-
terns (Morse et al. 1998).

Other geochemical techniques
The β-activity method is effective for measuring
net accumulation since the 1955–64 period; how-
ever, other geochemical tracers can be employed to
determine net accumulation over other intervals.
The production rate of the cosmogenic isotope
10Be in the stratosphere is modulated by the solar
and terrestrial magnetic field. Approximately 11-
year production rate variations due to regular solar
magnetic field reversals, associated with the
Schwabe sunspot cycle, are detectable in Antarctic
firn (Lal and Peters 1967; Steig et al. 1996). This
provides an opportunity for decadal-resolution ac-
cumulation rate measurements. The peak-to-peak
spacing of solar-modulated 10Be concentration
(Fig. 6a) in a firn core from site ‘40S’ (Fig. 2)
yielded accumulation rate of 7.2 cm a−1, in agree-

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Fig. 6. The $^{10}$Be profile from site 40S (a) shows modulation due to the Schwabe solar cycle (Steig et al. 1998). The two peak-to-peak intervals (right arrows) and one trough-to-trough interval are each taken to be 11 years apart. $^{210}$Pb activity from near site TM (b) decreases with depth due to radioactive decay with c. 22.7 year half-life. The dashed line is the expected $^{210}$Pb activity if we accept the $\beta$-core result of 2.26 cm a$^{-1}$ for the TM site.

The flux of $^{10}$Be to the snow surface is relatively constant over timescales longer than the Schwabe cycle and shorter than timescales for geomagnetic field variations (Raisbeck and Yiou 1988). Hence the concentration of $^{10}$Be in ice and snow is a measure of the relative fluxes of $^{10}$Be and $H_2O$ to the ice sheet surface (Raisbeck and Yiou 1985; Raisbeck et al. 1987). Estimates of net ice accumulation rate obtained from pit-average $^{10}$Be concentrations (Steig 1996) agree in general with the $\beta$-core results (Table 1). The $^{10}$Be concentration profile from the Taylor Dome ice core (Steig et al. 1999) provides an extended accumulation rate history that suggests relatively constant $H_2O$ accumulation through the Holocene and remarkably low accumulation during the Last Glacial Maximum, in agreement with the accumulation rate history inferred from the ice core timescale (Morse et al. 1998).

In areas where accumulation rates are extremely low, i.e. less than a few centimetres per annum, methods that require the detection of horizons, such as gross-$\beta$, can be imprecise since the characteristic shape of the activity profile may be indistinct, or the critical layers may be missing entirely. We encountered the former situation with the gross-$\beta$ results from the TM site (Fig. 4b). In such areas it is desirable to use another geochemical tracer that does not require very high sampling frequency or continuous layer preservation. Other workers have used the isotope $^{210}$Pb to estimate accumulation rate (Picciotto et al. 1968; Sanak and Lambert 1977), with mixed success compared with other methods. We have explored the utility of $^{210}$Pb to measure accumulation specifically in very low accumulation rate areas, where its special properties convey distinct relative advantages. $^{210}$Pb is the eleventh daughter of the $^{238}$U decay series. $^{238}$U is the most common form of natural uranium (accounting for 99.3% of natural uranium) and is present in soils and minerals. The sixth daughter of the decay series is the water-insoluble, inert gas $^{222}$Rn, some of which enters the atmosphere and is transported from the source-site. The radon decays ($t_{1/2} = 3.8$ days) into the next relatively stable element in the series, $^{210}$Pb. Since $^{210}$Pb is highly water soluble, it is quickly bound to aerosols and flushed from the atmosphere. This process results in a steady flux of $^{210}$Pb from the atmosphere to the snow. The $^{210}$Pb is then buried by subsequent accumulation. Thus profiles of $^{210}$Pb activity with depth can be used to determine the burial rate (Crozet et al. 1964). Since $^{210}$Pb has a half life of 22.7 years, it is an effective tracer to estimate accumulation rates on the scale of 20 to 60 years.

In the austral summer 1997/98, we collected 1–2 kg snow samples from a pit dug near the TM site (Fig. 2). The $^{210}$Pb activity was measured through $\alpha$ detection of the decay of a daughter isotope, $^{210}$Po (the method is described by Robbins 1978).
Preliminary results show that surface activity of $^{210}\text{Pb}$ in this area is comparable to values measured at the South Pole (Sanak and Lambert 1977). The observed decrease of $^{210}\text{Pb}$ activity with depth (Fig. 6b) suggests accumulation rates similar to the $\beta$-core result from same site, although still deeper sampling would be required to determine the accumulation rate with high confidence.

Satellite-based remote sensing

A new satellite-based remote sensing method of accumulation rate determination is currently being developed. In many locations, annual layers of Antarctic snow accumulation are separated by depth hoar which forms at the beginning of each winter. Recent work (West et al. 1996; Arthern et al. 1997) indicates that such density stratification in the upper few meters causes polarization of microwave energy that is thermally emitted from below the surface. A synthesis of ground observations with radiometric observations from the NIMBUS-7 Scanning Multichannel Microwave Radiometer (SMMR, $\lambda = 4.5$ cm) supports a theoretically predicted polarization/accumulation rate relationship (Arthern et al. 1997). A preliminary accumulation rate map, calibrated to Wilkes Land snow crust parameters (Goodwin 1988), agrees quantitatively on large scales with ground-based maps and qualitatively with ground observations in the Lambert drainage. This map indicates greater topographic influence on accumulation rates than do the maps based on ground data. For example, the map shows an asymmetry on the east and west sides of the Lambert basin that agrees quantitatively with recent ground data (Higham and Craven 1997). The results also indicate that the method is sensitive to millimeter-scale snow crust properties (e.g. thickness and spacing) that are not necessarily annual in nature. Thus the radiometric method at present relies on calibration with ground observations of crust parameters, accumulation rates, or both.

