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Recent Precipitation Trends over the Polar Ice Sheets

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With 8 Figures

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Summary

Meteorological and glaciological analyses are integrated to examine the precipitation trends during the last three decades over the ice sheets covering Antarctica and Greenland. For Antarctica, the best data source is provided by glaciologically-measured trends of snow accumulation, and for limited sectors of East Antarctica consistency with precipitation amounts calculated from the atmospheric water balance equation is obtained. For Greenland, precipitation rates parameterized from atmospheric analyses yield the only comprehensive depiction. The precipitation rate over Antarctica appears to have increased by about 5% over a time period spanning the accumulation means for the 1955–65 to 1965–75 periods, while over Greenland it has decreased by about 15% since 1963 with a secondary increase over the southern part of the ice sheet starting in 1977. At the end of the 10-year overlapping period, the global sea-level impact of the precipitation changes over Antarctica dominates that for Greenland and yields a net ice-sheet precipitation contribution of roughly $-0.2\,\mathrm{mm\,yr^{-1}}$. These changes are likely due to marked variations in the cyclonic forcing affecting the ice sheets, but are only weakly reflected in the temperature regime, consistent with the episodic nature of cyclonic precipitation.

These conclusions are not founded on high quality data bases. The importance of such changes for understanding global sea-level variations argues for a modest research effort to collect simultaneous meteorological and glaciological observations in order to describe and understand the current precipitation variations over both ice sheets. Some suggestions are offered for steps that could be taken.

1. Introduction

Precipitation changes over the ice sheets of Antarctica and Greenland are important considerations for variations of global sea level. These ice sheets hold large amounts of fresh water extracted from the global ocean, and are in approximate balance between mass input and mass wastage. Their annual precipitation rates, equivalent to $\sim 7.2\,\mathrm{mm\,yr^{-1}}$ of sea level decrease (5.8 for Antarctica, 1.4 for Greenland), are large in relation to the estimated $2\,\mathrm{mm\,yr^{-1}}$ of present global sea level rise (Douglas, 1991). For Antarctica, a precipitation increase results in a net removal of water from the ocean because the response time of the ice sheet is very long compared to the time over which the atmospheric changes take place (Whillans, 1981). For Greenland, the impact of a precipitation increase is less clear because the mass wastage of the ice sheet is equally shared by ice berg calving and by melting (Warrick and Oerlemans, 1990). The response time of the latter approaches that of the atmospheric changes and thus the actual situation in Greenland also depends upon the accompanying summer ablation conditions. It should be emphasized that the actual contribution of each ice sheet to global sea level change also depends on its exact mass balance state, so that, for example, if Antarctica was losing mass to the ocean a precipitation increase could make this net loss decrease or even become a net gain if the precipitation increase was large enough.

Direct precipitation measurements over polar ice sheets are problematic because of the difficulty of discriminating between snow that is actually falling at observation time and that which fell previously but has been picked up from the surface and is being transported by the wind (e.g., Bromwich, 1988). An additional complication in the much less windy high interior of Antarctica is that precipitation amounts are so small that routine gauge measurements are ineffective, with, for example, only a trace being recorded during 92% of the months over a ten-year period at South Pole (Bryazgin, 1986). Even beyond the ice-sheet margins precipitation measurements are questionable because of the profound impact of wind upon gauge collections of falling snow (the predominant precipitation type for both Greenlandic and Antarctic coastal stations). For example, in the Canadian Arctic precipitation measurements collected at weather stations amount to only one-third to three-quarters the amount inferred from accumulation values from adjacent ice fields (Woo et al., 1983). One strategy has been to devise corrections to gauge measurements to allow for the most important effects, such as those due to wind (Groisman et al., 1991). Such corrections are not often applied to precipitation records because of the effort required. Due to such lack of attention the reliability of any precipitation trends obtained from the long but uncorrected precipitation records from the coastal regions of Greenland (from some stations dating back to the late 1800's) should be treated with skepticism.

Another meteorological approach is to use indirect methods to calculate the precipitation rate. One technique is the atmospheric moisture balance where the precipitation is found as the residual from the budgeting of the fluxes of water vapor into and out of an atmospheric volume. This is suitable over seasonal and longer time scales for regions of at least 1 million square kilometers that are monitored by a good synoptic network of radiosonde stations (e.g., Rasmusson, 1977; Bromwich, 1988). Another approach is to sum precipitation amounts calculated from estimates of synoptic-scale vertical motion (Bromwich et al., 1993). These approaches have met with success, and both are considered here.

Another totally different approach to inferring precipitation amounts and their variations over the polar ice sheets is to measure the amount of snow accumulated over one, or more typically several, years. This net build-up of snow at the surface is the end result of almost exclusively solid precipitation (due to falls of snow and/or ice crystals) minus the net runoff of meltwater, the net flux of water vapor to the surface due to frost formation and condensation minus sublimation and evaporation, and the deposition minus the erosion of snow by drifting. These measurements are taken most easily in the dry snow zone where no melting takes place (Paterson, 1981, p. 7) using a variety of methods (Lorius, 1983), such as monitoring the height of the snow surface relative to stakes and searching for known radioactive horizons in the firn. Over the high interior of both ice sheets accumulation equals precipitation with little error. Generally, the relative contributions to accumulation other than precipitation become more important with approach to the coast. Comparatively little effort has been directed toward measuring the year-to-year accumulation changes. Of course, such methods require considerable effort to obtain representative spatial coverage, a limitation that tends not to be present in the indirect meteorological methods described above. On the other hand, the glaciological methods are much better understood than the polar meteorological ones, particularly with respect to the inherent errors. These atmospheric and ground-based methods provide complementary ways to describe and understand precipitation variations over the polar ice sheets.

This paper provides an overview of what is known about "recent" precipitation changes over the ice sheets covering Greenland and Antarctica. Here, recent is defined as extending from the mid 1950's or early 1960's to today. This period has been chosen because at least some complementary meteorological and glaciological data sets cover this time period. Accumulation measurements are much more extensive for Antarctica than Greenland whereas the meteorological observational network is better for the latter area than the former. Discussion is presented separately for Antarctica and Greenland with the former emphasizing the glaciological data and the latter the meteorological observations.

2. Antarctica

2.1 Background

Study of the atmospheric transport of moisture provides the necessary context to understand the precipitation characteristics over the continent as

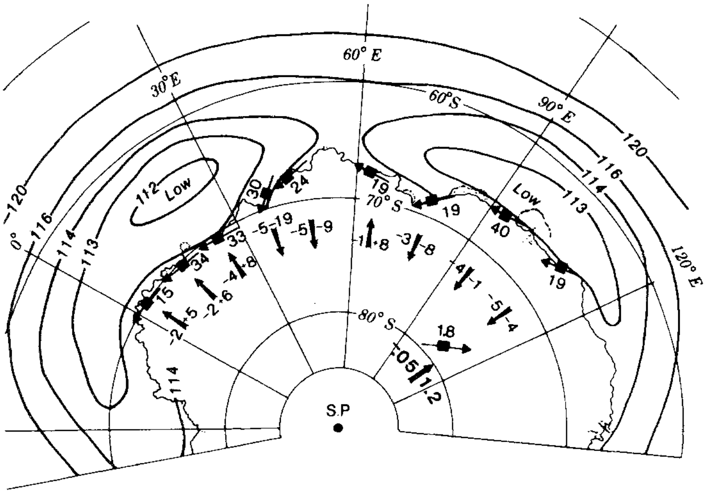

Fig. 1. Annual moisture transport for 1972 (in $kg\,m^{-1}\,s^{-1}$) in relation to the mean atmospheric circulation. The solid squares identify rawinsonde stations, and the attached arrows show the direction of annual transport. The transport magnitude in kilograms per meter per second is entered next to each arrow. The broad solid arrows give the direction of the meridional component of annual transport at the adjacent station; the eddy contribution is listed on the left side and the steady part on the right side of each arrow. The contours are annual geopotential heights in dekameters of the 850-hPa surface from Taljaard et al. (1969). (Figure adapted from Bromwich, 1988.)

well as their variations. However, the sparse rawinsonde data base limits this approach to East Antarctica. Furthermore, the limited attention paid to this topic means that it can not yet be used to describe the trends that do exist. Glaciological observations of accumulation rate variations, although not spatially comprehensive, do describe some large precipitation increases over the last three decades.

