JP2.12 ANTARCTIC NET PRECIPITATION ESTIMATE FROM NCEP-DOE REANALYSIS-2

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1. INTRODUCTION

The importance of Antarctic precipitation to global climate change has long been recognized. Douglas (1991) demonstrated that rate of atmospheric precipitation over Antarctica may have important role to global sea level variation via impact on the surface snow and ice accumulation. In particular, Cullather et al (1996) recently found that there is relationship between net precipitation over West Antarctica and the El Nino-Southern Oscillation (ENSO) phenomenon based on the European Center for Medium-range Weather Forecasts (ECMWF) operational analyses. Generally, ENSO is recognized as the most prominent signal of global climate variability. However, due to large errors caused by small rainfall amounts and drifting snow, direct precipitation observation over the Antarctica is very difficult (Bromwich 1988). In general, three kinds of datasets are used to derive the Antarctic precipitation: radiosonde observations of wind and moisture profiles (e.g.. Peixoto and Oort 1983; Bromwich 1988), analysis/reanalysis data (e.g., Bromwich et al. 1995; Cullather et al. 1996), and satellite data (Slonaker and Van Woert 1999; Zou and Woert 2001; Francis et al. 2002a and b). Although the Radiosonde data is much more reliable, the poor spatial observational network for large areas of Antarctic continent has hindered our understanding of the distribution and variability of precipitation. To avoid the impact of gap from radiosonde data, Slonaker and Woert (1999), Zou and Woert (2001, 2002) applied the Television and Infrared Observational satellite (TIROS) Operational Vertical Sounder (TOVS) data to estimate Antarctic Precipitation. In this method, atmospheric winds are derived from TOVS temperature sounding via the thermal wind relationship. Then, moisture flux and net precipitation are estimated from derived wind fields and TOVS moisture observations. However, net precipitation derived from TOVS Path A is found so small compared with surface observations and numerical analysis/reanalysis (Zou and Van Woert 2001; Zou et al. 2004). Perhaps numerical analysis/reanalysis data is attractive to compute the Antarctic atmospheric precipitation, especially for the discussion of the spatial distribution and temporal variability of precipitation. First, model inputs have contained various observations from satellite, ships, buoys, aircraft, and radiosonde network etc. Second, analyzed/reanalyzed data have been constrained to be consistent via comprehensive physical processes. In addition, these data are easy to be used with regular, gridded formats and no gaps in space and time.

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In this paper, the new improved NCEP reanalysis data called NCEP-Department of Energy (NCEP-DOE) Reanalysis 2 (R2) will be used. The R2 data has corrected many of the known problems found in the earlier NCEP-NCAR reanalysis (Kanamitsu et al. 2002). Especially, two problems related to the Southern Hemisphere have been corrected. One is PAOBS (Southern Hemisphere bogus data), which are shifted by 180° longitude for the period of 1979-94 during the NCEP-NCAR Reanalysis data assimilation. Second, perhaps more importantly, errors associated with polar moisture fields have been investigated. A simplification in the moisture diffusion parameterization used by the assimilation model leads to a spurious wave pattern in the polar region (Kanamitsu et al. 2002). In addition, the performance of R2 data in the southern high-latitude oceans has been validated by Zou et al. (2004). The results showed that the spatial distribution of the annual mean R2 moisture statistics has a good agreement with Special Sensor Microwave/Imager (SSM/I) data. In this study, we focus on the R2 performance of net precipitation over the Antarctica. Atmospheric net precipitation is derived using wind and moisture field via atmospheric moisture budget method.

2. DATA

The R2 data used in this study were obtained from the Climate Diagnostics Center (CDC). The data are archived four times daily with a global horizontal resolution of 2.5 latitude-longitude, 28 vertical sigma levels from January 1979 though December 2003. In the practical calculation of moisture transport and convergence, based on the vertical distribution features of atmospheric moisture, only data under the 100hPa level are selected. That is, only the moisture, wind and temperature fields, defined on the 12 levels:1000, 925, 850, 700, 600, 500, 400, 300, 250, 200,150 and 100 hPa, are used.

