Predictability of snow-depth anomalies over Eurasia and associated circulation patterns

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SUMMARY

This study investigates the variability and predictability of snow depth anomalies over the Eurasian continent at the end of winter, as represented in 12 ensembles of General Circulation Model simulations performed at the European Centre for Medium-Range Weather Forecasts. Each ensemble includes nine integrations performed with the same prescribed sea surface temperature, but started from time-lagged initial conditions. An empirical orthogonal function (EOF) analysis shows that the leading EOF of Eurasian snow depth in March has a zonally-oriented dipole structure, with a band of positive anomalies covering northern Europe and Siberia, and negative anomalies over central Europe, the Himalayas and north China. A significant relationship is found between the positive/negative phase of this snow-depth anomaly and warm/cold El Niño Southern Oscillation events. The positive phase of the snow-depth EOF1 is associated with a wintertime circulation characterized by a strengthening of the westerly winds over Europe and Siberia; in the upper troposphere, this westerly anomaly is accompanied by negative zonal wind anomalies over Eurasia around 30–40°N and positive zonal wind anomalies between the equator and 25°N over Africa and south-east Asia. A good degree of predictability is found in the snow-related circulation anomalies: considering 500-hPa height, 850-hPa zonal wind and 200-hPa zonal wind, the interannual variations of the ensemble-mean fields show a correlation of 48%, 56% and 65% (respectively) with the corresponding observed anomalies over the eastern half (0° to 180°E) of the northern hemisphere. The tropical component of the zonal wind anomaly associated with snow-depth EOF1 is strongly predictable; it shows a marked persistence from winter to the early summer, and affects the large-scale circulation over south Asia in the early and central periods of the monsoon season.

KEYWORDS: Eurasian circulation anomalies Seasonal predictability Snow-depth variability

1. INTRODUCTION

The prospect for seasonal atmospheric prediction is based primarily on the assumption that lower-boundary forcing can influence the statistical properties of the atmospheric circulation on time-scales longer than the limit of predictability of individual weather patterns. The fundamental problem to be addressed is the separation between internal chaotic atmospheric fluctuations, which have no predictability beyond one or two weeks, and the influences of anomalies in sea surface temperature (SST) and land surface conditions.

In the case of SST anomalies, it is recognized that the ocean dynamics generates large-scale modes of variability which are predictable for at least a few months (e.g. Latif et al. 1994). Land surface conditions, on the other hand, are totally dependent on local exchanges of energy and moisture between the surface and the atmosphere, and it is not yet understood whether their influence is limited to a ‘low-pass filtering’ of local atmospheric variations, or may extend to wider regions through atmospheric teleconnection patterns.

Snow cover over the northern continents has long been recognized an important component of the climate system on seasonal as well as longer time-scales, with the capacity of influencing the atmospheric conditions in the northern spring and early summer. In particular, the effect of Eurasian snow cover on the development of the Asian summer monsoon has been advocated for more than a century (Blandford 1884), with supporting evidence coming from both observational (Hahn and Shukla 1976; Dickson 1984; Sankar-Rao et al. 1996) and modelling studies (Barnett et al. 1989; Vernekar et al. 1995; Douville and Royer 1996).

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Recently, evidence of a precursor signal for Eurasian snow anomalies has been found in El Niño-Southern Oscillation (ENSO) indices (Yang 1996). This result suggests that the snow cover may be predictable a few months in advance as a result of the ocean forcing during the winter season. On the other hand, it raises the possibility that the Eurasian snow–monsoon relationship is just a passive reflection of global-scale connections between ENSO and the Eurasian circulation (Webster and Yang 1992; Ju and Slingo 1995), and therefore may diminish the importance of snow cover as an independent predictor of the monsoon.

The difficulty in establishing the significance of any relationship between Eurasian snow cover (or snow depth) and atmospheric circulation anomalies arises partly from the scarcity of snow observations over large parts of Asia. This uncertainty also affects the interpretation of general circulation model (GCM) studies on the snow–monsoon relationship, since one should evaluate how realistic are the snow anomalies imposed or generated in such experiments. On the other hand, simulations made with state-of-the-art GCMs may themselves be a source of information on the space and time variability of snow anomalies, and their association with modes of atmospheric variability which are affected by SST forcing.