Diagnosis of coastal rawinsonde observations for 1972 (Bromwich, 1988) showed that the total poleward flux of water vapor across the periphery of East Antarctica is dominated by the eddy terms arising from the cross correlation between the meridional wind component and the specific humidity; northerly winds tend to transport warm, moist air poleward, and southerly winds tend to carry cold, dry air equatorward. Figure 1, from the same analysis, shows that there is a recurring spatial organization to the meridional eddy moisture fluxes. All these fluxes are poleward, but they increase from west to east in conjunction with the change from net mass outflow to net mass inflow associated with the quasi-stationary vortices in the low-pressure trough surrounding the continent. The longitudinally-averaged coastal dominance of the eddy flux term means that precipitation variations in this area will only be weakly related to the mean temperature conditions, as the most important precipitation events (and thus most of the eddy transports) occur during relatively limited periods.

As shown by Fig. 1, the meridional eddy transport becomes a small part of the total meridional transport in the high interior of East Antarctica at Vostok where the eddy component of the moisture transport is $-0.05\,kg\,m^{-1}\,s^{-1}$ and the steady component is $1.20\,kg\,m^{-1}\,s^{-1}$ (Bromwich, 1979). Over the high plateau the majority of precipitation occurs by the semi-continuous fall of ice crystals from apparently clear skies (Schwerdtfeger, 1984). In this situation it is reasonable to presume that the amount of precipitation is limited by the average amount of water vapor which is concentrated near the top of the surface temperature inversion. Robin (1977) proposed that the precipitation amounts are related to the average temperature at this level and thus implicitly to the saturated specific humidity over ice. In support of this hypothesis he presented a convincing log-linear plot of annual precipitation amounts versus annual free atmospheric temperatures for a range of sites in Antarctica. This spatial relationship reflects the strong topographic forcing affecting both temperature and precipitation, but does not necessarily reflect recent temporal variations at a fixed site, as has frequently been assumed. Furthermore, Vostok is not representative of the western part of the Antarctic plateau where the influence of the steady transport associated with the "circumpolar" vortex is much less. Clear sky precipitation observations by Kuhn (1970) at Plateau Station (79.2° S, 40.5° E), which were summarized by Bromwich (1988), indicate that eddy moisture fluxes strongly modulate the ice crystal production rate. This implies that the temporal relation between precipitation and free atmospheric temperature over the western Antarctic plateau may not be well defined.

2.2 *Glaciological Observations*

Based upon radioactive bomb horizons in polar firn, Pourchet et al. (1983) reported average accumulation rate increases of about 30% for the 1965–75 decade compared with the decade that preceded it. The areas involved covered parts of East and West Antarctica, including the Ross Ice Shelf. This result did not attract much attention. Recently this topic was brought to the forefront by Morgan et al. (1991) who noted that the accumulation rate over Law Dome (near 110° E) and adjacent coastal parts of East Antarctica declined during the 1950's, returned to the previous level by 1960, and then increased by 20% by 1985. The most recent values are also about 20% above the mean for the past 200 years. Based on other evidence including that of Pourchet et al. (1983), Morgan et al. (1991) argued that the precipitation had increased over large parts of Antarctica with an implied impact on global sea level amounting to -1.0 to $-1.2\,\mathrm{mm\,yr^{-1}}$.

In an article on the impact of Antarctica on global sea level, Jacobs (1992) assembled a more extensive survey of observations of "recent" accumulation rate variations over Antarctica than referred to above. The most relevant aspect was that both increases and decreases were found. However, these trends covered a variety of time periods and were presented qualitatively. The original sources quoted by Jacobs have been consulted (see Table 1) and the percentage changes of decadal-average accumulation between 1955–65 and 1965–75 have been calculated. The results are presented in Fig. 2. Where possible the percentage change over the longer period of 1950–60 to 1970–80 was also derived and is entered on the figure in parentheses next to the decadal change. For the shorter time period both increases and decreases of 15% or less are common, but there are four sites where increases of 28% or more are found. The result of spatially integrating these changes with the meteorological analyses described below to get a percentage change for the entire

Table 1. *Accumulation Data Sources Used to Construct the Trend Analysis in Fig. 2, Mostly Following Jacobs (1992)*

Site	Approx. location	Reference	Comments
James Ross Is.	64° S, 58° W	Peel (1992)	5-yr binomially-weighted series
Dolleman Is.	71° S, 61° W	Same as above	Same as above
Gomez	74° S, 71° W	Same as above	Same as above
Siple	76° S, 84° W	Mosley-Thompson (1992)	10-yr. average values
Ice Stream B Catchment Area	83° S, 120° W	Whillans and Bindschadler (1988)	Original data: 1955–65 vs. 1965–85
Ross Ice Shelf	80° S, 180°	Pourchet et al. (1983)	~10 yr. averages based on radioactive bomb horizons
South Pole	90° S	Same as above	Same as above
South Pole	90° S	Jouzel et al. (1983)	Smoothed time series
Vostok	79° S, 107° E	Pourchet et al. (1983)	Same as Ross Ice Shelf
Dome C	75° S, 123° E	Same as above	Same as above
Marginal slopes, East Antarctica	67° S, 139° E	Same as above	Same as above
Site GD 15	69° S, 131° E	Morgan et al. (1991)	Smoothed by gaussian filter with 3-yr bandwidth
Site GD 03	69° S, 116° E	Same as above	Same as above
Law Dome	67° S, 113° E	Same as above	Same as above
Sites W200 and S18	70° S, 45° E	Satow and Watanabe (1990)	Annual values; two-station average used
Neumayer	71° S, 8° W	Graf et al. (1990)	Quantitative trend determination not possible
Filchner Ice Shelf: Druzhnaya Is.	78° S, 40° W	Same as above	Inferred from 1952–85 mean rate
Ronne Ice Shelf	79° S, 55° W	Graf et al. (1988)	Annual values

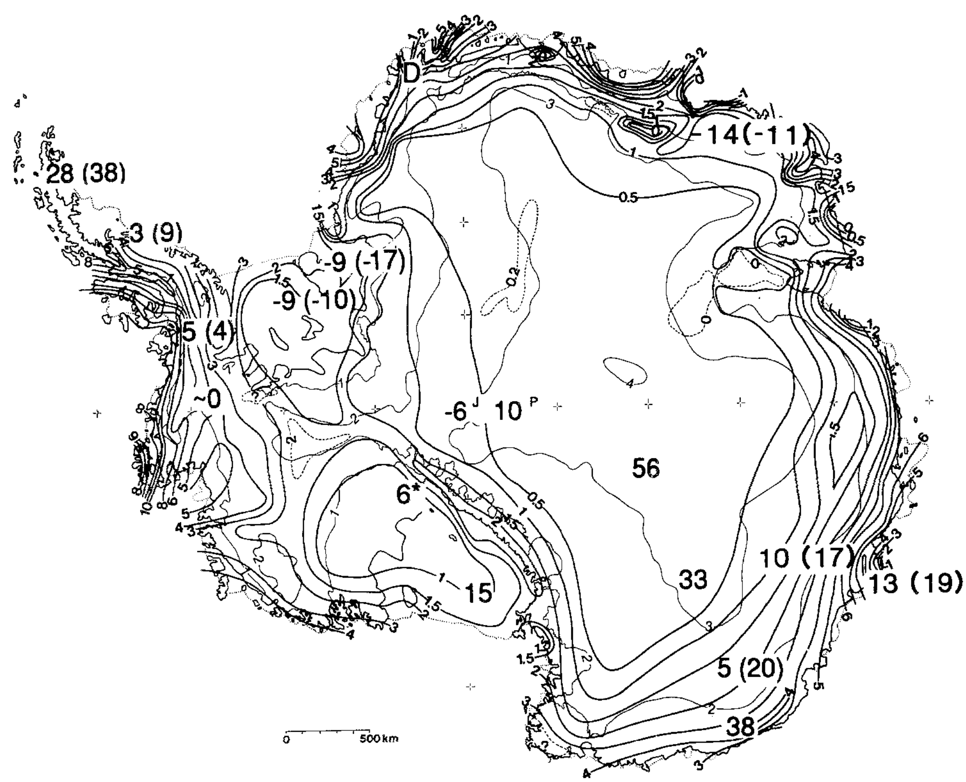

Fig. 2. Percentage changes in accumulation (large numerals) from the 1955–65 yearly mean to the 1965–75 yearly mean and from the 1950–60 yearly mean to the 1970–80 yearly mean (in parentheses, where available) for locations given in Table 1. The South Pole estimates based on Jouzel et al. (1983) and Pourchet et al. (1983) are indicated by J and P superscripts, respectively. D stands for the recent accumulation decrease at Neumayer. The * for the Ice Stream B catchment area estimate is based on the change between the 1955–65 and 1965–85 yearly mean, and is not statistically significant over the sampling sites in the catchment area. The background map consists of Antarctic surface elevations (thin lines, every 1000 m) and surface mass balance rates (thick lines, in units of $10^2\,\mathrm{kg\,m^{-2}\,yr^{-1}}$, adapted from Giovinetto and Bentley (1985))

continent is presented in Section 2.4. If a uniform linear trend in time was present at each site then the percentage changes for the longer period should be twice as large as those for the shorter period. For the nine sites where changes over the longer period could be determined only three showed twice the increase or decrease of the shorter period. Thus, it appears that the Antarctic rate of snowfall increase is smaller for the 1950–60 to 1970–80 period than for 1955–65 to 1965–75.