3. NET PRECIPITATION CALCULATION

Atmospheric net precipitation can be estimated using the moisture budget equation. The basic equation in an isobaric coordinate system is written as

$$P - E = -\frac{\partial W}{\partial t} - \nabla . \frac{1}{g}\int_{p_0}^{p_s} qVdP \qquad (1)$$

where P represents precipitation, E surface evaporation/sublimation, W the precipitable water of a whole atmospheric column, q is the specific humidity, V the horizontal wind vector, p_s the surface pressure, and g the gravity constant. $\partial w / \partial t$ is the moisture storage term, $\nabla \bullet$ term is moisture convergence.

To solve the equation above, the moisture flux is first computed at each pressure level as the product of specific humidity and horizontal wind vector. To calculate the time-averaged vapor flux, moisture transport are divided into mean and eddy components such that

$$\overline{qV} = \overline{q}\overline{V} + \overline{q'V'} \qquad (2)$$

the over-bar denotes temporal averaging, and the superscript prime denotes the eddy component. Generally, the moisture storage term $\partial w / \partial t$ may be ignored because it is much smaller than the moisture convergence for seasonal time scales (Peixoto and Oort 1992). Thus, the net precipitation is only determined by the convergence of the moisture fluxes.

The spatial averaged net precipitation can be written as:

$$< P - E >= -\frac{1}{S}\oint[(\int_{p_0}^{p_s}\overline{q}\overline{V}dP) + \oint(\int_{p_0}^{p_s}\overline{q'V'}dP)].ndl \qquad (3)$$

where dl represents the line increment along the Antarctic border, n represents the outward normal to the area perimeter, $<>$ represents area average, and S is area.

4. WIND CORRECTION METHOD

Keeping atmospheric mass conservation is a fundamental requirement in the atmospheric model. The failure to satisfy the dry-air mass budget will lead to inaccurate calculation on any other atmospheric budget being calculated (Trenberth at al. 1994). However, in practical calculation, the conservation of dry air mass is not satisfied because of lots of reasons. For example, the imbalances arise from post-processing variables onto pressure surfaces, problems of dealing with the lower boundary and substituting an artificial atmosphere below ground, and diurnal pressure tendencies associated with the semidiurnal tide and the timing and distribution of observations. Any errors in the budget of dry air will therefore be reflected strongly in the budgets of all these other quantities. For instance, typical errors in the vertically integrated heat budget due to mass imbalances, easily exceed 100 Wm^{-2} locally (Alexander and Schuert 1990). This dry-air mass imbalance is especially larger in Antarctica because of the complex topography there. In the Antarctic precipitation calculation, Zou and Van Woert (2001) indicated that the introduction of the conservation of mass reduces the estimates of the moisture flux and net precipitation dramatically in comparison with the nonmass-conserved method. For instance, the estimates of the zonally averaged, vertically integrated moisture flux across 50S are reduced by 56% and the net precipitation between 50S and 60S latitude belt are reduced by 63%. Accordingly, it is not only important to recognize errors in the mass budget but also to correct them in some way (Boer and Sargent 1985; Alexander and Schubert 1990). To keep atmospheric mass conservation and then get accurate net precipitation estimation, this study solves the residual equations for the mass balance to compute the wind correction term.

Assuming m_d represents dry air mass, the conservation of mass integrated in the vertical is:

$$\frac{\partial m_d}{\partial t} + \nabla \bullet \frac{1}{g}\int_0^{p_s} m_d V dp = 0 \qquad (4)$$

Dry air mass m_d equals also the difference of the total atmosphere mass m minus the precipitable water w

$$m = \frac{1}{g}\int_0^{p_s} dp = \frac{p_s}{g} \qquad (5)$$

$$w = \frac{1}{g}\int qdp \qquad (6)$$

accordingly,

$$m_d = m - w = \frac{p_s}{g} - w \qquad (7)$$

Thus, equation 4 is rewritten using equations 5 and 6,

$$\left(\frac{\partial p_s}{\partial t} - g\frac{\partial w}{\partial t}\right) + \nabla.\int_0^{p_s}(1-q)V\,dP = 0 \qquad (8)$$