This paper is focused on the variability and predictability of snow depth anomalies over the Eurasian continent at the end of winter, as represented in 12 ensembles of GCM simulations performed at the European Centre for Medium-Range Weather Forecasts (ECMWF). Each ensemble includes nine integrations performed with the same prescribed SST but started from time-lagged initial conditions. The analysis of this dataset, which has much greater extent and homogeneity than the available observational record, allows a clear separation of the effects of boundary forcing and internal atmospheric dynamics in the generation of snow anomalies. The present study is intended to provide a guidance to further numerical experimentation aimed at clarifying the role of snow anomalies in the short-term climate variability of the Eurasian circulation.

The paper is structured as follows. The ensemble integrations are described in more detail in section 2. The leading modes of variability of modelled Eurasian snow depth and their predictability are described in section 3. In section 4 the relationship between snow depth and SST anomalies is considered. Associations with wintertime flow anomalies are discussed in section 5. In section 6, the time-evolution of snow-related flow anomalies into the following summer season, and their link with monsoon indices, are investigated. A summary of the results is given in section 7.

2. DATASET

This study analyses 12 winter ensembles of 120-day integrations performed at ECMWF as a part of the European project ‘PRediction Of climate Variations On Seasonal to interannual Time-scales’ (PROVOST). The model is the ECMWF spectral model (Simmons et al. 1989) cycle 13r4 (part of ECMWF’s Integrated Forecasting System), used at triangular truncation T63 with 31 levels in the vertical. The integrations considered here have been run over the winters 1981/82–1992/93 using prescribed observed SSTs extracted once a day from the ECMWF re-analysis (ERA; see Gibson et al. 1997) and based on the optimum-interpolation analyses by Reynolds and Smith (1994). Each ensemble consists of nine integrations started using a time-lagged technique, with initial data spatially interpolated from ERA fields and separated by 24 hours. All runs start between 22 and 30 November and end on 31 March. Within a given ensemble, each
integration differs from the others only in the initial conditions (Branković and Palmer 1998).

Monthly mean values of surface and upper air fields have been extracted from the 108 (12 × 9) integrations. For the snow data, the analysis is focused on March monthly means, which are representative of the maximum snow cover at the end of each winter. To investigate the relationship with the atmospheric circulation in the winter season, data from January, February and March are used. December data are not analysed, since the individual ensemble members cannot be considered as independent integrations during a period when the influence of initial conditions is still strong. ECMWF re-analysis is used to verify the ensemble experiments; data from the re-analysis of the United States National Centers for Environmental Prediction (NCEP) (Kalnay et al. 1996) will also be used in section 6.

3. SNOW DEPTH VARIABILITY OVER EURASIA

(a) Model simulations

Model simulations of Eurasian snow depth anomalies in March are shown in Fig. 1. For each year the difference between the ensemble mean and the model climate (i.e. the mean over the whole dataset) is plotted. Some interesting features can be noticed. There are several cases (see for example years 1983, 1984, 1985, 1986, 1987) in which snow depth anomalies of the same sign cover major portions of the Asian continent while in others years (1982, 1990, 1991, 1992) anomalies are located over relatively small portions of the Eurasian mountain ridges. March 1993 exhibits high positive snow depth anomalies almost everywhere, appearing the most anomalous year of our sample (see discussion in section 3(b) below).

The leading variability patterns of the model snow depth in March were searched for by computing empirical orthogonal functions (EOFs) of monthly anomalies from the whole dataset, in the sector from 0° to 180°E. Figure 2 shows the first three EOFs, which explain 16.3%, 11.5% and 10.7% of the total variance respectively. It is worth noting that none of the leading EOFs exhibits anomalous patterns of the same sign over the whole target area. Anomalies covering a large portion of Eurasia are always accompanied by significant anomalies of the opposite sign centred somewhere else on the continent, suggesting a shift in the location of the main precipitation areas. EOF1 exhibits a north–south dipole structure marked by a positive anomaly of large longitudinal extent, positioned on the northern part of Eurasia (from northern Europe to the eastern part of Siberia) and three (more localized) negative anomalies over eastern Europe, the Himalayas and north China. An essentially east–west dipole characterizes EOF2, while EOF3 shows smaller-scale anomalies located over west Asia, Scandinavia and Tibet.

