Instrument- and Tree-Ring-Based Estimates of the Antarctic Oscillation

JULIE M. JONES AND MARTIN WIDMANN
Institute for Coastal Research, GKSS Research Centre, Geesthacht, Germany

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ABSTRACT

An estimate of the strength of the austral summer Antarctic Oscillation using station sea level pressure records for the period 1878–2000 is presented, the first to the authors’ knowledge. The reconstruction was obtained by relating the Antarctic Oscillation (AAO) intensity derived from NCEP–NCAR reanalysis data to the leading principal components of station records using multiple regression analysis. Particular effort has been made to fit the model in a way that is robust to the questionable trends in the NCEP–NCAR data in the Southern Hemisphere. The trends in the reconstruction are derived from the station data, not from the NCEP data. Cross-validation with the NCEP data and comparison with other analyses of the AAO over the late instrumental period give confidence that this station-based reconstruction can be regarded as trustworthy. With regard to the whole reconstruction period, some unquantifiable uncertainty stems from potential instability of the statistical relationships.

To extend this record further back, a reconstruction using tree-ring chronologies back to 1743 has also been undertaken. Comparison with the station-based reconstruction shows moderate agreement on interannual and decadal timescales, but the comparison also points toward the inherent uncertainties of proxy-based climate reconstructions. In particular, it was found that this tree-based reconstruction may have been influenced by a warming that is not related to changes in the Antarctic Oscillation index during the twentieth century. Comparison of the tree-based reconstruction with a published reconstruction of zonal flow over New Zealand before the twentieth century shows common features.

The temperature and precipitation signals of the Antarctic Oscillation have been calculated and show that the response of the chronologies to Antarctic Oscillation variability is physically plausible. In addition, it was shown that a substantial fraction of the observed warming over much of Antarctica between the late 1950s and the 1980s can be linked to changes in the Antarctic Oscillation, whereas the observed warming over New Zealand is related to other influences.

1. Introduction

Detection of climatic changes during the last century, and their potential attribution to increasing concentrations of atmospheric greenhouse gases and other anthropogenic activities, requires a realistic estimation of the level of natural climate variability on decadal and longer timescales. As the observational record goes back mostly only to the mid-nineteenth century, other sources of information on past climate variability are required. One source of such information are palaeoclimatic proxy data derived from natural archives, such as tree rings and ice cores. These have previously been used both to reconstruct the local climate, and larger-scale indices or patterns of atmospheric circulation. Reconstructions of past climate variability from proxy data include, for example, estimates of global and hemispheric (Mann et al. 1998, 1999; Crowley and Lowery 2000; Briffa 2000) and regional temperatures (Briffa et al. 2001), and of the North Atlantic Oscillation (NAO) from tree rings (Cook et al. 2002), ice cores (Appenzeller et al. 1998), and multiproxy data (Glueck and Stockton 2001; Cullen et al. 2001). Such reconstructions complement the use of transient numerical climate models forced with past solar and other forcings for estimating climate variability in the preinstrumental period.

This paper uses long observational sea level pressure (SLP) records and tree-ring width chronologies to produce reconstructions of the dominant mode of Southern Hemisphere extratropical circulation, the Antarctic Oscillation (AAO). The AAO, which has also been termed the “Southern Annular Mode” (Thompson and Wallace 2000), is a zonally symmetric mode representing exchange of mass between the midlatitudes near 45°S and high latitudes poleward of 60°S, and characterizes fluctuations in the strength of the circumpolar vortex. This mode has been found to be present at various atmospheric levels (von Storch 1999), for example, SLP (Gong and Wang 1999; Rogers and van Loon 1982), 500 hPa (Rogers and van Loon 1982; Mo and White 1985), and 850 hPa (Thompson and Wallace 2000).
The Southern Hemisphere has large areas of ocean, and thus few land surfaces from which to obtain proxy and observational data. This lack of data produces problems in finding sufficient data to produce and evaluate reconstructions. Nevertheless, a number of aspects of the Southern Hemisphere circulation have been reconstructed using proxy data. The Trans-Polar index (TPI), which Pittock (1980) suggests is a measure of the displacement of the Southern Hemisphere vortex (the strength of which is described by the Antarctic Oscillation Index (AAOI)) towards either South America or New Zealand, has been reconstructed by Villalba et al. (1997) using high-latitude tree-ring chronologies (the TPI has been confused with the AAOI, Folland et al. 2001). Other dendroclimatic reconstructions include SLP variability in the Tasman Sea area (D’Arrigo et al. 2000), the latitude of the east Pacific anticyclone (Villalba 2000), and the Southern Oscillation index (Stahle et al. 1998). Despite interest in the AAO, as far as we know, no reconstruction has been published.