To evaluate the conservation of dry air mass, the residual of equation 8 is represented as R,

$$R = \left(\frac{\partial p_s}{\partial t} - g\frac{\partial w}{\partial t}\right) + \nabla.\int_{p_0}^{p_s}(1-q)V^*\,dP \qquad (9)$$

where V^* is the original velocity field. Here, it is the velocity field from R2.

In 1991, Trenberth developed a method to solve equation 9. Here following Trenberth method, but give a more detailed description and further discussion for the limitations of this method. In this method, key assumption is that the largest uncertainty and greatest source of errors in the global analyses are associated with the divergent wind component (Boer and Sargent 1985). Thus, the strategy for evaluating and correcting the overall mass balance is to assume that any errors are in the divergent wind.

Following this assumption, original velocity equals the difference of real velocity and wind correction. That is,

$$V^* = V - V^c \qquad (10)$$

putting equation 10 into equation 9, then get:

$$R = -\nabla \bullet \int_{p_0}^{p_s}(1-q)V^c\,dp \qquad (11)$$

To simply solve procedure, wind correction V^c is assumed to be barotropic, that means V^c is same at every level, not a function of pressure. Thus,

$$\int_{p_0}^{p_s}(1-q)V^c\,dp = V^c(p_s - p_0 - gw) \qquad (12)$$

$$R = \nabla \bullet [V^c(p_s - p_t - gw)] \qquad (13)$$

Further assume that there is a potential function A, which is defined as,

$$R = \nabla \bullet \nabla A \;= \nabla^2 A \qquad (14)$$

Compared the right sides of equation 13 and 14, wind correction can be expressed as,

$$V^c = \nabla A \;/(p_s - p_t - gw) \qquad (15)$$

when potential function A is solved, V^c is obtained. In fact, equation 14 is a Poisson's equation. In this study, it is solved by a relaxation method.

5. RESULTS

5.1 Evaluation of wind correction

Figure 1a shows the spatial distribution of annual mean zonal velocity correction for 1988. Seeing from this figure, an interesting spatial distribution pattern can be found. The whole southern high latitude area is clearly separated into two parts: positive and negative wind correction areas. According to clockwise rotation