2.3 Meteorological Analyses

Connolley and King (1993) used radiosonde observations transmitted over the Global Telecommunications System to calculate the water vapor transports along the coast of Antarctica for 1980–82 and 1988–90. Their six-year average patterns approximately agreed with those obtained by Bromwich (1988). Of particular relevance here is that their transport values can be used to derive the net precipitation (precipitation minus sublimation/evaporation, or $P-E$) from the atmospheric water balance equation for the same region as studied by Bromwich (1988, 1990) for 1972. Thus, the calculated net precipitation change from 1972 to $\sim$1985 provides a comparison with the accumulation trends outlined above.

The atmospheric water balance equation for annual time scales can be written (Peixoto and Oort, 1983; Bromwich, 1988)

$$\langle \bar{P} - \bar{E} \rangle = -\frac{1}{A} \oint \left\{ \int_0^{p_s} \frac{\overline{qV}}{g}\, dp \right\} \cdot \boldsymbol{n}\, dl \tag{1}$$

where A is the area over which the net precipitation is to be determined, V is the wind vector, q is the specific humidity, and g is the acceleration due to gravity; the angle brackets denote a spatial average, and the overbar a time average. The water vapor flux is integrated from the surface ($p=p_s$) to the top of the atmosphere ($p=0$). The

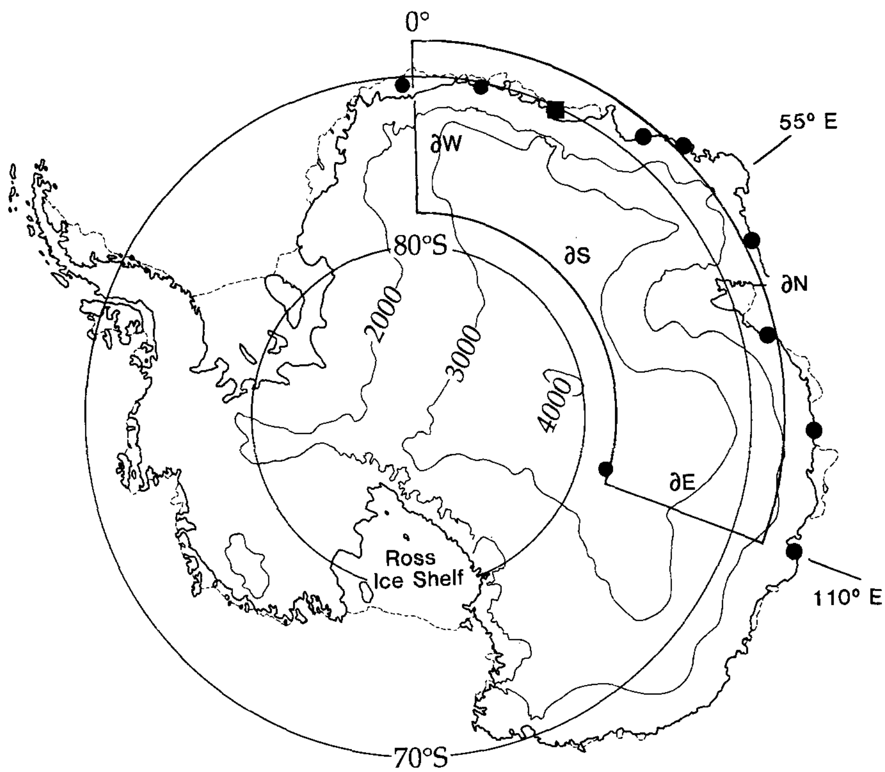

Fig. 3. 0° to 110° E sector of Antarctica for which the moisture budget results in Table 2 apply. Solid circles and square indicate radiosonde stations used. The solid square represents Roi Baudouin station, which ceased operations in the 1960's, but for which synthetic values for 1972 and 1985 were generated using earlier data (Bromwich, 1979). The north, south, east, and west boundaries of the largest moisture budget area are indicated with the notation used in the text and Table 2

unit vector $\boldsymbol{n}$ is the outward normal to the periphery. Equation 1 states that the net convergence of atmospheric moisture into the volume overlying area A must equal the net flux of moisture to the surface (i.e., the net precipitation). By carrying the integration to the top of the atmosphere the moisture flux through the top of the volume must be zero; the change of moisture storage within the atmospheric volume is negligible on these time scales.

The line integral is carried out around the periphery of area A along parallels of latitude (northern and southern boundaries) and along meridians (western and eastern boundaries). The decomposition of (1) along these four boundaries takes the form

$$\langle \bar{P} - \bar{E} \rangle = -\frac{1}{A}\left[\int_{\partial \mathrm{N}} Q_\phi dl - \int_{\partial \mathrm{S}} Q_\phi dl + \int_{\partial \mathrm{E}} Q_\lambda dl - \int_{\partial \mathrm{W}} Q_\lambda dl\right] \tag{2}$$

where

$$Q_\phi = \int_0^{p_s} \left(\frac{\overline{qv}}{g}\right) dp; \quad Q_\lambda = \int_0^{p_s} \left(\frac{\overline{qu}}{g}\right) dp \tag{3}$$

with v and u being the meridional and zonal components of $\boldsymbol{V}$, respectively, and ∂N, ∂S, ∂E, and ∂W refer to the integral taken along the north, south, east, and west boundaries, respectively.

Figure 3 shows the radiosonde locations used by Bromwich (1988, 1990). For the sector 0° to 110° E only the northern boundary is well monitored by radiosonde stations, spaced at an average

Table 2. *Net Moisture Transport Convergence (mm yr^{-1}) Across the Boundaries Defined by Bromwich (1988).* 1972 values are after Bromwich (1990). ~1985 values are derived from Connolley and King (1993) as described in the text. Sum represents $\langle \bar{P} - \bar{E} \rangle$. Accumulation values (given by Bromwich, 1990) are derived from the analysis of Giovinetto and Bentley (1985) and are based on average accumulation values for a few to several years between 1940 and 1980 (Giovinetto et al., 1992)

Area	Year	∂N	∂S	∂E	∂W	Sum
0°–110° E,	1972	158	3	−6	−50	105
68.4° S–78.2° S	~1985	168	5	−6	−59	108
Accumulation						108
0°–55° E,	1972	240	−8	−1	−100	131
69.3° S–76.8° S	~1985	160	−8	−1	−118	33
Accumulation						109
55° E–110° E,	1972	102	9	−4	−2	105
67.2° S–79.8° S	~1985	171	14	−4	−2	179
Accumulation						108

of 660 km apart. There are no radiosonde stations on the eastern and western boundaries, and only one on the southern boundary. The longitudinal boundaries were selected to enclose the well sampled coastal section of East Antarctica and the southern boundary was located at the average latitudinal position of the crest of the terrain. As outlined previously (Bromwich, 1988), the transports across the eastern, western and southern boundaries were estimated by combining available radiosonde data with the climatological data given by Taljaard et al. (1969) and Jenne et al. (1971). Over the ice sheet above ~2500 m elevation the transports are mostly small because the lower tropospheric layer where most of the moisture resides is no longer present. Table 2 summarizes the net moisture transports into the section lying between 0° and 110° E, and for two subsections of this area. For 0° to 110° E, the transport across the northern boundary dominates the budget, but a very important secondary contribution is the net outflow across the western boundary. The latter arises because the parallel approximating the coastline is located to the north of the coast at the Greenwich meridian (see Fig. 3); this component is almost entirely due to the large coastal westward transport which is monitored by SANAE station (2.4° W) observations. This component becomes even more important for 0° to 55° E, but does not influence the subsection from 55° E to 110° E. The small convergence contribution from the southern boundary arises because of the equatorward transport at Vostok.

To derive the corresponding moisture transport convergence values for the Connolley and King (1993) analysis, the 1972 ∂N value was multiplied by the ratio of the longitudinal average of the coastal meridional transports obtained by Connolley and King (1993) for ~1985 to those obtained by Bromwich (1988, 1990) for 1972. Synthetic values for coastal Roi Baudouin site (24.3° E, not operational since the 1960's), following the procedures given by Bromwich (1979), were derived for the Connolley and King (1993) period and incorporated into the longitudinal average for ~1985. The ~1985 ∂W value was derived by multiplying the 1972 ∂W value by the ratio of the zonal moisture transports found for SANAE. The ~1985 ∂S value was derived from the 1972 ∂S value by the ratio of the meridional transports at Vostok.