We present a reconstruction of the austral summer [November–December–January (NDJ)] AAOI, using multiple linear regression between the leading principal components (PCs) of standardized station SLP data and the AAOI derived from National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data, termed the station-based reconstruction (SBR). We also present a second reconstruction using the same method and tree-ring chronologies from Argentina and New Zealand, the tree-based reconstruction (TBR). Since the SBR is based on direct pressure measurements, it can be expected to have substantial skill. However it is based predominantly on data from one center of action (New Zealand and Australia), which could be problematic for reconstructing the intensity of a hemispheric scale pattern. Validation against independent data for the last few decades indicates that the reconstruction is successful. The quality of the SBR for the whole reconstruction depends on whether the statistical relationships derived during the calibration period hold true for the rest of the reconstruction period. Because the SBR is based on data predominantly from one center of action, the relationships may be less stable than if it were based on information from several centers. This adds additional, but unquantifiable (outside the calibration period), uncertainty to it. Nonetheless we assume that the SBR is more reliable than the TBR, which is based on indirect relationships between tree growth and the AAOI. Thus comparison of the TBR with the SBR allows a systematic assessment of the uncertainties in the TBR. Given the uncertainties in circulation reconstructions from proxy data found by our analysis, and also by Schmutz et al. (2000), for example, we regard the TBR as preliminary.

2. Data and methods
   a. NCEP–NCAR data

   The SLP data that were used to define the AAO were taken from the NCEP–NCAR reanalysis (Kalnay et al. 1996; Kistler et al. 2001), regridded to a 5° × 5° latitude–longitude grid using a Gaussian space interpolation technique (Santer 1988). The reanalysis involves assimilation of surface, satellite, radiosonde, and other observations of variables including pressure, temperature, and humidity using a global forecast model. Several errors have been identified in this dataset in the Southern Hemisphere. An error of 180° longitude was found in the location of the Australian surface pressure bogus data for the Southern Hemisphere for the period 1979–92. These data enter the NCEP–NCAR reanalysis and are estimates of the SLP produced by Australian analysts using satellite data, conventional data, and time continuity for the data-poor Southern Ocean. This problem is considered not to be significant on climatological timescales (Kistler et al. 2001), and is therefore not relevant to this work.

   The second, potentially more serious error, are the possible spurious trends identified by Hines et al. (2000) in the NCEP–NCAR Southern Hemisphere SLP for the region south of 45°S, with the largest trend near 65°S. We have therefore designed our analysis in a way that is robust against the influence of the 65°S region and the trends in the NCEP data. First, the southern limit of the analysis domain was chosen to be 60°S, to exclude the region of strongest trends identified to be at 65°S by Hines et al. (2000). Second, the data were linearly detrended at each grid point before analysis in order to remove the influence of trends on the definition of the AAO pattern and its statistical relation to station or tree records. The northern limit of 20°S was chosen to be consistent with the annular modes definitions of Thompson and Wallace (2000). We consider here the NDJ AAO, but the AAO is present year-round in the troposphere (Thompson and Wallace 2000; Gong and Wang 1998; Thompson and Solomon 2002) and trends in recent decades agree in all except one month (Thompson et al. 2000).

b. Station data

   Southern Hemisphere station SLP records were obtained from P. Jones, Climatic Research Unit, United Kingdom (Jones et al. 1999; Jones 1991). These data originate mostly from the World Weather Records and some earlier data from yearbooks from the U.K. National Meteorological Library at Bracknell, and have been homogenized according to Jones (1987). A total of 28 stations, those with data since at least 1878, were used. To select those containing an AAO signal, the stations were correlated with the AAOI (see section 2d for the exact definition). Stations were retained that were significantly correlated at the 5% level, a total of 10 (Table 1). This was done for the period 1948–85 (the common period of trees and NCEP data) so that the tree and station-based reconstructions are fitted on the same data. All of the rejected stations except one (Chatham) were located outside of AAO centers of action.
There are a number of possible reasons why so few chronologies show a relationship with the NDJ AAOI. First, chronology growth may be influenced by climate in other seasons. It will also be shown in section 3c that there are also large geographical areas that do not have significant temperature and/or precipitation signals from the AAO. For example, much of South America does not have a temperature signal, thus temperature-sensitive chronologies located in these regions would not be expected to show a relationship with the AAOI. Additionally, even in areas with strong correlations between temperature or precipitation and the AAOI on the grid-box scale (e.g., New Zealand), the relationship of local temperature and precipitation with the AAOI may differ greatly within a grid box, according to local meteorology.

The original chronologies have been standardized to account for the age–growth effect, that is, as a tree ages the growth is dispersed around an increasing radius. The standardization will have been done in different ways for many of the chronologies. To ensure consistency between chronologies, the raw ring-width chronologies for the selected trees were therefore restandardized [by C. Woodhouse, NOAA, Boulder, CO], using a cubic spline two-thirds of the length of the measurement series used, which is regarded as relatively conservative (C. Woodhouse 2001, personal communication). This standardization places a limit to the amount of low-frequency variability that can be retained in tree-ring chronologies that is also dependent on the length of the segments making up the series (Cook et al. 1995).

Tree-ring width chronologies may contain autocorrelation due to processes believed to be unrelated to climate, for example, stand dynamics and physiological effects (Cook et al. 1999). Thus autoregressive modeling has been used to remove autocorrelation from the standardized series, to produce so-called whitened, or residual, chronologies. The level of model fit was determined using the Akaike information criterion (Akaiké 1974), and ranges from AR 1 to AR 6 (Table 2). A potential problem with this approach is that climate-induced low-frequency variability might also be removed. It will be shown in section 3b that this is not the case.