The time-series of principal components associated with EOF1 are shown in Fig. 3. For each integration the EOF1 principal component (PC1 hereafter) is represented by a black dot, while the solid line denotes the average over the nine members of each ensemble. Shading marks the interval of one ensemble standard deviation centred around the ensemble mean. A considerable interannual variability is found. The peculiarity of March 1993, which is denoted by the highest ensemble mean, is confirmed by the PC analyses.

The PC spread within an ensemble, which can be assumed as a measure of predictability of the anomaly in a given year, also displays a significant interannual variability. Usually, low spread is considered as an indication that the lower boundary forcing can induce a significant reproducibility on snow anomalies development, in spite of the
Figure 1. Snow depth anomalies in March from winter ensembles, defined as differences between the ensemble mean for each year and the mean over the whole dataset. (a): 1982 to 1987; (b): 1988 to 1993. Unit = mm of equivalent water. Contours at ±5, 10, 30, . . . mm; negative values dashed.

different atmospheric initial conditions. The ensemble spread of PC1 does not show any correlation either with the sign or with the magnitude of the mean snow anomaly of a given year.

To quantify the seasonal predictability of any climatic variable, an index of reproducibility $R$ (Webster et al. 1998) can be defined as:

$$R = \frac{V_e}{V_{\text{tot}}}$$

where $V_{\text{tot}}$ and $V_e$ are the total and ‘external’ variance of the model. In our case $V_e$ is defined as the variance of the ensemble means. For the snow-depth PC1, this index is equal to 38% when the full 12-year sample is considered, but it decreases to 23% if data for March 1993 are excluded (see below). If the means of nine-member ensembles were just averages of nine random values taken from a standardized gaussian distribution, their expected variance would be equal to $1/9$ (i.e. about 11%). According to standard tests, the hypothesis that the interannual variability of ensemble-mean PC1 is random can be rejected with a confidence level exceeding 99.5% for the 12-year record, with a 98% confidence level excluding March 1993.

If the interannual variability of PC1 has a dynamical origin, a likely explanation (consistent with results by Yang 1996) is that such a variability is a consequence of changes in the large-scale circulation over Eurasia which are forced by SST anomalies. Indeed, large-positive values of ensemble-mean PC1 are present in March 1983 and 1987, corresponding to El Niño conditions during winter, while wintertime La Niña
events (e.g. in 1984–85 and 1989) are marked by negative ensemble means. Although mild El Niño conditions occurred in winter 1992/93, the ensemble-mean PC1 for that year seems far too large compared to the PC values in other (and stronger) ENSO events.

Another possible explanation is that the snow anomalies in late winter reflect to some extent the snow anomalies already present at the end of November, i.e. included in the initial conditions of the ensembles. This hypothesis will be checked in the next subsection by examining the snow-depth data in the ECMWF re-analysis.

**(b) Re-analysis dataset**

In the study of interannual variations, re-analysis data are particularly valuable because they are the product of a fixed data assimilation scheme. The snow mass (water equivalent snow depth) is analysed six-hourly in the ECMWF re-analysis, using surface-based observations of precipitation and of snow depth (Mahfouf and Viterbo 1998). The snow assimilation scheme can be summarized in three points:

- First a snowfall analysis is produced: depending on temperature, precipitation is interpreted as snowfall and a gridded value is built by spatial interpolation.
- Then a background field for snow mass is built using persistence and the snowfall, with a very simple model for snow melting (depending on temperature). To avoid drifts in data void regions, a relaxation to climatology is also applied.
- Finally observations of snow depth, assumed at a constant snow density, are combined with the background field to provide an analysis of snow mass.
Although realistic in data rich areas, the snow analysis defaults to climatology and loses realism in data void areas. Unfortunately this problem affects mainly the former Soviet Union (the 'core' region of EOF1), where the data used are very inhomogeneous in time. Actually, the number of available observations in that area increased dramatically (from about 1000 to about 10000) from 1991 to 1992 (P. Kallberg, private communication).