Table 2 shows that the $P-E$ value derived from the ~1985 moisture budget for 0° to 110° E does not differ significantly from that obtained by Bromwich (1988, 1990) for 1972 nor from the derived accumulation value which represents a composite over many years. This result is probably consistent with the accumulation trend analysis in Fig. 2 as decreases are found in the western part of the domain and increases in the eastern part. Confirmation for this hypothesis is found by looking at the convergence calculations for the two subsections; the western area reveals a very marked net precipitation decrease since 1972 and the eastern one a sharp increase. These roughly 75% changes approximately cancel for 0° to 110° E, and are primarily caused by changes in the meridional transports across ∂N. The calculated net precipitation changes are qualitatively consistent with the accumulation trends. However, they appear to be much too large, although it is difficult to be sure because the time periods compared are not the same and the spatial coverage differs. This finding calls for a much more careful treatment of the convergence calculations than presented here, particularly with respect to the impact of missing soundings which preferentially occur during storms (Bromwich, 1979) and of a mass budget constraint (Alestalo, 1983). Neither factor was considered by Connolley and King (1993), and the latter was not evaluated by Bromwich (1988, 1990).

It is uncertain how the moisture transport convergence results are influenced by the potentially large errors in radiosonde moisture content measurements at low temperatures (Elliot and Gaffen, 1991). Connolley and King (1993) estimate their fluxes to have an uncertainty of 20%, with the dominant contribution coming from moisture measurement uncertainties. However, they also note some locations where the moisture amounts are likely to be systematically underestimated, as well as others where the amounts are likely overestimated. This suggests some error cancellation. Furthermore, the largest poleward fluxes occur at the coast in conjunction with storms, whose comparatively warm conditions minimize the moisture measurement uncertainties.

Additional confirmation from the meteorological record was sought by examining the temporal trend in synoptic variability around the Antarctic coast because, at least for East Antarctica,

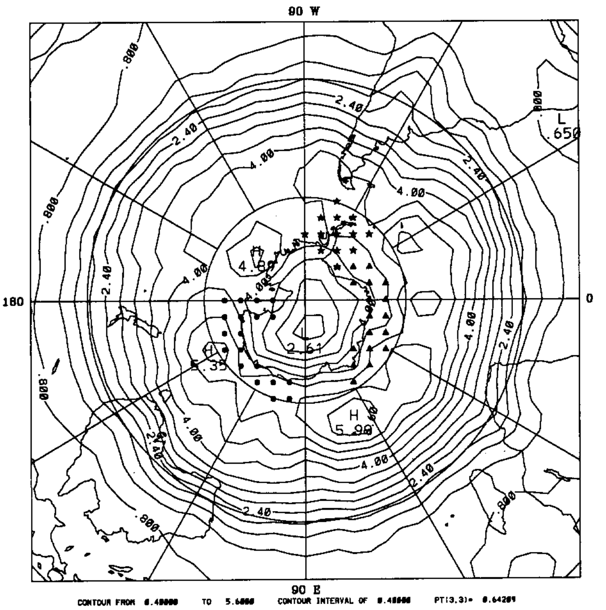

Fig. 4. Daily mean variability on synoptic time scales (2–6 days, contoured every 0.4 hPa) of sea level pressure from the Australian Bureau of Meteorology analyses for 1972–89. The solid stars, triangles, and circles indicate the grid points used to diagnose the variability trends shown in Fig. 5

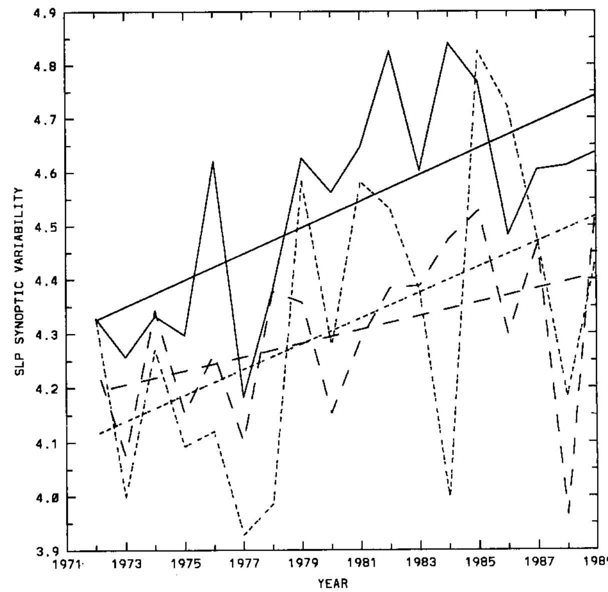

Fig. 5. 1972–89 time series of synoptic variability (in hPa) of sea level pressure for the three areas of Fig. 4, with best-fit linear trends: 100° E–180° (circles) = solid, Peninsula area (stars) = large dash, 40° W–60° E (triangles) = small dash

the poleward eddy flux is the dominant component leading to ice sheet precipitation (Bromwich, 1988). The twice-daily numerical analyses produced by the Australian Bureau of Meteorology are available from April 1972 to December 1989. As a result they cover only the later part of the interval spanned by Fig. 2, and essentially test the accumulation tendency for recent increases. The synoptic variability is represented by the standard deviation of the 2 to 6 day band pass filtered sea-level pressure field (Duchon, 1979) and was averaged by year. Figure 4 presents the 18-yr average pattern. The maximum values lie along 50–60° S where the cyclones travel quickly and are close to their maximum intensity. The synoptic variability decreases toward the coast of Antarctica near where the circumpolar trough is located; this is a region of slow-moving and dissipating synoptic-scale cyclones (Streten and Troup, 1973). Three coastal sectors based upon the above accumulation analysis (see Fig. 2) have been chosen to examine the trends in synoptic variability: 100° E to 180° (the solid circles in Fig. 4), the Antarctic Peninsula region (the stars in Fig. 4), and from 40° W through the Greenwich meridian to 60° E (the triangles in Fig. 4). The first and third sectors are areas of increased accumulation while in the second area the accumulation has decreased. The sectors have been chosen so that both the ice-sheet accumulation trends and the coastal meteorological conditions are reasonably well defined; consequently, the meteorologically data sparse West Antarctic sector is omitted.

In all three sectors the synoptic variability shows a statistically significant (better than 91%) upward trend (Fig. 5), consistent with the idea advanced by Morgan et al. (1991) that the accumulation over Antarctica has recently increased. On the regional scale, the synoptic variability and accumulation trend analyses both point to increased accumulation for 100° E to 180° and the Antarctic Peninsula area. The former result agrees with Morgan et al. (1991) who reported an increase in cyclonic activity near 110° E that accompanied their measured accumulation increase. However, for the area between 40° W and 60° E the two approaches apparently disagree. It is possible that if accumulation data were available from this sector for the period 1972–89 that an overall increase would be found; this is definitely not the case for the site at 40° W, but some support for this situation is found at the sites at 55° W and 45° E. From the above evaluation of the atmospheric moisture budget, it was found for 0° to 55° E that the precipitation rate has sharply decreased between 1972 and about 1985. Clearly, there is an urgent need to have glaciological and meteorological observations cover the same time period.

As a step towards checking the reliability of the time series of synoptic variability derived from the Australian analyses, the same quantity for the three areas of Fig. 4 was computed from the European Center for Medium-Range Weather Forecasts (ECMWF) sea level pressure fields. These data were only available for 1980–89. The shortness of this time period makes definitive comparisons difficult. No major differences between the time series from the two analyses were found, and some correspondence in interannual variations was in evidence. Most of the linear trends in synoptic variability from both analyses were not statistically different from zero. For the Australian analyses, this was in spite of the clear trends towards increasing variability over the longer 1972–89 period (Fig. 5). It is interesting to note that 1979 was the year of enhanced data coverage associated with the Global Weather Experiment (Hollingsworth, 1989), and that improvements in numerical analyses followed this event. Thus the trends in the Australian analyses may to some extent be an artefact. Trenberth and Olson (1988) have discussed spurious changes in numerical analyses over the Southern Hemisphere produced by ECMWF and NMC, with major discrepancies between the two analyses over and near Antarctica. These operational products are continually being upgraded and this can result in artificial shifts that are of major importance for climatic analyses. More work is needed concerning possible changes of this type in the Australian analyses. However, the areas chosen for the synoptic variability analyses cover well observed Antarctic coastal regions that had stable station coverage for 1972 to 1989, thus station changes likely had no artificial influence on the variability trends.