### Table 2. Chronologies that have a statistically significant relationship with the NDJ NCEP AAOI (5% level).

<table>
<thead>
<tr>
<th>Chronology</th>
<th>Species</th>
<th>Lat, lon (°)</th>
<th>Autoregressive model level</th>
<th>Time period</th>
</tr>
</thead>
<tbody>
<tr>
<td>1) Estancia Carmen Camino</td>
<td><em>Nothofagus pumilio</em></td>
<td>−54.26, −67.55</td>
<td>4</td>
<td>1730–1986</td>
</tr>
<tr>
<td>2) Estancia san Justo</td>
<td><em>Nothofagus pumilio</em></td>
<td>−54.03, −68.34</td>
<td>4</td>
<td>1729–1985</td>
</tr>
<tr>
<td>3) Lago Yehuin</td>
<td><em>Nothofagus pumilio</em></td>
<td>−54.28, −67.43</td>
<td>4</td>
<td>1735–1986</td>
</tr>
<tr>
<td>4) Estancia Maria Cristina-Bosque Virgen</td>
<td><em>Nothofagus pumilio</em></td>
<td>−54.30, −67.05</td>
<td>4</td>
<td>1743–1986</td>
</tr>
<tr>
<td>5) Monte Grande, Magallanes</td>
<td><em>Nothofagus pumilio</em></td>
<td>−53.12, −72.10</td>
<td>2</td>
<td>1681–1988</td>
</tr>
<tr>
<td>6) Peninsula Brunswick</td>
<td><em>Nothofagus pumilio</em></td>
<td>−53.30, −71.10</td>
<td>6</td>
<td>1666–1988</td>
</tr>
<tr>
<td>7) Norquinco</td>
<td><em>Austrocedrus chilensis</em></td>
<td>−39.07, −71.07</td>
<td>2</td>
<td>1567–1989</td>
</tr>
<tr>
<td>8) Moa Park</td>
<td><em>Libocedrus bidwillii</em></td>
<td>−40.56, 172.56</td>
<td>1</td>
<td>1491–1991</td>
</tr>
<tr>
<td>9) Putara</td>
<td><em>Halocarpus biformis</em></td>
<td>−40.40, 175.31</td>
<td>1</td>
<td>1646–1993</td>
</tr>
</tbody>
</table>
d. Method

Multiple linear regression (e.g., von Storch and Zwiers 1999, Wilks 1995) was applied to estimate the AAOI from the leading PCs of normalized tree-ring records or station observations. This so-called principal component regression (PCR) was first used for climate reconstructions by Briffa et al. (1986) to estimate the leading PCs of summer SLP over England from tree-ring PCs, and has also been used to estimate drought indices from tree rings (Cook et al. 1999); and to estimate the winter North Atlantic Oscillation index from multiproxy data (Cook et al. 2002) and from early instrumental and documentary data (Luterbacher et al. 2002).

For the purpose of model fitting we define the AAOI as the first PC of detrended NCEP NDJ SLP for the domain 20°–60°S and the AAO pattern as the first empirical orthogonal function (EOF) of these data. The detrending of the NCEP data ensures that neither the structure of our AAO pattern nor the statistical relation of its amplitude to the predictor variables is contaminated by potential unrealistic trends in the NCEP data. Note that this AAOI, which has by definition no trend for the fitting period 1948–85 and therefore most likely differs from the true, unknown AAOI (defined as the projection of perfect SLP data onto the AAO pattern), is only used for model fitting. The trends in the reconstructions are weighted means of the trends in the chronologies or in the station data, which are influenced by the true AAOI trends, and are therefore not affected by trends in the NCEP data.

An overall scaling factor for the AAO pattern and index can be arbitrarily chosen. Our normalization is such that the variance of the AAO index obtained from detrended NCEP seasonal means equals one for the period 1948–85. The physical units are associated with the AAO pattern. Local SLP signals associated with a given AAOI can be obtained by multiplying the local AAO pattern loading with the AAOI. The zero line is defined such that the AAOI obtained from detrended NCEP seasonal means has zero mean over the period 1948–85.

The predictor PCs in our analysis are derived from NDJ seasonal means of long station records or from annually resolved tree-ring width chronologies. Again, all records were detrended prior to analysis in order to be consistent with the detrended predictand data. The station and tree-ring data were normalized to unit variance so that records with large variance do not dominate the statistical model (i.e., the EOFs and PCs were derived from the correlation matrix). In order to avoid overfitting only the leading PCs were used for the PCR model. The optimal number of PCs to be retained was determined based on the correlation between the reconstructed and the true AAOI in an independent validation period. Three PCs were retained for the stations and four for the trees. All EOFs and PCs were calculated with respect to the period 1948–85, which is when the tree-ring and NCEP data overlap.

To obtain the reconstruction, the station or tree-ring width records were normalized by dividing by the standard deviation from the fitting period, and the PCs were calculated by projecting the normalized values on the EOFs defined in the fitting period. Multiplication of these PCs with the PCR weights derived from the fitting data yields the reconstruction. The reconstruction can equivalently be expressed as a sum of weighted normalized station or tree records. These weights for the stations or trees are shown in the respective plots (Figs. 1 and 5). They were obtained by multiplying the tree or station EOF loadings by the PC weights.