The spatial resolution of the SMMR observations is roughly 150 km (interpolated onto a 25 km grid). This is too coarse for direct comparison with ground observations, but the SMMR results do provide a valuable regional context (Fig. 7). The SMMR map shows that the regional accumulation rate spatial gradient in the vicinity of Taylor Dome is in fact east–west, rather than north–south. In fact, the east–west gradient measured by $\beta$-counting at sites WS-DS-TM (Figs 1 and 2, Table 1) has the opposite sense to the regional pattern in Fig. 7. Evidently, local factors that influence the net accumulation rate, notably the interaction of surface topography and the prevailing weather direction, are sufficient to overwhelm the regional pattern in some cases. The accumulation rate inferred from SMMR observations in the vicinity of Taylor Dome is roughly double that of the ground measurements. This may be due to a combination of two effects: accumulation at Taylor Dome may in fact...
be lower than the regional mean that the SMMR observations represent, and second, the snow crust parameterization from the Wilkes Land calibration site may be inappropriate for this region of South Victoria Land.

Summary

The low accumulation rates and active wind-driven surface reworking prevalent over the East Antarctic plateau make measurements of local accumulation rates a challenge. New snow precipitation and reworking operate on different spatial scales. Spatial patterns of precipitation depend on the thermodynamics of synoptic systems transporting moisture, and their interaction with the regional topography. Ice sheet topography on a scale comparable to the thickness of the katabatic flow (i.e., a few hundred meters) can affect wind velocity, and therefore also the post-precipitational redistribution of snow (Whillans 1975; Takahashi et al. 1994). In addition, drifts and sastrugi represent enhanced erosion and deposition at spatial scales of meters. Burial rate measurements can be made with high precision, but these effects impart spatial and temporal variability that can be large compared with the magnitude of the desired mean annual accumulation rate signal. A determination of accumulation rate is useful only if its coefficient of variation is significantly less than unity; averaging is the key to reducing the coefficient of variation. Precise accumulation rate determinations for specific sites can be achieved with temporal averaging, while spatial averaging can return temporally resolved results. The $\beta$-activity method succeeds due to its inherent multidecadal temporal averaging. Radar stratigraphy mapping can probe to even greater depths, and thus longer time intervals. By employing many closely spaced measurements, the bamboo pole survey achieves spatial averaging that allows accumulation rate determinations at temporal resolution of a few years. Similarly, ground-truthed satellite-based remote sensing can provide accumulation rate observations with much lower spatial resolution, but over continental scales. Because of their innate spatial averaging, satellite-based methods should also provide reliable estimates of regional and continental-scale temporal changes as the period of observational record increases.

Both spatially and temporally averaged accumulation rates are important for long-term mass balance analyses, and for climate studies. Short-term (e.g., decadal or shorter) averages can be strongly influenced by El Niño–Southern Oscillation (ENSO) cycles and other variations. Averaging over several decades can minimize this effect. Shorter-term measurements can then give information about current climate oscillations by determining if the observable accumulation rate is greater or less than the long-term average. An accurate multidecadal accumulation rate is also an imperative for longer-term accumulation rate measurements, such as for ice core interpretation. While steadily deposited geochemical indicators (e.g., $^{10}$Be) can give long-term accumulation rate histories, they depend upon knowledge of current $\mathrm{H}_2\mathrm{O}$ accumulation rate to calibrate the longer-term geochemical record by making it possible to directly measure the modern flux of the species at the site. Gross-$\beta$ activity firm cores are the standard tool for decadal mean, point observations in Antarctica for sites with low enough accumulation to be easily hand-augered, but high enough to preserve regular stratigraphy. In many areas of East Antarctica with even lower accumulation rates, $^{210}$Pb and $^{10}$Be techniques may be required for decadal mass balance observations. Radar stratigraphy, mapped from the surface, provides invaluable data on spatial gradients over tens of kilometers that affect ice core records. Ultimately, relating point observations over continental scales must be achieved with airborne mapping of radar stratigraphy or with satellite-based observations.

Finally, our observations confirm that local surface topography on the scale of a kilometer or less can dramatically modify the net accumulation rate; furthermore, the effect depends on orientation of the topography in relation to the wind and accumulation sources. This relationship must be considered around remote-sensing ground control points, and it should be explored and utilized around every ice core site.

Acknowledgements

We are grateful to Pieter Grootes for organizing the Taylor Dome ice core program, and to Minze Stuiver for providing the snow pit $\delta^{18}$O data. We thank the many capable field assistants who made this work not only possible, but enjoyable. We also thank Antarctic Support Associates for logistical assistance. Support for this work was provided by NSF/OPP grants 8915924, 9316162, 9421644, 9526979 and NASA grant NAG5-6819.
ACCUMULATION RATES AT TAYLOR DOME

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Steig, E.J., 1996: Beryllium-10 in the Taylor Dome Ice Core: Ap-


Manuscript received August 1998, revised and accepted January 1999.