2.4 Implications

Given the continent-wide temporal accumulation changes (Fig. 2) and the results of the moisture budget analysis for part of East Antarctica (Table 2), an accumulation change for the entire Antarctic continent can be found. The ice sheet was divided into subsections with borders approximately equidistant between the adjacent points of Fig. 2. The moisture budget area shown in Fig. 3 was kept intact as its own subsection. Only the 1955–65 to 1965–75 accumulation changes were used, as these were more numerous than the 1950–60 to 1970–80 changes. Each accumulation change was multiplied by the total average accumulation for the subsection (in $\mathrm{kg\,m^{-2}\,yr^{-1}}$, from the mean accumulation map of Giovinetto and Bentley, 1985) to yield a change in mass for that subsection. The 0% moisture budget change for 1972–~85 was assumed to apply to the 1955–65 to 1965–75 period as well. A 5% accumulation decrease was used for Neumayer, and a 0% change at South Pole. Accumulation changes over the floating ice shelves (which approximately cancel) were not included, nor were accumulation changes over the tip of the Antarctic Peninsula, where average accumulation data were not available. The accumulation changes for all the subsections were integrated to yield a ~5% increase in accumulation over Antarctica as a whole. This number varies by ±2.5% if the above assumptions are altered realistically. Given the large number of assumptions, and the differing time periods involved, this number is used for scaling purposes only, and to set the large accumulation increases at individual points into a continental context. Based upon an average temperature increase from 1957–82 of 0.74 °C for all of Antarctica (Raper et al., 1984), a relation of ~18% $^{\circ}\mathrm{C}^{-1}$ is found between Antarctic precipitation and temperature variations. This estimate is larger than other estimates of the association between changes of precipitation and temperature, consistent with the findings of Morgan et al. (1991).

To summarize, both glaciological and meteorological observations generally support the idea that the precipitation rate over Antarctica has increased in recent decades. The likely cause is systematic increases in the cyclonic forcing around the coast of the continent, and is only weakly reflected in mean air temperatures (Morgan et al., 1991). The data coverage in space and time with both approaches leaves a great deal to be desired, and a null result can not be ruled out. The proposed International Trans-Antarctic Scientific Expedition (ITASE, 1992) project to collect accumulation rate observations over Antarctica from oversnow traverses is a very timely effort and coupled with modern atmospheric numerical analyses may supply the data sets necessary to describe and understand continental precipitation changes.

3. Greenland

3.1 Background

The Greenland Ice Sheet is located on the northern fringes of the Atlantic Ocean and is flanked to the west by North America and to the east by western Europe. Surprisingly, the atmospheric behavior over the ice sheet, especially as regards precipitation, remains poorly understood (e.g., Scorer, 1988). In part, this has to do with the availability of observational data which are abundant from the populated ice-free coastal margins, but are sparse from the ice sheet proper. Recent deployments of satellite-transmitting automatic weather stations (AWS) are alleviating some of this deficiency (Weidner and Stearns, 1991). Glaciological snow accumulation measurements have been made in many locations but cover widely differing time periods. Extensive efforts are being made to retrieve deep ice cores from the Summit region of the ice sheet, an area not well understood meteorologically; the goal is to describe and understand climate change on the glacial-interglacial time scale (e.g., Johnsen et al., 1992).

As a focal point for this overview of snowfall variations over Greenland over the past several decades, we consider the results of a model used to simulate precipitation over the Greenland Ice Sheet (Bromwich et al., 1993). The model is essentially based upon daily radiosonde data from the Greenland region. The coverage of such data here is adequate for the present purposes given the smaller area of the Greenland Ice Sheet as opposed to Antarctica. Throughout the time period to which the model was applied at least seven radiosonde stations were operational surrounding the ice sheet. The results from this model provide important information concerning recent precipitation trends over this ice sheet, and attempts to verify these results will serve to summarize precipitation trend information from other, more conventional, sources.

3.2 Precipitation Model: Description and Results

The diagnostic model mentioned above parameterizes precipitation as essentially the product of vertical motion and a measure of atmospheric moisture content. The vertical atmospheric motion is divided into that arising from the dynamics of synoptic-scale atmospheric circulation systems and that due to uplift of air over the terrain of Greenland. The first or "dynamic" component is based upon the flux of relative geostrophic vorticity (or "Vorticity Flux Index"; Keen, 1984) at the 500-hPa level (calculated from daily 500-hPa geopotential height fields for 1963–88 analyzed by the National Meteorological Center, or NMC) and makes use of the equivalent-barotropic assumption. The second or "orographic" component is based simply upon the component of the 850-hPa geostrophic wind parallel to the terrain gradient vector at each point (Alpert and Shafir, 1991). It is only significant over the steep coastal ice sheet slopes. The final form of the model is

$$\text{Precipitation} = A_0 q|\boldsymbol{VFI}| + A_1 q \boldsymbol{V}_{850} \cdot \nabla H \qquad (4)$$

where q is the climatological specific humidity at 700 hPa, $\boldsymbol{VFI}$ is the vorticity flux index, $\boldsymbol{V}_{850}$ is the geostrophic wind at the 850-hPa level (also based on geopotential heights from NMC), H is the terrain height, and A_0 and A_1 are empirical coefficients. These coefficients are used to fix the

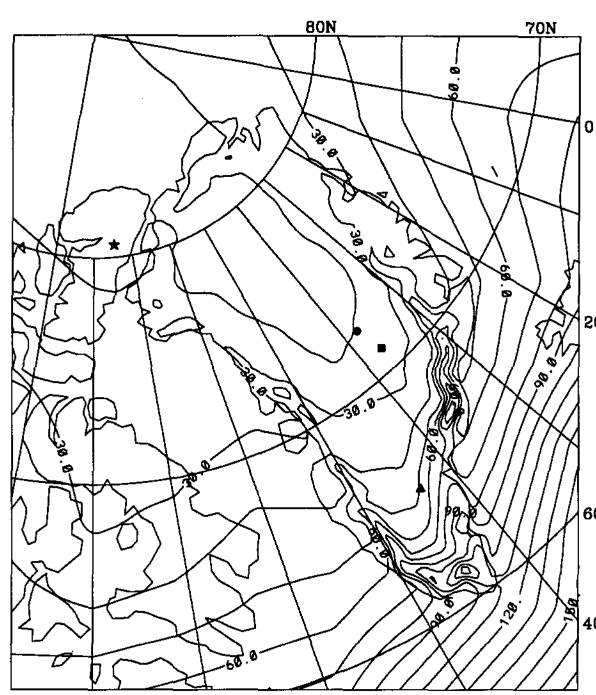

Fig. 6. Modeled mean annual precipitation for Greenland, contoured every 10 cm w.e. Also indicated are the locations of Summit (circle), Crete (square), Dye 3 (triangle), and Agassiz (star)

model output to actual precipitation values, and are derived primarily through linear regression of modeled precipitation against corresponding observed accumulation values found from glaciological investigation near Summit (see Fig. 6), where an array of shallow ice cores were drilled and the resulting accumulation measurements areally averaged into one value for each of the years 1964–86 (Bolzan and Strobel, 1993). Both components are multiplied by a monthly climatological specific humidity value at each grid point to incorporate the annual cycle of the water vapor content of the lower troposphere.

The most striking result of the application of this model to the period 1963–88 is a trend in time towards less total precipitation over the ice sheet which amounts to roughly 15% (Fig. 7). No trend towards increasing observed mean specific humidities was found which would counteract the modeled trend. This trend is not uniformly manifested over the ice sheet, but varies from point to point. Figure 8 shows the linear trend in modeled precipitation for 1963–88 for each grid point in and near Greenland, expressed in both cm w.e. and percent of mean annual modeled precipitation. Statistically significant decreases are seen over most of the northern 2/3 of the ice sheet. The largest trends relative to mean annual precipitation are observed in the northern interior. There is

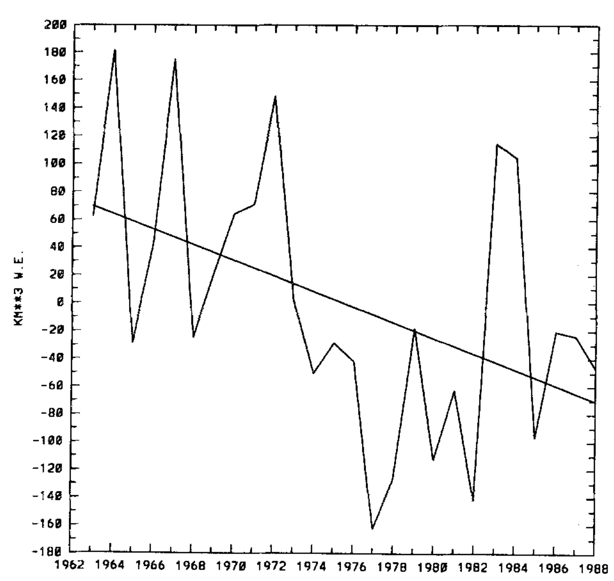

Fig. 7. Time series of the yearly volume of simulated precipitation over the Greenland Ice Sheet, expressed in km^3 w.e. deviation from the 1963–88 mean (733). The best-fit linear trend, significant at the 95% level, is also superimposed

some evidence for positive trends in precipitation in the far northern and eastern portions of Greenland. From Fig. 7 it can also be seen that the modeled precipitation displays a large degree of interannual variability, the precipitation often changing by 30% from one year to the next. This variability shows a 3–5 year periodicity.