An example is given in Fig. 4 which compares the snow depth re-analysis over North Asia in 1982, 1987, 1991 and 1992, averaged over the 9 days (22 to 30 November) in which the ensemble runs are started. A climatological constant value characterizes all the late-November periods except that of November 1992, suggesting that snow-depth initial conditions over Siberia had little influence on the simulations preceding 1992–93. Thus, apart from winter 1992/93, the Eurasia snow-depth anomalies found at the end of the integrations are mostly due to the atmospheric circulation during the winter, rather than to the snow initial condition. On the other hand, the anomalous nature of the snow
SNOW DEPTH ANOMALIES OVER EURASIA

Figure 3. Time series of the standardized principal component associated to snow-depth EOF1 from each experiment of the winter ensemble integrations (black dots). The solid line denotes the average over the nine members of each ensemble. The interval of one ensemble standard deviation centred around the ensemble mean is shaded.

Figure 4. Snow depth fields from ECMWF reanalysis averaged over 22 to 30 November. (a) 1982; (b) 1987; (c) 1991; (d) 1992. Unit = mm of equivalent water. Contours at 30, 50, 70, ... mm.

The depth field of March 1993 (Fig. 1(b)) simulated by the seasonal ensemble suggests that November initial conditions may have a large influence on the snow simulations, which will not be captured for earlier years because of the deficiencies of the ERA snow fields.

Snow depth re-analyses in March show a similar inconsistency (not shown). Anomalies of small amplitude are found in the monthly means from 1982 to 1991; while 1992 and 1993 are characterized by high positive snow depth anomalies almost everywhere over the area (50–80°N, 40–180°E). This inhomogeneity in the ECMWF re-analysis makes it unsuitable for a direct verification of the seasonal prediction of Eurasian snow.
It is important to establish whether the bias in the snow simulations of March 1993 is such as to affect the EOF structure in any significant way. Fortunately, recomputations of such EOFs without the 1993 data have produced almost identical spatial patterns, and consequently only marginal differences in the PC time series in all other years. Therefore, in the rest of the paper we will continue to use PCs from the 12-winter EOF decomposition, but the 1993 values will be excluded from the analysis of the relationships between snow-depth fields and either forcing fields or circulation patterns.

4. RELATIONSHIP BETWEEN SNOW DEPTH AND SST ANOMALIES

After assessing the impact of the initial conditions in the snow field, we will now analyse the relationship between the interannual variability in snow depth and SST anomalies in a more objective way. Figure 5 shows the covariance map between SST anomalies in the ECMWF re-analysis, averaged in January, February and March of each year, and the values of the March snow-depth PC1 in ensemble experiments (excluding 1993 data because of the initial-condition bias). Since the SST field is the same for each member of a given winter ensemble, the covariance map can be computed directly from the ensemble-mean values of PC1. Consistently with Yang's (1996) results, the covariance shows a strong and coherent signal in the eastern tropical Pacific, reminiscent of a typical El Niño anomaly, with a maximum exceeding 0.5 K.

When the covariance map is turned into a correlation map (not shown), the eastern Pacific maximum corresponds to a correlation of 92% between SST and ensemble-mean values of PC1. However, when individual ensemble members are considered, the correlation decreases to 44% because of the larger variability of single realizations. (By definition, the correlation of SST with data from individual members cannot exceed the square root of the reproducibility index). Since the real atmosphere provides just one realization per year, the 44% value may be considered as an estimate of the correlation between observed snow anomalies and ENSO indices. However, the multiple realizations provided by the ensemble experiments allow a much stronger confidence (exceeding the 99.5% level) in the statistical significance of this relationship than the available observed record.
Since correlations between SST anomalies and other PCs of late-winter snow depth show a weaker and less coherent signal (not shown), we focus our investigations on the atmospheric variability associated with the first snow-depth EOF. In the ECMWF model, this EOF pattern describes the response of the Eurasian snow depth to tropical forcing from ENSO events, even though the substantial intraseasonal variability of this mode suggests that it would be prominent even in the absence of anomalous SST forcing. The link between SST and Eurasian snow should be interpreted as the result of changes in the northern extra-tropical circulation caused by tropical SST forcing, which in turn modify the intensity and distribution of precipitation over Eurasia. The following analyses will identify the circulation patterns that generate (but may also be affected by) the leading snow-depth anomaly, estimate their predictability from the ensemble experiments and examine their seasonal evolution into the following summer season from observational data.