As there is a relatively short overlap period of only 38 years (37 summers) (1948–85) between the tree-ring records and the NCEP data, the model fitting was performed with respect to interannual variability. Calibration based on lower-frequency variability, although desirable, would not be robust due to the small number of independent time steps. The degree to which our model can estimate variability on longer timescales depends on the degree to which the statistical link between the AAO and proxy data is not strongly timescale-dependent. With respect to the TBR there is also the question of how the preprocessing of the tree-ring data affects the reconstruction. This issue will be discussed in section 3b.

For independent validation of the reconstruction on annual and on longer timescales we did not want to withhold more than a few years in order to keep the fitting period sufficiently long. Instead we use a cross-validation procedure (Michaelson 1987; Wilks 1995; Glueck and Stockton 2001) during which the available data are repeatedly divided into calibration and short validation datasets. We performed the PCR 37 times, each time estimating a different year not included in the fitting data. These 37 individual years were then concatenated to produce a validation record. Using all data except the validation time step for model fitting has been found to be problematic as, due to autocorrelation, the withheld data may not be independent from the data used to estimate the model (Michaelson 1987). However, Michaelson (1987) found lag 1 autocorrelations of less than 0.25 to probably not have any measurable effect on estimates of forecast skill. Because the lag 1 autocorrelation of the detrended NCEP AAOI is 0.14, of the detrended SBR 0.01, and of the detrended TBR -0.05, we do not expect a substantial overestimation of model skill. Nevertheless, to be safe, we left out two years on either side of the validation year so that the time step that is reconstructed is practically uncorrelated with all time steps used for model fitting. The final model used for producing the reconstructions was fitted using data for all 37 years.

The reliability of the reconstructions depends on the stability of the statistical relationship between local climate and the AAOI throughout the entire reconstruction period.
period, this skill cannot be directly quantified. Prior to the period that the relationships can be directly validated, instabilities can be caused by either changes in the response of the local data to the local climate or by changes in the relationship between the local climate and the large-scale patterns that are reconstructed (Cook et al. 2002). The first source of instability is of concern only for the TBR, since the response of tree growth to local climate may have changed, but not for the SBR, if we assume reasonably accurate measurements of the seasonal SLP mean for the whole record. The second source of instability applies to both reconstructions. Further discussion of instabilities is provided in section 3.

3. Results

a. The station-based reconstruction

The AAO pattern and the station weights used to obtain the reconstruction are shown in Fig. 1. The AAO pattern (detrended SLP EOF1) explains 28% of the variance of the detrended data and is characterized by opposite signs at mid (approximately 45°S) and high latitudes. In the positive phase of the AAO, the westerly flow at high latitudes is strengthened and that at mid latitudes weakened, and in the negative phase this is reversed.

The regression weights are positive at all stations except for Perth Airport and Ushuaia. This reflects the fact that all stations but these two are located within areas of positive loadings of the AAO pattern. The large common weight of the Australasian stations may be expected because they are located in a center of action of the AAOI. Note that this large common weight is not a result of the large number of stations in this region. The weights are dimensionless because the input series to the PCA filtering prior to the multiple regression are normalized and therefore dimensionless.

The coefficient of multiple determination during the fitting period is 0.92 (Fig. 2), thus 85% of the variance of the NDJ AAOI can be explained by the PCR model. The correlation of the reconstructed AAOI and the detrended NCEP AAOI in the validation period is 0.91 (statistically significant at the 1% level). The reduction of error (RE) in the validation period is 0.82, which also indicates good prediction skill. The RE compares the
residuals from our reconstruction (validation series) with the residuals relative to an estimate based on no knowledge, which here is taken as the calibration period mean of the predictand (Fritts et al. 1990; Cook et al. 1999). The RE can have values from $\infty$ to $+1$, with positive values indicating skill relative to the no-knowledge estimate, and a value of $+1$ indicating perfect reconstruction for the validation period. Some overoptimistic comparison may be expected because most of the predictor stations have also entered the NCEP reanalysis, but this effect is probably small because numerous other stations, including those from other centers of action such as the Antarctic, also enter the reanalysis. Although a very good reconstruction skill could be expected as direct pressure measurements have been used, it is noteworthy that the AAOI can be reconstructed so well during the calibration period from stations dominantly located in New Zealand and Australia.

Station data for the period after 1985 are completely independent, as they were not used for fitting the model, thus the SBR for this period can also be compared on an interannual basis with the NCEP AAOI for model evaluation. Figure 3 shows the SBR over this period (and also the TBR) compared to the NCEP AAOI, which was calculated by projecting the NCEP SLP data onto the 1948–85 EOF. This PC agrees well with the SBR over the period 1986–98, with a correlation of 0.69 (statistically significant at the 1% level) and an RE of 0.54.