Vigorous efforts have been taken to establish the validity of these modeled trends in precipitation. Data used for these comparisons include ice-core derived accumulation values and coastal gauge precipitation measurements. All of these data possess limitations which make comparisons with the modeled precipitation difficult. The modeled precipitation amounts are essentially based upon 500-hPa heights which were obtained from NMC on a roughly 400-km grid, and comparisons to precipitation or accumulation observations at a single point likely stretch the capability of the model to the limit.

3.3 Model Verification: Meteorological Data

Changes in the large-scale atmospheric flow corresponding to this modeled trend towards smaller precipitation rates were sought through examining changes in the synoptic-scale (2–6 days) variability of the 500-hPa height field (through use of the same band-pass filtering scheme mentioned above). Regions of high synoptic-scale variability roughly correspond to "storm tracks". It was found that for years of higher Greenland precipitation the axis of maximum variability near Greenland (in the vicinity of the "Icelandic Low") was more intense and closer to Greenland than during periods of lower precipitation over Greenland, as would be expected.

Corroboration of the trends over the Greenland Ice Sheet through the use of coastal gauge records was much more problematic. These measurements were obtained for Greenland coastal stations, as well as for some Icelandic stations, from *Monthly Climatic Data for the World*. No attempt was made to directly compare modeled and observed precipitation magnitudes, as gauge measurements of snowfall notoriously under-represent actual values (Groisman et al., 1991; Woo et al., 1983). The coastal reporting stations in Greenland are also located in or adjacent to complex mountainous topography, and the small scale variations which result from these station locations are not known. As evidence of these problems, Barry and Kiladis

(1982) found annual precipitation amounts to be poorly correlated between the various coastal stations, even between those located less than 10 km apart. However it was hoped that these gauge measurements might adequately portray interannual trends in precipitation, assuming the type and exposure of the gauge remained the same throughout the period. To further complicate the use of these data, no precipitation measurements for the Greenland stations were available from *Monthly Climatic Data for the World* for the years 1971–75. Time series from these precipitation observations and from the nearest model grid points were compared. Both agreement and disagreement were found. It is not at all clear whether the model should be applied to coastal sites as low-level blocking should strongly affect precipitation generation in these areas (Bromwich, 1988; Overland, 1984; Hsu, 1988, pp. 163–165), and this mechanism is not included in the model. Given all these caveats, no firm conclusions can yet be drawn from these comparisons, and no coherent picture of precipitation trends over the Greenland Ice Sheet from gauge precipitation measurements was gained.

3.4 Model Verification: Glaciological Data

As an overall check of the model results against glaciological accumulation records, the mean annual precipitation over the entire ice sheet as simulated by the model was computed, converted into accumulation, and compared against various estimates of mean annual accumulation over the ice sheet based on glaciological data. The model's value was seen to be within the range of values obtained by other investigators. The model's mean annual spatial distribution of precipitation over Greenland was also plotted (Fig. 6) and compared to a spatial distribution of mean annual accumulation from Bender (1984). This comparison revealed that the model captures the essential features of the accumulation distribution: the general trend of high values in the south to low values in the north, zones of high precipitation along the southern coastal margins, and a large area of very low precipitation dominating the northern interior of the island. The major shortcoming of the simulated precipitation distribution is the lack of a band of higher accumulation values along the northern west slopes of the ice sheet. The model seriously underestimates the precipitation in this region, and it is believed that this is due to a shortcoming in the NMC analyses: namely that they underrepresent the intensity of the storm track along the western coast of Greenland, a shortcoming also observed by Serreze and Barry (1988).

Glaciological data with which to ascertain recent precipitation changes in Greenland are surprisingly sparse. Only ice cores from Summit, Dye 3, and Crete (see Fig. 6) provided accumulation data coverage to overlap with 1963–88. There are many data concerning interglacial, long-term climatic variations, but far fewer high temporal resolution accumulation records from the recent past.

Recent accumulation values from an array of shallow ice cores near Summit (Bolzan and Strobel, 1993), spatially averaged into one value, provided the data necessary to tie the output from the precipitation model to actual precipitation amounts (through linear regression between model output and these accumulation data). No trend in time over the 23 years from 1964–86 towards higher or lower accumulation values was found in the spatially averaged Summit data (although a trend towards lower precipitation rates was simulated for 1963–88 for this location, see Fig. 8), a result that matches the modeled precipitation trend for 1964–86 here. (A matching trend in time between the two time series was not forced by the linear regression.)

The trends from two other Greenland ice-core records were also examined for consistency with model results. Accumulation records from Crete (Clausen et al., 1988) and Dye 3 (Reeh et al., 1977) were available for portions of the 1963–88 period. Both records showed mean annual accumulation rates to be decreasing over periods of time overlapping the 1963–88 period; for both sites, these decreases were part of a longer term annual accumulation decrease which started in the late 1930's. Modeled annual precipitation rates for the corresponding time periods were also seen to be decreasing.

Through correspondence with David Fisher of the Geological Survey of Canada we obtained a preliminary analysis of recent accumulation rates taken from the Agassiz site on Ellesmere Island in the northeast Canadian Arctic Islands. These data show increasing accumulation rates over the 1963–88 period (amounting to about a 35% increase). The trend in modeled precipitation rates for this

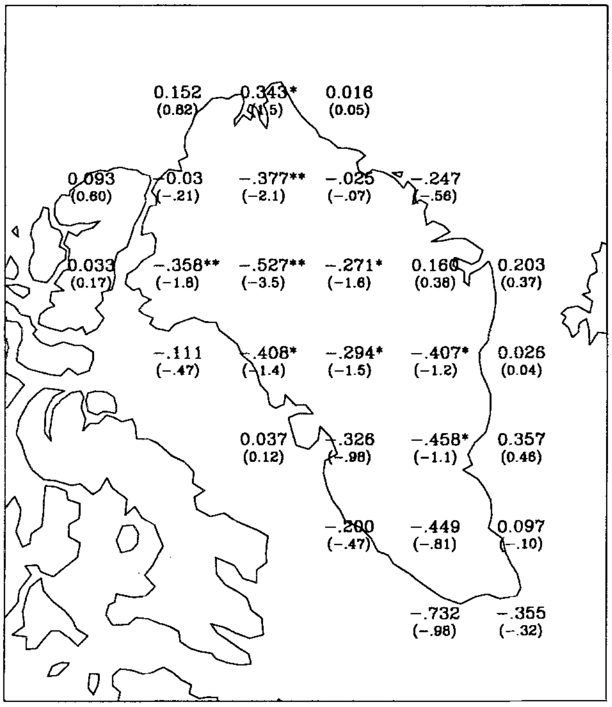

Fig. 8. 1963–88 linear trends expressed in cm w.e. yr^{-1} (top) and yearly percentage of modeled annual mean precipitation (bottom, in parentheses) for grid points in and near Greenland. Trends significant at the 95% level are indicated by *, and at the 99% level by **

location is very slightly positive (though not statistically significant), and some consistency between the model and ice-core accumulation values is observed.

Work carried out by Zwally (1989) found, using satellite radar altimetry, that the southern half of the ice sheet (south of 72° N) is thickening at about 200 mm yr^{-1}. This was attributed to precipitation increases over the decade to century time scale. This finding is being debated (e.g., Douglas et al., 1990). The satellite measurements were taken over 1978–86, a time period which exhibits a secondary upturn in modeled precipitation (see Fig. 7), primarily over southern Greenland. However, the elevation changes south of 72° N which would result from the modeled precipitation trend for 1978–86 would only amount to ~10% of Zwally's increase. Thus the altimetry-based results may be reflecting to some extent recent precipitation changes.

3.5 Implications

This trend has important implications for global sea level, which appears to be rising at the rate of ~2 mm yr^{-1} (Douglas 1991). Sensitivity studies have indicated that falls of ~0.1 mm yr^{-1} in global sea level would result from a 5% increase in precipitation over Greenland (Warrick and Oerlemans, 1990; ignoring effects of corresponding changes in temperature and/or cloud cover). Thus, considered in isolation, the modeled trend implies a possible contribution to global sea level rise of ~0.3 mm yr^{-1} by 1988 due to precipitation over Greenland if this ice sheet was in balance in 1963.