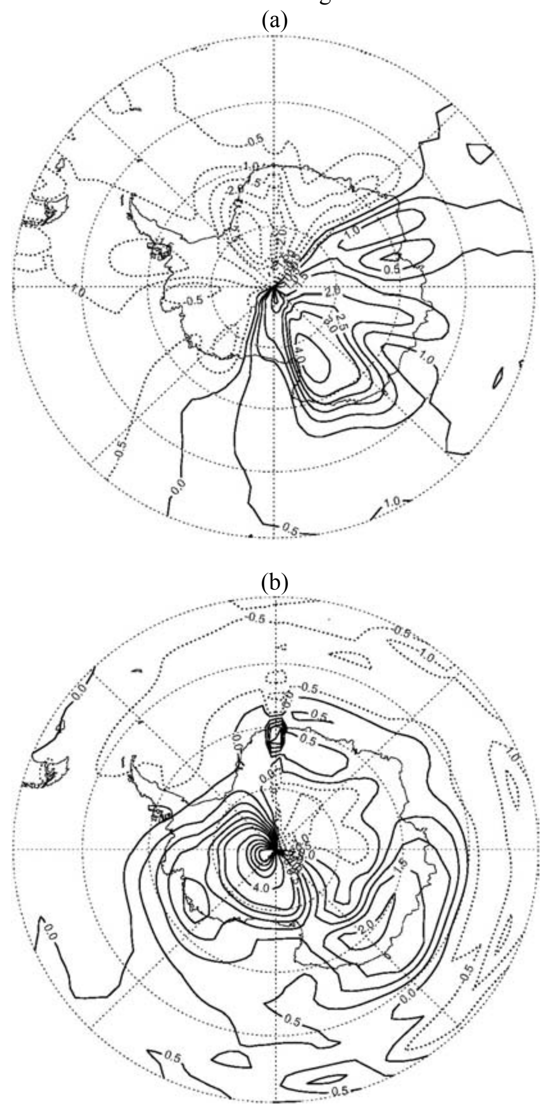


Figure 1 Annual mean R2 wind correction for 1988: (a) Zonal wind; (b) Meridional wind. Unit: 0.1m/s

rotation, the positive zonal wind correction area ranges from 45°E to 225°E. In general, magnitudes of wind correction are less than 0.5 m/s. The largest area, in which values are larger than 0.4 m/s, is located at Wilkes Land. In contrast, from 225°E to 45°E, zonal wind correction is characterized by negative values. The maximum negative corrections (absolute value) are found in the land area between 0 and 45°W. This kind of spatial distribution characteristic of wind correction implies that wind field from R2 model has a systematical error.

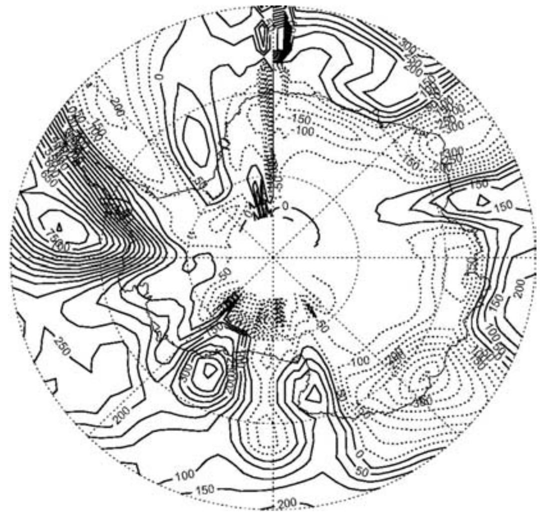
The spatial distribution of meridional wind correction is shown in Figure 2. Seeing from the figure, the positive wind correction can be seen at almost whole Antarctic continent, except a small area closed to South Pole in Eastern Antarctica. This means that real meridional wind velocity is decreased by R2 model. The maximum positive correction values are found in Marie Byrd Land and Wilkes Land. Generally, the magnitudes of wind corrections are also less than 0.5 mm/s. However, over the Southern Ocean, most parts show the negative wind correction features. That is, real meridional winds are enlarged by the reanalysis procedures.

5.2 Atmospheric moisture convergence and net precipitation

Figure 2 demonstrates the Antarctic moisture convergence calculated from R2 dataset for 1988. In general, the annual mean moisture convergence (Figure 2a) has similar spatial distribution pattern with previous studies (Bromwich et al. 1995; Cullather et al. 1996). Such as, large moisture convergence areas are along the coast of Antarctic continent and over the South Ocean and spatial distribution shows a atmospheric wave pattern. The largest convergence center is over the Bellinghausen Sea, which are locally greater than 800 mm/year. In fact, these largest moisture convergence centers are associated with atmospheric cyclone activity areas.

Figure 2b indicates the spatial distribution of eddy component of moisture convergence. Two main characteristics over the Antarctica can be seen. One is that magnitude of moisture convergence varies with orographic height. Moreover, moisture convergence shows a symmetric pattern along the latitude cycle. A gradient of increasing towards to north from the South Pole can be seen. The cycle with minimum value, which is less than 50 mm, is right over the highest terrain area of the Eastern Antarctica. The larger eddy convergence with about 200mm value is along the coast of Antarctica. The maximum convergence occurs over the southern oceans. Four maximum convergence centers, which have the values of 400 mm or larger, are respectively located at South Atlantic Ocean, South Indian Ocean and South Pacific Ocean.