The SBR is shown in Fig. 4 (top). The black line is a 9-yr running mean. The error bars are confidence intervals, which cover the true value with a probability of 95%. They are defined as $\pm 1.96$ standard deviations of the residuals from the model fitting because during the fitting period and for normally distributed residuals about 95% of the true AAOI values lay inside an interval of this size. We assume that the uncertainty remains constant back in time and that the link between predictor and predictand derived during the calibration period remain constant throughout the reconstruction. However, there is the possibility that the relationships derived during calibration change through time. Cook et al. (2002) suggest, for example, that an anthropogenic signal may have been superimposed on the NAO or Arctic Oscillation (AO) signal during the twentieth century, thus making model fitting problematic when only twentieth century data are used. Because the SBR is based predominantly on information from one center of action of the AAOI, it is possibly more at risk from instability in the relationship between the predictors and the AAOI than if it were based on data from several centers, thus is subject to more uncertainty than suggested by the validation statistics and the confidence intervals. Inclusion of stations from additional centers of action may reduce the uncertainty. Station measurements from the other main AAO centre of action (the Antarctic) start first in the early twentieth century.

It can be seen from Fig. 4 that the SBR shows a period of dominantly negative AAOI from around 1900 up to the late 1950s, then a period of more positive values during the 1960s, followed by a trend to negative values until 1980, and then again to positive values until the end of the record.

b. The tree-based reconstruction

The AAO pattern (the same as in Fig. 1) and the tree-ring weights used to obtain the TBR are shown in Fig. 5. The total influence of the New Zealand chronologies is similar to that of the South American chronologies, thus in contrast to the SBR, the TBR is based on information from two centers of action of the AAO. The coefficient of multiple determination during the fitting period is 0.72 (significant at the 1% level) (Fig. 6), thus 52% of the variance of the AAOI is explained by the regression model. The correlation between the reconstructed AAOI and the detrended NCEP AAOI in the validation period is 0.66. The RE of 0.46 indicates moderate skill relative to a no-knowledge prediction, but shows that the TBR has lower skill than the SBR.

To determine whether the tree growth/climate response implied by the regression relationships are consistent with the local climate signal of the AAO, and the local tree response for the chronologies where this is known, we calculated correlation maps between the AAOI and detrended temperature or precipitation, using NCEP 850-hPa temperatures and the precipitation dataset of New et al. (1999, 2000), respectively. The temperature map shows coherent large-scale features (Fig. 7) with positive correlations in the regions of positive AAO loadings in the mid/southern ocean basins, including over New Zealand, and a weakly positive signal over South America. The surface precipitation signal of

![Graph](image-url)
the AAOI (Fig. 8) also shows large areas with a relatively coherent signal, including South Island, New Zealand, although less so than temperature, as precipitation is more spatially variable and also some grid boxes may contain few input data.

Of the two New Zealand chronologies, Moa Park has a negative weight (Fig. 5), suggesting decreased (increased) growth with a positive (negative) index, whereas the positive weight of Putara indicates the opposite response. The response of these chronologies to local climate is known. Putara, together with other pink pine chronologies, is positively correlated with November–April local station temperatures and New Zealand average temperatures (D’Arrigo et al. 1998). Moa Park, conversely, is negatively correlated with New Zealand average December temperatures (Xiong and Palmer 2000a). The AAOI temperature and precipitation correlation maps (Figs. 7 and 8) show that lower pressure over New Zealand in the negative AAOI phase is associated with reduced temperatures and increased precipitation over South Island, which is characteristic of midlatitude maritime climates. The relationship of the chronologies with climate implied by the regression weights is thus physically sensible, indicating that these two chronologies are truly related to the AAOI, rather being part of the 5% which may be selected by chance in the prefiltering.

Norquino, the northernmost of the South American chronologies, is located in a region of negative precipitation and weakly positive temperature correlation coefficients (Figs. 8 and 7), indicating increased precipitation and lower temperatures due to low pressure to the west of South America during the negative phase of the AAOI, during which the negative weight of this chronology suggests increased growth and vice versa. This is in agreement with Villalba and Weblen (1997), who also found growth for this chronology to be strongly positively correlated with spring and early summer precipitation and negatively correlated with temperature during the current growing season.

The northernmost chronologies of Tierra del Fuego (Peninsula Brunswick and Monte Grande Magallanes) appear to be precipitation sensitive, as they are located in an area of significant negative precipitation correlations and of insignificant temperature correlations. The negative regression weights thus indicate increased growth with increased precipitation. The local precipitation response to the AAO in this region (Fig. 8) we surmise is due to the fact that a negative AAO is associated with decreased westerly/increased easterly
flow, which results in increased precipitation, possibly due to the orographic effect of mountains in the southwest of Tierra del Fuego. The other Tierra del Fuego chronologies, Estancia Carmen Camino, Estancia San Justo, Lago Yehuin, and Estancia Maria Cristina-Bosque Virgen, are located in the drier part of the island to the north of mountains (Boninsegna et al. 1990). A positive relationship with late spring and summer precipitation was found for a regional chronology that includes these chronologies as well as some positive relationships with summer temperatures (Boninsegna et al. 1990). Because the AAOI temperature and precipitation correlation coefficients (Figs. 7 and 8) are very small, it appears that they do not fully capture the effect of the AAOI on the...
local climate, which in turn affects tree growth. However, this region suffers from a sparsity of instrumental records (J. A. Boninsegna 2001, personal communication), thus the gridded climatologies are relatively uncertain. Additionally, as stated in section 2c, variation can be expected within a grid box, particularly in regions such as this with complex topography.