5. SNOW DEPTH AND WINTER FLOW ANOMALIES

In this section, the relationship between the first PC of Eurasian snow depth in March (PC1) and the atmospheric circulation during the previous winter is studied. Monthly means of geopotential height at 500 hPa and zonal wind at 200 and 850 hPa from the winter ensemble integrations are considered.

(a) Geopotential height

The covariance patterns of March snow PC1 and 500-hPa monthly-mean anomalies in January, February and March have been computed from a dataset including all winter integrations, except those in winter 1992/93 (see section 3(b)). The January covariance pattern (hereafter SGH, 'Snow-Geopotential Height') is shown in Fig. 6(a). The largest covariances are found over the Eurasian continent, from 30° to 90°N and from 0° to 120°E, with a large negative anomaly centred around 70°N, 45°E and two positive centres over the Mediterranean Sea and central Asia. The SGH pattern associates increased snow in northern Eurasia with a reinforcement of zonal winds over most of northern Europe and Siberia, a favourable condition for increased baroclinic activity. The correlation with snow depth PC1s reaches 0.6 (−0.6) over the three main centres.

The SGH pattern is very similar to the 500-hPa Euro-Atlantic EOF4 (Fig. 6(d)) computed from 45 winters of NCEP analyses by Pavan and Molteni (1998). Furthermore SGH bears some resemblance to the 500 hPa height component of the 'cold ocean–warm land' (COWL) pattern (Wallace et al. 1995), which is related to the Northern Hemisphere mean surface-air-temperature anomalies during winter. The February covariance pattern (Fig. 6(b)) looks like the January pattern, while the corresponding pattern for March (Fig. 6(c)) shows much weaker features, confined to high latitudes.

Projections of the 500-hPa height anomaly fields onto the SGH pattern have been computed for all the individual integrations in the 12-winter dataset. The correlation between the time-series of SGH projections and snow PC1 reaches 78% when the mean anomaly of January and February is considered. (In computing the correlation index, the ensemble runs of 1992/93 have been excluded). The SGH pattern shown in Fig. 6(a) can be used to verify (indirectly) the ensemble seasonal prediction of the snow field using the re-analysis data. In practice, the correlation between the projections of analysed and simulated anomalies of height onto the SGH pattern can be used as proxy of the correlation between observed and modelled snow depth PC1.
Some measures of the predictability of the SGH pattern are given by the statistics in the first row of Table 1, which lists the total variance of the projections of January–February (JF) 500-hPa height anomalies from individual experiments, the ensemble-mean variance, the reproducibility index and the correlation between ensemble-mean and re-analysis projections. The reproducibility index for the SGH pattern is equal to 20%, slightly less than the value obtained for the snow PC1 (23%) when the data of March 1993 are excluded.

The model performance with respect to the re-analysis shows a moderate degree of predictive skill. The correlation between re-analysis and ensemble mean values is 48%. Such a correlation can be further explored by looking at Fig. 7, which displays the time-series of JF-mean projections onto SGH in same format used for PC1 values in Fig. 3. In addition to data from ensemble members (black dots) and ensemble means (solid line), the projections of the re-analysis anomaly (computed with respect to the observed mean field) are represented by a dashed line.
TABLE 1. (a) TOTAL VARIANCE ($V_{tot}$), ENSEMBLE-MEAN VARIANCE ($V_e$), REPRODUCIBILITY INDEX ($R$) AND CORRELATION BETWEEN ENSEMBLE-MEAN AND RE-ANALYSIS VALUES ($\text{cor}$) FOR THE PROJECTIONS OF 500-hPa HEIGHT ANOMALIES ONTO THE COVARIANCE PATTERN BETWEEN SNOW-DEPTH PC1 AND JF 500-hPa HEIGHT; (b) AS IN (a), BUT FOR THE PROJECTIONS OF WIND ANOMALIES ONTO THE COVARIANCE PATTERN BETWEEN SNOW PC1 AND JF ZONAL WIND AT 200 hPa; (c) AS IN (b), BUT FOR ZONAL WIND AT 850 hPa

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<th>$V_{tot}$</th>
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<tr>
<td>(a) Z 500 hPa</td>
<td>1.71</td>
<td>0.35</td>
<td>0.20</td>
<td>0.48</td>
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<td>(b) $u$ 200 hPa</td>
<td>2.37</td>
<td>0.90</td>
<td>0.38</td>
<td>0.65</td>
</tr>
<tr>
<td>(c) $u$ 850 hPa</td>
<td>1.99</td>
<td>0.62</td>
<td>0.31</td>
<td>0.56</td>
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Figure 7. Time series of the projections of January–February mean 500-hPa height anomalies onto SGH. Each model integration is represented by a black dot. The solid line denotes the average over the nine members of each ensemble, the dashed line represents the projections of the re-analysis anomalies. The interval of one ensemble standard deviation centred around the ensemble mean is shaded.