It is starting to be appreciated that Greenland is an area of large spatial and temporal variations in climate. Chapman and Walsh (1993) examined the linear trends in annual surface air temperature poleward of 40° N for 1961–90. Pronounced cooling (around 1 °C) was found in central and southern Greenland and over the ocean areas to the west and east. Along the northeast coast of Greenland warming of a similar magnitude was found. The Summit area of the Greenland Ice Sheet, where the deep ice cores mentioned above are being drilled, was close to the transition zone. The cooling over Greenland during the last 30 years contrasts with the remainder of the Northern Hemisphere north of 60° N. Correspondence between these temperature changes and the spatial pattern of modeled precipitation trends (Fig. 8) is observed: an area of increases to the north of a large area of decreases. The modeled precipitation changes can not, however, be simply explained by the temperature changes as these precipitation changes (~25% per °C) are much larger than those indicated by the usually cited relationships of precipitation to temperature. Rather, precipitation and temperature changes are both related to the complex dynamics of the cyclones affecting this region.

4. Discussion

It has been shown that within the last three decades the precipitation rate appears to have been increasing over Antarctica, and decreasing over Greenland with a secondary upturn since 1977. Both trends (+5% over Antarctica over a ten year period; −15% over Greenland over a 26 year period) are caused by changes in the cyclonic forcing, and are only weakly reflected in the variables often thought to correlate with annual precipitation amounts. The sea-level change equivalent of precipitation over Antarctica is

5.8 mm yr^{-1} (Budd and Simmonds, 1991) and for Greenland is 1.4 mm yr^{-1}. Thus the above results suggest that the accumulation changes over the two ice sheets have offsetting effects on global sea level change, with the changes over Antarctica (5% times 6 mm yr^{-1}) overwhelming those over Greenland (−6% times 1.5 mm yr^{-1}) and resulting in a net 0.2 mm yr^{-1} being taken from the global ocean by the end of the ten-year period. This is of course considering only precipitation and ignoring all other ice-sheet mass balance effects. However, these conclusions are not founded on quality data bases. The importance of such changes for global sea level strongly argues for vigorous efforts to resolve the many outstanding questions. In particular, complementary analyses of meteorological and glaciological observations would not only describe the precipitation changes that are taking place but also why these are occurring.

For Antarctica, several new efforts could be contemplated. First, the large gap of radiosonde stations along the coast of West Antarctica, with few prospects for improvement, argues for vigorous efforts to exploit satellite data over the ocean to calculate the poleward flux of water vapor across the entire continental periphery, and thus use the atmospheric water budget to monitor large scale precipitation changes over Antarctica. A very promising sensor is the SSM/I (Special Sensor Microwave/Imager) instrument on the DMSP (Defense Meteorological Satellite Program) polar orbiting meteorological satellites, whose retrievable quantities over water surfaces include integrated water vapor amounts, integrated cloud water amounts, and surface wind vectors (Alishouse et al., 1990; Wentz, 1992). Second, the model approach described above to calculate precipitation amounts over Greenland can not presently be used for Antarctica because the sparse data base renders the 500-hPa analyses unreliable (Godin, 1977; Trenberth and Olson, 1988). However, the method presented by Phillpot (1991) to calculate the 500-hPa geopotential heights over Antarctica from AWS observations at sites above 2000-m elevation has shown impressive reliability (e.g., Carrasco and Bromwich, 1993). Should a dense enough AWS network be deployed and be used routinely to help construct the 500-hPa analyses over the continent, the Bromwich et al. (1993) approach could be used to infer precipitation variations over Antarctica, provided a parameterization for clear-sky precipitation was included. Third, planned efforts to systematically collect accumulation observations over Antarctica, like ITASE, deserve full support and encouragement.

For Greenland, the meteorological analyses are progressing and a full atmospheric moisture budget study is underway to complement the Bromwich et al. (1993) approach. The paucity of records of accumulation rate variations could in theory be redressed with a modest research effort. The ice sheet is readily accessible by aircraft flying out of the numerous near-coastal airfields (Brozena et al., 1993). The limited size of the ice sheet means that a systematic aircraft-based shallow ice coring survey could be completed in a few summer field seasons to describe the recent precipitation changes over Greenland and thus provide the definitive data base to test the results from the meteorological analyses.

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References

Alestalo, M., 1983: The atmospheric water budget over Europe. In: Street-Perrott, A., Beran, M., Ratcliffe, R. (eds.) *Variations in the Global Water Budget*. Hingham, Mass.: D. Reidel, pp. 67–79.

Alishouse, J. C., Snyder, S. A., Vongsathorn, J., Ferraro, R. R., 1990: Determination of oceanic total precipitable water from the SSM/I. *IEEE Trans. Geosci. Remote Sensing*, **28**, 811–816.

Alpert, P., Shafir, H., 1991: Role of detailed wind-topography interaction in orographic rainfall. *Quart. J. Roy. Meteor. Soc.*, **117**, 421–426.

Barry, R. G., Kiladis, G. N., 1982: Climatic characteristics of Greenland. In: Radok, U., Barry, R. G., Jenssen, D., Keen, R. A., Kiladis, G. N., McInnes, B. (eds.) *Climatic and Physical Characteristics of the Greenland Ice Sheet*. Boulder, Colorado: CIRES, Univ. of Colorado, pp. 7–33.

Bender, G., 1984: The distribution of snow accumulation on the Greenland Ice Sheet. M.S. thesis, University of Alaska, Fairbanks, Alaska, 110 pp.

Bolzan, J. F., Strobel, M., 1993: Accumulation rate variations around Summit, Greenland. *J. Glaciol.* (in press).

Bromwich, D. H., 1979: Precipitation and accumulation estimates for East Antarctica, derived from rawinsonde information. Research report, Dept. of Meteorology, Univ. of Wisconsin, Madison, Wisconsin, 142 pp.

Bromwich, D. H., 1988: Snowfall in high southern latitudes. *Rev. Geophys.*, **26**, 149–168.

Bromwich, D. H., 1990: Estimates of Antarctic precipitation. *Nature*, **343**, 627–629.

Bromwich, D. H., Robasky, F. M., Keen, R. A., Bolzan, J. F., 1993: Modeled variations of precipitation over the Greenland Ice Sheet. *J. Climate* (in press).

Brozena, J., Chalona, M., Forsberg, R., Mader, G., 1993: New technology allows remoter areas to be probed. *Eos, Transactions, Amer. Geophys. Union*, **74**, 18.

Bryazgin, N. N., 1986: Method of preparing monthly charts of atmospheric precipitation in Antarctica. In: Dolgin, I. M. (ed.) *Climate of Antarctica*. New Delhi: Oxonian, pp. 109–116.

Budd, W. F., Simmonds, I., 1991: The impact of global warming on the Antarctic mass balance and global sea level. In: Weller, G., Wilson, C. L., Severin, B. A. B. (eds.) *Proceedings of the International Conference on the Role of the Polar Regions in Global Change, Vol. 1*. June 11–15, 1990, University of Alaska Fairbanks, Fairbanks, Alaska, 489–494.

Carrasco, J. F., Bromwich, D. H., 1993: Mesoscale cyclogenesis dynamics over the southwestern Ross Sea. *J. Geophys. Res.* (in press).

Chapman, W. L., Walsh, J. E., 1993: Recent variations of sea ice and air temperature in high latitudes. *Bull. Amer. Meteor. Soc.*, **74**, 33–47.

Clausen, H. B., Gundestrup, N. S., Johnsen, S. J., Bindschadler, R., Zwally, H. J., 1988: Glaciological investigations in the Crete area, central Greenland: a search for a new deep-drilling site. *Annals Glaciol.*, **10**, 10–15.

Connolley, W. M., King, J. C., 1993: Atmospheric water-vapour transport to Antarctica inferred from radiosondes. *Quart. J. Roy. Meteor. Soc.*, **119**, 325–342.

Douglas, B. C., 1991: Global sea level rise. *J. Geophys. Res.*, **96**, 6981–6992.

Douglas, B. C., Cheney, R. E., Miller, L., Agreen, R. W., Carter, W. E., Robertson, D. S., 1990: Greenland Ice Sheet: Is it growing or shrinking? *Science*, **248**, 288.

Duchon, C. E., 1979: Lanczos filtering in one and two dimensions. *J. Appl. Meteor.*, **18**, 1016–1022.

Elliot, W. P., Gaffen, D. J., 1991: On the utility of radiosonde humidity archives for climate studies. *Bull. Amer. Meteor. Soc.*, **72**, 1507–1520.