The TBR [Fig. 9 and Fig. 4 (lower)] was produced...
by multiplying the prewhitened tree-ring width series by the regression weights. The uncertainty bands on this reconstruction are much larger than those of the SBR because of the lower fraction of AAOI variance explained by the model (52% for the TBR compared to 86% for the SBR). Figure 9 shows that the decadal-scale signal in the TBR is less variable prior to 1860 than after. The decade from 1870 to 1880 has the lowest running mean index, there is then an upward trend to a maximum in the mid-1950s, followed by a trend towards a negative index until the record ends in the mid-1980s. This reconstruction shall be further discussed later.

Removing autoregression from a time series by prewhitening can reduce the low-frequency variability contained in that series, and hence any reconstruction derived from this series. To investigate whether the prewhitening has influenced the low-frequency variability in the reconstruction, a reconstruction was also produced using the unprewhitened chronologies (not shown). The interannual correlation between these two reconstructions is 0.76, and between the 9-yr running means 0.84, and the features in the TBR described above are present. Thus the prewhitening has not changed the main features of the TBR. The NCEP AAOI contains little autocorrelation (Fig. 10) and the TBR autocorrelation is much closer to the NCEP autocorrelation at most lags than the unprewhitened reconstruction, thus we consider the TBR to be more reliable than the unprewhitened reconstruction.

c. Comparison of the tree-ring- and station-based reconstructions

The validation of the SBR in section 3a shows that the SBR has considerable skill. Despite the uncertainties in the SBR related to the fact that it is based mainly on data from one center of action, we assume that the SBR is closer to the true AAOI than the TBR. Based on this assumption, we can validate the TBR against the SBR over the full period 1878–1985, therefore allowing also a validation of the low-frequency variability.

Given the various error sources for proxy-based reconstructions, the correlations between the SBR and the TBR are encouraging. 0.43 for the full period 1878–1985, 0.50 since 1900, and 0.56 for the fitting period (1948–85) (all significant at the 1% level). Although the significance of the correlations shows that the SBR and the TBR are clearly related, their relatively low values reflect the substantial differences between the two reconstructions. Despite these differences, the SBR and TBR can be regarded as consistent, because their 95% confidence intervals overlap for all but four time steps.

The low-frequency variability, as represented by the 9-yr running mean (Fig. 4), shows some common features, although the correlation over the whole period is only 0.26 (significant at the 1% level) and 0.38 for the fitting period (significant at the 5% level). The minima at around 1900 and 1940–50 are present in both reconstructions, as is the peak between 1930 and 1940, and the trend from positive to negative index between 1960 and 1980. The main difference is that the TBR shows a positive trend from 1900 to the late 1950s. This upward trend is less evident in the SBR, where the period from 1900 until the mid-1950s is dominated by negative values, followed by a period of positive index for five years, before a return to more negative values (Fig. 4). The linear trends over the period 1900–60 are 0.008 yr\(^{-1}\) for the TBR and 0.003 yr\(^{-1}\) for the SBR. This discrepancy shall be discussed further in section 4.

4. Discussion

A number of investigations of Southern Hemisphere circulation for recent decades have been undertaken, but no other studies have investigated the behavior of the AAO for the pre-NCEP period. We use these studies, particularly those that use other atmospheric data than the NCEP–NCAR SLP, as a consistency check for the SBR.

The trends in the SBR are compatible with the analyses of Thompson et al. (2000) and Chen and Yen (1997) over their common periods. The former identified a trend towards high index polarity of their Southern Hemisphere annular mode, based on 850-hPa NCEP geopotential height (GPH), over the period 1968–97, in all months except June. Similarly, using Australian Bureau of Meteorology (ABM) data for the period 1972–92, the latter identified a deepening of the circumpolar trough and an increase in midlatitude SLP in summer. Both correspond to a positive AAOI. The linear trends in the SBR over the periods 1968–97 and 1972–92 are indeed positive (0.01 and 0.02 yr\(^{-1}\)) respectively. It is important to reiterate that, because the SBR and TBR were calibrated against the detrended NCEP data on an
annual basis and because the reconstructions are produced using undetrended station data and chronologies, the trends in the two reconstructions are those contained in the trees and the stations and not those in the NCEP data.

Analyses of shorter intervals within the reconstruction period are also encouraging. A decrease in the strength of westerly flow at Southern Hemisphere high latitudes measured as DJF 500-hPa height differences between Chatham Island and McMurdo, Antarctica, was noted between 1957/58 and 1978/79 (Rogers and van Loon 1982), which is evident in both the SBR and the TBR. There is also agreement between the PC1 of Mo and White (1985) from Australian analysis Southern Hemisphere 500-hPa height anomalies for the period 1973–80 and the SBR and to some extent with the TBR (e.g., low AAOI summer 1976/77, high AAOI summer 1973/74). Thus our claim that the SBR is trustworthy, based on comparison with the detrended NCEP AAOI, is corroborated by these additional comparisons, and conversely the SBR confirms the findings of these studies.

The undetrended NCEP AAOI, as discussed previously, agrees relatively well with the TBR and SBR on an interannual basis, both for the fitting period and for the independent period of the TBR (1985–98) (Fig. 3). However, fitting a linear trend to the NCEP AAOI gives larger trends than for the SBR, of 0.03 yr⁻¹ over the period 1968–97 and of 0.06 yr⁻¹ over the period 1972–92. We cannot determine whether this is due to spurious trends in the NCEP data or because the SBR does not fully capture the true trend.