The simulations appear more accurate during the first half of the 12-year period considered. In particular, the large oscillation of the SGH index between 1983 and 1987 is quite well reproduced by the model. During the second period, the re-analysis index exhibits quite large variations from year to year (for example from 1988 to 1989) which are mostly missed by the model. Comparing Fig. 3 with Fig. 7, it is notable that also the correlation between ensemble mean values of the snow PC1 and the SGH index is higher during the 1982–87 period than in the second half of the record (1988–93), when the absolute value of the SGH projections is weaker.

(b) Zonal wind

The anomaly wind patterns associated with the leading EOF of Eurasian snow depth were identified by computing, for each winter month, the covariance between the snow
Figure 8. Covariance patterns between the time series of March snow-depth PC1 and 200-hPa zonal wind anomalies from winter ensembles in (a) January, (b) February, and (c) March. Contours every 1 m s^{-1}, negative values dashed, zero-contour omitted.

PC1 and the zonal wind anomaly time series simulated by all ensemble experiments in winter 1981/82 to 1991/92. Figures 8 and 9 show the covariance patterns at 200 and 850 hPa respectively.

At the upper level (Figs. 8(a), (b), (c)), the January and February anomalies are quite similar, and are characterized by zonally-elongated patterns covering a major portion of the northern hemisphere. Anomalies over Eurasia and northern-central Africa exhibit a north–south three-pole-like structure. A strengthening of the jet at high latitudes (45–70°N) is accompanied by weakened westerly winds over the east Mediterranean and north-east China, while a westerly anomaly prevails over south Asia and central Africa (0–25°N). A similar pattern, but with opposite sign and a generally smaller amplitude, characterizes the upper level circulation over the western half of the northern hemisphere (30–180°W). Here the zonal wind anomalies are positive between 20° and
Figure 9. As Fig. 8, but for 850 hPa zonal wind anomalies. Contours every 0.5 m s$^{-1}$, negative values dashed, zero-contour omitted.

45°N, negative over North America and the eastern equatorial Pacific. The signal over America and east Pacific is strongest in February.

The upper-level covariance pattern in March shows a similar (zonal) structure and polarity to the January and February patterns, but its amplitude is differently distributed as a function of latitude. Most of the signal is now located in the tropics, while the anomalies over the extratropics are considerably weaker.

The zonal wind pattern at the 850-hPa level (see Figs. 9(a), (b), (c)) associated with the snow-depth PC1 shows more localized anomalies than its upper-level counterpart. Again the results for January and February are quite similar. The strongest anomalies, located over Eurasia between 40 and 70°N, are associated with increased westerly winds over northern Europe and Siberia, while an easterly anomaly is located north of 70°N. Apart from this equivalent barotropic signal, the most significant anomaly is found in the tropics, where the trade winds are anomalously weak over most of the equatorial
Pacific. This anomaly, typical of a warm ENSO event, persists into March, when (as for
the upper level case) the signal at high latitudes becomes weaker.

Considering together the patterns shown in Figs. 8 and 9, the zonal-wind covariance
with PC1 seems to be part of a planetary-scale circulation extending in both the tropics
and the extratropics of the northern hemisphere. In the tropical regions, the upper-
tropospheric anomalies are strongly reminiscent of the ENSO-related global signal (see
Fig. 13 of Webster and Yang 1992, hereafter WY92). Therefore, it is logical to interpret
the correlation between the snow PC1 and Pacific SST, discussed in section 4, as the
combined result of global and regional phenomena, namely:

(a) the forcing of the circulation pattern described in Figs. 8 and 9 by tropical SST
anomalies;
(b) the change in the distribution of precipitation (i.e. snow deposition) over Eurasia
caused by the shift in the areas of maximum baroclinicity which are associated with the
zonal wind anomalies.