Giovinetto, M. B., Bently, C. R., 1985: Surface balance in ice drainage systems of Antarctica. *Antarctic J. U.S.*, **20**(4), 6–13.

Giovinetto, M. B., Bromwich, D. H., Wendler, G., 1992: Atmospheric net transport of water vapor and latent heat across 70° S. *J. Geophys. Res.*, **97**, 917–930.

Godin, R. H., 1977: An investigation of synoptic and associated mesoscale patterns leading to significant weather days at McMurdo Station, Antarctica. M.S. Thesis, Naval Postgraduate School, Monterey, California, 114 pp.

Graf, W., Moser, H., Oerter, H., Reinwarth, O., Stichler, W., 1988: Accumulation and ice-core studies on Filchner-Ronne Ice Shelf, Antarctica. *Annals Glaciol.*, **11**, 23–31.

Graf, W., Reinwarth, O., Oerter, H., Dyurgerov, M., 1990: Isotopic and stratigraphical interpretation of a 16m firn core nearby Druzhnaya I. In: *Filchner Ronne Ice Shelf Programme Report No. 4*, Alfred-Wegener-Institute for Polar and Marine Research, Bremerhaven, Federal Republic of Germany, 46–47.

Groisman, P. Y., Koknaeva, V. V., Belokrylova, T. A., Karl, T. R., 1991: Overcoming biases of precipitation measurement: a history of the USSR experience. *Bull. Amer. Meteor. Soc.*, **72**, 1725–1733.

Hollingsworth, A., 1989: The Global Weather Experiment – 10 years on. *Weather*, **44**, 278–285.

Hsu, S. A., 1988: *Coastal Meteorology*. New York: Academic Press, 260 pp.

ITASE, 1992: The International Trans-Antarctic Scientific Expedition: a planning document. Institute for the Study of Earth, Oceans, and Space, University of New Hampshire, Durham, New Hampshire, 18 pp.

Jacobs, S. S., 1992: Is the Antarctic ice sheet growing? *Nature*, **360**, 29–33.

Jenne, R. L., Crutcher, H. L., van Loon, H., Taljaard, J. J., 1971: Climate of the upper air: Southern Hemisphere, Vol. 3, Vector mean geostrophic winds. *Rep. NCAR-TN/STR-58*, National Center for Atmospheric Research, Boulder, Colorado, 61 pp.

Johnsen, S. J., Clausen, H. B., Dansgaard, W., Fuhrer, K., Gundestrup, N., Hammer, C. U., Iversen, P., Jouzel, J., Stauffer, B., Steffensen, J. P., 1992: Irregular glacial interstadials recorded in a new Greenland ice core. *Nature*, **359**, 311–313.

Jouzel, J., Merlivat, L., Petit, J. R., Lorius, C., 1983: Climatic information over the last century deduced from a detailed isotopic record in the south pole snow. *J. Geophys. Res.*, **88**, 2693–2703.

Keen, R. A., 1984: Statistical-dynamical model of accumulation on the Greenland Ice Sheet. *Annals Glaciol.*, **5**, 69–74.

Kuhn, M., 1970: Ice crystals and solar halo displays, Plateau Station, 1967. *International Association for Hydrological Sciences*, Pub. No. 86, 298–303.

Lorius, C., 1983: Accumulation rate measurements on cold polar glaciers. In: Robin, G. de Q. (ed.) *The Climatic Record in Polar Ice Sheets*. New York: Cambridge University Press, pp. 65–70.

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

Mosley-Thompson, E., 1992: Paleoenvironmental conditions in Antarctica since A.D. 1500: ice core evidence. In: Bradley R. S., Jones, P. D. (eds.) *Climate Since A.D. 1500*. New York: Routledge, pp. 572–591.

Overland, J. E., 1984: Scale analysis of marine winds in straights and along mountainous coasts. *Mon. Wea. Rev.*, **112**, 2530–2534.

Paterson, W. S. B., 1981: *The Physics of Glaciers, 2nd edn.* New York: Pergamon Press, 380 pp.

Peel, D. A., 1992: Ice core evidence from the Antarctic Peninsula region. In: Bradley, R. S., Jones. P. D. (eds.) *Climate Since A. D. 1500*. New York: Routledge, pp. 549–571.

Peixoto, J. P., Oort, A. H., 1983: The atmospheric branch of the hydrological cycle and climate. In: Street-Perrott, A., Beran, M., Ratcliffe, R. (eds.) *Variations in the Global Water Budget*. Hingham, Mass.: D. Reidel, pp. 5–65.

Phillpot, H. R., 1991: The derivation of 500 hPa height from automatic weather station surface observations in the Antarctic continental interior. *Aust. Meteor. Mag.*, **39**, 79–86.

Pourchet, M., Pinglot, F., Lorius, C., 1983: Some meteorological applications of radioactive fallout measurements in Antarctic snows. *J. Geophys. Res.*, **88**, 6013–6020.

Raper, S. C. B., Wigley, T. M. L., Mayes, P. R., Jones, P. D., Salinger, M. J., 1984: Variations in surface air temperatures. Part 3: The Antarctic, 1957–82. *Mon. Wea. Rev.*, **112**, 1341–1353.

Rasmusson, E. M., 1977: Hydrological application of atmospheric vapour-flux analyses. *Operational Hydrology Report 11*, World Meteorological Organization, Geneva, 50 pp.

Reeh, N., Clausen, H. B., Gundestrup, N., Johnsen, S. J., Stauffer, B., 1977: $\delta(^{18}O)$ and accumulation rate distribution in the Dye 3 area, south Greenland. *International Association for Hydrological Sciences*, **118**, 177–181.

Robin, G. de Q., 1977: Ice cores and climatic change. *Phil. Trans. R. Soc. Lond.*, **B.280**, 143–168.

Satow, K., Watanabe, O., 1990: Seasonal variation of oxygen isotopic composition of firn cores in the Antarctic Ice Sheet. *Annals Glaciol.*, **14**, 256–260.

Schwerdtfeger, W., 1984: *Weather and Climate of Antarctica, Dev. Atmos. Sci.*, Vol. 15. New York: Elsevier, 261 pp.

Scorer, R. S., 1988: Sunny Greenland. *Quart. J. Roy. Meteor. Soc.*, **114**, 3–29.

Serreze, M. C., Barry, R. G., 1988: Synoptic activity in the Arctic Basin, 1979–85. *J. Climate*, **1**, 1276–1295.

Streten, N. A., Troup, A. J., 1973: A synoptic climatology of satellite observed cloud vortices over the Southern Hemisphere. *Quart. J. Roy. Meteor. Soc.*, **99**, 56–72.

Taljaard, J. J., Van Loon, H., Crutcher, H. L., Jenne, R. L., 1969: Climate of the upper air: Southern Hemisphere, Vol. 1, Temperature, dew points and heights at selected pressure levels. *Rep. NAVAIR 50-1c-55*, Naval Weather Service Command, Washington, D.C., 134 pp.

Trenberth, K. E., Olson, J. G., 1988: Intercomparison of NMC and ECMWF Global Analyses: 1980–1986. *NCAR Technical Note NCAR/TN-301+STR*, National Center for Atmospheric Research, Boulder, Colorado, 81 pp.

Warrick, R. A., Oerlemans, J. R., 1990: Sea level rise. In: Houghton, J. T., Jenkins, G. J., Ephraums, J. J. (eds.) *Climate Change: The IPCC Scientific Assessment*. Cambridge: Cambridge University Press, pp. 257–281.

Weidner, G. A., Stearns, C. R., 1991: A two-year record of the climate on the Greenland crest from an automatic weather station. In: Weller, G., Wilson, C. L., Severin, B. A. B. (eds.) *Proceedings of the International Conference on the Role of the Polar Regions in Global Change, Vol. 1.* June 11–15, 1990, University of Alaska Fairbanks, Fairbanks, Alaska, 220–222.

Wentz, F. J., 1992: Measurement of oceanic wind vectors using satellite microwave radiometers. *IEEE Trans. Geosci. Remote Sensing*, **30**, 960–972.

Whillans, I. M., 1981: Reaction of the accumulation zone portions of glaciers to climatic change. *J. Geophys. Res.*, **86**, 4274–4282.

Whillans, I. M., Bindschadler, R. A., 1988: Mass balance of Ice Stream B, West Antarctica. *Annals Glaciol.*, **11**, 187–193.

Woo, M.-K., Heron, R., Marsh, P., Steer, P., 1983: Comparison of weather station snowfall with winter snow accumulations in high Arctic basins. *Atmosphere-Ocean*, **21**, 312–325.

Zwally, H. J., 1989: Growth of Greenland: interpretation. *Science*, **246**, 1589–1591.

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