The larger positive trend in the TBR than in the SBR over the period 1900–60 identified in section 3c can possibly be explained by the observed increase in New Zealand temperatures since the beginning of the twentieth century, the rate of which reduced during the 1970s (Folland and Salinger 1995). The New Zealand chronologies, Moa Park and Putara, have negative and positive responses to temperature, respectively (section 3b). The tree regression weights (Fig. 5) are negative and positive for the former and the latter, respectively. Thus a temperature increase over New Zealand would result in an increased reconstructed AAOI, assuming no compensating trends caused by the Argentinean chronologies.

Because of the lower positive trend in the SBR over this period, we can conclude, however, that the observed temperature changes cannot be related to changes in the strength of the AAOI, thus either must have resulted from another mode of circulation, for example ENSO (Folland and Salinger 1995), and/or from noncirculation-related causes, for which anthropogenic warming is a candidate. This is also backed up by the regression coefficients between the AAOI and temperature over New Zealand of 0.2–0.4 (not shown), from which it can be calculated that the trend in the SBR of 0.003 yr⁻¹ would give a very small temperature increase of 0.04°–0.07°C over the period 1900–60.

We therefore suggest that the temperature-sensitive New Zealand chronologies are responding to a local temperature increase that is unrelated to the AAOI, leading to a positive trend in the TBR over the period 1900–60. However, this hypothesis is tentative, first, because the trend in the TBR is small and not significant and, second, because other trends may also be possible if one takes into account the confidence intervals.

Given the uncertainties in the TBR during the twentieth century, an important question is how the TBR compares to other climate reconstructions over New Zealand in the period before the documented non-AAO-related warming over New Zealand. As there are no reconstructions of the AAOI, we have to turn to reconstructions of other circulation features that may be related to the AAO. Salinger et al. (1994) produced reconstructions of zonal flow (the Auckland–Christchurch pressure gradient) from New Zealand tree-ring chronologies. During the pre-1900 period a number of features in the TBR are also present in this zonal flow reconstruction, namely the minima between 1760 and 1770 and between 1820 and 1830. The minima in the TBR during the 1860s and 1870s are also present, although to a lesser extent, in the Salinger et al. (1994) reconstruction. This may suggest that some confidence can be placed in the TBR before the twentieth century. However, the Salinger reconstruction is based on chronologies that are sensitive to temperature (as well as to precipitation), so it is also subject to potential changes in the link between the local climate and large-scale circulation. It should be noted that none of the chronologies are common between those used to construct the TBR and those used by Salinger et al. (1994).

Another Southern Hemisphere circulation index, that has been reconstructed is the TPI (Villalba et al. 1997). It represents the latitudinal displacement of the circumpolar vortex, and was defined by Pitttock (1980) as an index of pressure differences between Hobart and Stanley and by Villalba et al. (1997) as an index of differences between pressure over the New Zealand and the South America–Antarctic Peninsula sectors of the Southern Ocean. The relationship between the AAOI and the TPI seems to be unclear. The fact that the first of these two locations is located in a positive center of action of the AAO, and the second in the negative center, indicates that correlation between the TPI and the AAOI is likely. However, Rogers and van Loon (1982) suggest that the TPI is associated with their summer SLP PC2. Thus the TPI would be expected to have low correlations with their PC1 (and thus with the analogous AAOI). The comparison of the TPI and the TBR indeed shows similarities, for example, the upward trend between 1900 and the mid-1950s, and the subsequent downward trend between the mid-1950s and 1980s, and between 1780 and 1830. For other periods however, for example, between 1860 and 1890, the agreement is not so good.
Three of the nine chronologies used here were among the 12 used by Villalba et al. (1997) to produce their reconstruction. This may lead to some commonality between the reconstructions in the case that the relative weights of the three common chronologies in both reconstructions are similar.

The AAO is also connected to temperatures over much of Antarctica. The links between the AAOI and temperature evident from the correlation map in Fig. 7 enable investigation of whether climate variability in this region could be related to changes in the AAOI. Much of the Antarctic has negative temperature correlation coefficients with the AAOI. Thus increased (decreased) temperatures over these areas are related to a decreased (increased) AAOI. Note that this excludes the Antarctic Peninsula, which has been found to exhibit differing behavior to much of the rest of the continent (Rogers 1983). These correlations agree with the results of Rogers and van Loon (1982), who found that higher (lower) temperatures at Amundsen–Scott Station occurred in years with weak (strong) Southern Hemisphere westerlies during both summer and winter. This link between the strength of the southern westerlies and Antarctic temperatures was also found by Raper et al. (1984).

Instrumental temperature measurements over the Antarctic between 1957 and 1982 showed a statistically significant positive trend in DJF of 0.032°C yr⁻¹ (Raper et al. 1984), consistent with the negative AAOI trend in the SBR, indicating that at least some of this warming could be related to AAO-related circulation changes. We can estimate the warming that could be expected from the influence of the AAO over this period from regression coefficients calculated between the NCEP AAOI and NCEP 850-hPa temperature. The trend calculated over 1957–82 in the SBR is −0.037 yr⁻¹. The regression coefficient over much of Antarctica varies between −0.2 and −0.6 (not shown), thus, depending on the location, the warming associated with this trend in the AAOI would be between 0.007°C yr⁻¹ and 0.022°C yr⁻¹, that is, 22% to 69% of the observed warming can be linked to AAOI changes when we assume that the observed trend at all locations is the same as the trend found by Raper et al. (1984).