The variability and predictability of the snow-wind covariance patterns described
above is analysed in Fig. 10. The two panels, referring to 200 hPa (panel a) and 850 hPa
(panel b) wind respectively, are constructed in the same way as Fig. 7, and show the
projections of JF wind anomalies (in the area 0–180°E, 0–90°N) simulated by the
ensemble integrations onto the corresponding covariance patterns. Ensemble means
and re-analysis projections are also shown, and the ensemble standard deviations are
indicated by a shaded band. The total and ensemble-mean variance, the reproducibility
index and the correlation between ensemble mean and re-analysis are listed in rows (b)
and (c) of Table 1.

The indices listed in Table 1 indicate a stronger predictability for the snow-wind
covariance patterns than for the corresponding 500-hPa height field (SGH). For the 850-
hPa zonal wind covariance, which is mainly localized in the Eurasian extratropics, the
difference in correlation with respect to SGH is relatively small. On the other hand,
the 200-hPa zonal wind pattern, which includes significant tropical features, has a
substantially larger reproducibility than the SGH pattern, and a better agreement with
the re-analysis (with a 65% correlation). In common with the SGH index, the interannual
variability of wind projections (shown in Fig. 10) is better simulated by the ECMWF
ensembles in the first half of the 12-year period.

6. SNOW-DEPTH ANOMALIES AND LARGE-SCALE MONSOON CIRCULATION

The results presented in section 5(b) appear in agreement with both observational
and modelling studies on wintertime tropical precursors of Eurasian snow depth anoma-
lies. In the ECMWF ensemble experiments analysed here, positive snow depth anoma-
lies over north Eurasia are found at the end of winters characterized by strong upper-
troposphere westerlies over south Asia. Barnett et al. (1989), using a low-resolution
GCM, found that weak westerlies over south Asia were accompanied by deficient winter
snowfall over Asia. In a recent observational study, Yang (1996) found a precursor signal
for Eurasian snowfall in (the opposite of) the Southern Oscillation Index. Yang’s results
are consistent with the snow-SST covariance shown in Fig. 5 and the snow-related wind
patterns in Figs. 8 and 9, which are characteristic of a global circulation associated with
warm ENSO events.

It is interesting to investigate whether our results are also consistent with the statisti-
cal relationship between Eurasian snow cover in winter/spring and monsoon rainfall
in the following summer found in many studies (e.g., Hahn and Shukla 1976, Dickson
1984; Yasunari and Seki 1992; Sankar-Rao et al. 1996). The tropical covariance between March snow PC1 and JF upper-troposphere zonal wind (Figs. 8(a), (b)) resembles the DJF large-scale circulation patterns found by WY92 as precursor to weak monsoons. In WY92 no signal was found at 850-hPa level in winter between 45°N and 45°S: in the tropical lower troposphere, a significant difference between weak and strong monsoon years occurs only in the late spring and summer. Our analysis yields a similar result: the
snow-related 850-hPa zonal wind anomalies over Eurasia (Figs. 9(a), (b)) are positioned north of 45°N.

These findings suggest that the same large-scale, long-lasting anomalous circulation system may determine both winter/spring snow depth anomalies over Eurasia and anomalous Asian summer monsoons (although, as discussed in the next section, this does not necessarily mean that the snow anomaly is a purely passive component). An indicator of such a broad-scale winter flow over Eurasia has been defined using the anomaly wind patterns shown in Figs. 8 and 9. Figure 11(a) shows the Hovmoeller diagram for January, February and March (JFM) of the zonal mean difference between the 200 and the 850 hPa covariance patterns, computed for each month in the sector 40°-110°E. This three-pole pattern is representative of the wintertime evolution of the large-scale anomalous system which, in the ECMWF model, induces positive snow-depth anomalies over north Eurasia, and may evolve into a weak monsoon circulation in the following summer.

It is not possible to verify directly the modelled evolution of this circulation anomaly into summer, since the ECMWF winter ensemble integrations analysed here terminate on 31 March. However, this issue may be explored using re-analysis data (monthly means from the NCEP re-analysis are used in this section in order to extend the analysis to 1994). Figures 11(b) and 11(c) show two Hovmoeller diagrams of analysed wind shear anomaly (\(U_{200}' - U_{850}'\)), zonally averaged between 40° and 110°E. The first diagram is for years 1982 to 1988, the second shows the period 1988-1994. Anomalies are filtered using a three-month running mean.