(a)



(b)

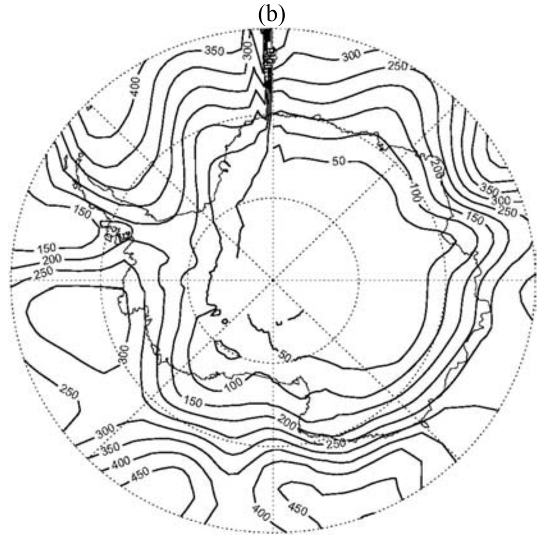


Figure 2. Annual mean moisture convergence from R2 for 1988: (a) Mean component; (b) Eddy component. Unit: mm/year

The accumulated annual net precipitation is calculated from moisture convergence fields above. To assess the performance of R2 data, the net precipitation derived from R2 is compared with those from different studies. A comparison of net precipitation from different studies is listed in Table 1. The calculated annual accumulated precipitation from different data and methods show a big difference. The minimum value is only 46 mm/year. However, the maximum value reaches 166 mm/year. This big difference partly implies the difficulty of estimating Antarctic precipitation. Generally, the multiyear glaciological data is recognized as more accurate. Bromwich (1990) gave a value of 151-156 mm/year using a multiyear glaciological dataset. Giovinetto and Bentley (1985) got an estimate of 143 mm/year. Net precipitation calculated from ECMWF analysis (Bromwich et al. 1995) is about 151-157 mm/year, which tends to agree with the glaciological estimate. However, analyses from U.S. National Meteorological Center and Australia Bureau of Meteorology (ABM) only have estimates of 98-108 mm/year and 84-89 mm/year, respectively. In this study, net precipitation is calculated using R2 atmospheric moisture and

Table 1. Comparison of the Antarctic net precipitation estimates from different studies (Unit: mm/year)

Studies	Data source	Year	Precipitation
Giovinetto and Bentley (1985)	Glaciological accumulation	1960-85	143
Bromwich (1990)	Glaciological P-E	Multiyear	151-156
Vaughan et al. (1990)	In situ surface observation	Multiyear	166
Cullather et al. (1998)	ECMWF analysis P-E	1985-1995	151
Bromwich et al. (1995)	ECMWF analysis P-E	1985-1992	157
	NMC analysis P-E	1985-1992	108
	ABM analysis P-E	1985-1992	89
Bromwich et al. (1995)	ECMWF analysis P-E	1988	148
	NMC analysis P-E	1988	98
	ABM analysis P-E	1988	84
Zou et al (2004)	R2 P-E, equivalent Antarctic area of 70°S	1988	137
	TOVS/R2 P-E, equivalent Antarctic area of 70°S	1988	46
	ECMWF, equivalent Antarctic area of 70°S	1988	139
This study	R2 P-E, equivalent Antarctic area of 70°S	1988	152

corrected wind fields. The result is quite encouraging. The annual accumulated Antarctic net precipitation is 152 mm, which greatly improves estimate from NCEP analysis and agrees with estimates from multiyear glaciological data and ECMWF analysis/reanalysis.

6. CONCLUSION

The newly released NCEP-DOE Reanalysis 2 data has been used to estimate net precipitation over the Antarctica. The results demonstrate that R2 data provides generally good estimates for atmospheric precipitation. However, it is worth to note that, due to some post-processing procedures, the dry air mass conservation in R2 dataset is destroyed. To get a better estimate for net precipitation via atmospheric method, wind correction is necessary.

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