The influence of the AAO on Antarctic climate in individual years is also apparent from the reconstructions. An interesting result is the presence of the lowest value of the SBR series in summer 1911/12, which is also one of the lowest TBR values. From the correlation map relationships described above one would expect increased temperatures over much of Antarctica. Indeed, this year was described as having difficult weather conditions, with unusually high temperatures, for the Scott Expedition to the South Pole (Schwerdtfeger 1984; Villalba et al. 1997), with temperatures of +1.1°C over the Beardmore Glacier (83°20′S, 170°W). We can again estimate the temperature anomaly expected from the SBR AAOI value using the regression coefficients, which are −0.4 to −0.5 in this region. Combined with the AAOI in this year of 2.35, this would give a warming above the fitting period mean of 0.94°–1.18°C.

Regarding the New Zealand regional atmospheric circulation, it appears that the AAO has had a strong influence on the circulation over this region in the period 1930–75 and had a weak influence in the period 1976–94. Salinger and Mullan (1999) used station records from the National Institute of Water and Atmospheric Research to investigate circulation changes over New Zealand. They found that the period 1930–94 could be divided into three circulation periods: 1930–50 with anomalous south to southwest airflow, 1951–75 with increased easterly and northeasterly airflow, and 1976–94 with west to southwesterly airflow predominating. The first period is indeed one of dominantly negative AAOI in the SBR (Fig. 4), which brings westerly or southwesterly flow to New Zealand (Fig. 1). The second interval shows a shift to a positive AAOI, that is, easterly anomalies in both reconstructed series, again agreeing with Salinger and Mullan. The final period however shows a short period of westerly anomalies and then a move to easterly anomalies, suggesting that another climatic influence must be overriding the AAO.

5. Conclusions

A reconstruction of the austral summer (Nov–Jan) Antarctic Oscillation index (AAOI) has been produced for the period 1878–2000 from long station SLP records (termed the SBR). This reconstruction can be regarded as relatively reliable because of the high coefficient of multiple determination in the fitting period and the high coefficient of multiple determination and the high reduction of error during the validation period. The SBR is also in agreement with trends in the AAOI from other datasets for the period for which comparison is possible. This is the first time that the behavior of the AAO has been discussed for a period longer than that of the NCEP or the Australian Bureau of Meteorology data of approximately 50 years. The SBR shows that for the first half of the twentieth century the NDJ AAOI was dominantly negative, changing to a period of positive values during the 1960s, followed by a trend to negative values until 1980, then by a trend towards positive values from 1980 until the end of the reconstruction. However, when interpreting the SBR outside of the calibration period, the increased uncertainty related to the stability of the underlying regression model outside of the calibration period, which is particularly pertinent here because the SBR is based dominantly on stations from one center of action of the AAO, should be borne in mind.

A second reconstruction has been developed using tree-ring width chronologies from New Zealand and South America for the period 1743–1985 (termed the TBR). The tree–climate relationships implied by the regression model are physically sensible, agreeing in many cases with relationships found in published lit-
temperature on the individual chronologies. As high confidence can be placed in the SBR in comparison to the TBR, the former has been used to evaluate the latter. Significant correlations on interannual and longer timescales were found, but also clear differences. For example, the TBR contains an upward trend from the beginning of the twentieth century until 1960 that is weaker in the SBR. Differences between the two reconstructions may be partly due to the different spatial distribution of the stations and the chronologies, but may also be caused by nonclimatic influences on the chronologies or by temperature or precipitation changes that are unrelated to the AAO. We hypothesise that this TBR trend, which corresponds to a period of observed temperature increase over New Zealand, may be a result of increased growth of the temperature-sensitive New Zealand chronologies caused by a non-AAO-induced warming, which leads to a decreasing reconstructed AAOI.

Support for the TBR before the late nineteen-tenth century is given through moderate agreement with the New Zealand zonal flow reconstruction of Salinger et al. (1994), although this reconstruction was also produced from temperature- (and precipitation-) sensitive chronologies and therefore may also be affected by temperature changes that are not induced by the reconstructed circulation. There is also some agreement between the SBR and the Trans-Polar index reconstruction of Villalba et al. (1997).

The TBR should be regarded as a first estimate for the preinstrumental AAO, the optimal given the data available. The uncertainties are clearly too high to draw definite conclusions on climate variability during the preinstrumental period. Data from additional geographical areas where the AAO has a surface climate signal, thus where proxy data may contain an AAOI signal, could help reduce these uncertainties. Such areas, evident from temperature and precipitation correlation maps, are southeastern and northwestern Australia, southern Africa, and Antarctica.

For the NCEP period, the regional climate responses to the NDJ AAO have been investigated. This has shown that the temperature increases over New Zealand since the 1950s are not linked to the AAO. Moreover, a substantial fraction of the observed austral summer temperature changes over much of Antarctica between the late 1950s and the 1980s can be linked to a negative AAOI trend during this period.

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