The JFM snow-related pattern shown in Fig. 11(a) (hereafter SWS, 'Snow-Wind Shear') is clearly recognisable in the NCEP data for winters 1983, 1990, 1992 and 1993, while JFM 1984, 1985, 1988 and 1994 are characterized by its opposite phase. The high-latitude signal (50°-80°N), when it is significant, appears in general during the cold season, while the tropical and mid-latitude anomalies seem more persistent. In particular, the anomalies in the tropical belt persist through winter and spring up to mid-summer.

In order to evaluate the relationship between the monsoon circulation and the SWS anomaly, an index was computed by projecting the time-series of JFM zonal winds from the re-analysis onto the SWS pattern itself. An even simpler index of the tropical wind-shear pattern (TWS) was also defined as the zonal wind shear anomaly (\(U_{200}' - U_{850}'\)) averaged over the area (0-20°N, 40-110°E). For the summer months, the TWS index is the opposite of the index defined by Webster and Yang (WY92) as a measure of intensity of the large-scale monsoon circulation. A high value of the TWS index during the summer indicates a weak monsoon, while a low value is related to an enhanced monsoon circulation.

The two indices are highly correlated not only during winter (as expected) but also up to late spring. Lagged correlation values (listed in Table 2(a)) decrease in summer, after the onset of the Asian summer monsoon, but are still significant until mid-summer. In the last period considered (July-August-September), the correlation drops to zero. These results confirm the relationship between the continental-scale snow-related flow anomaly during winter and the monsoon circulation in the following summer.

The high correlation values imply a marked persistence of the TWS index from winter to the early-mid summer, consistent with the findings of WY92 and Ju and Slingo (1995) on the persistence of the tropical wind anomalies over south Asia. Lagged correlations of the TWS index are listed in Table 2(b). The monsoon circulation index exhibits a considerable auto-correlation up to June-July-August. Again, the auto-correlation is broken in the late summer monsoon season.
Figure 11. (a) Hovmoeller diagram in January, February and March of the zonal mean difference between the 200 and 850 hPa wind covariance patterns shown in Figs. 8 and 9, computed in the sector 40°-to-110°E. (b) Hovmoeller diagram of wind shear anomaly \( (\bar{U}_{200} - \bar{U}_{850}) \) from the NCEP/NCAR re-analysis, zonally averaged between 40° and 110°E. Anomalies are for the years 1882 to 1988 and are filtered using a three-month running mean. (c) as in (b), but for the period 1988-1994. Contours every 0.5 m s\(^{-1}\) in (a), every 1.5 m s\(^{-1}\) in (b) and (c); negative values dashed, zero-contour omitted.
TABLE 2. **TIME-LAG CORRELATION COEFFICIENTS BETWEEN:**

(a) **THE SWS INDEX IN JFM AND THE TWS INDEX (AFTER WY92) DURING THE SAME WINTER AND THE FOLLOWING SPRING AND SUMMER**.

(b) **THE TWS INDEX IN JFM AND THE SAME INDEX IN THE FOLLOWING SEASONS**

<table>
<thead>
<tr>
<th></th>
<th>TWS index (DJF)</th>
<th>TWS index (FMA)</th>
<th>TWS index (AMJ)</th>
<th>TWS index (MJJ)</th>
<th>TWS index (JJA)</th>
<th>TWS index (JAS)</th>
</tr>
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<tbody>
<tr>
<td>SWS index (JFM) (a)</td>
<td>0.97</td>
<td>0.96</td>
<td>0.87</td>
<td>0.69</td>
<td>0.64</td>
<td>−0.08</td>
</tr>
<tr>
<td>TWS index (JFM) (b)</td>
<td>−</td>
<td>−</td>
<td>0.86</td>
<td>0.82</td>
<td>0.67</td>
<td>−0.21</td>
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On the basis of these results, it seems possible that some indications on the development of the monsoon circulation in the following summer may be available even at the beginning of winter, at least in years with strong SST anomalies. If GCMs were able to predict correctly the value of the SWS and TWS indices during the winter, then such forecasts could be used to predict the strength of the Asian summer monsoon, taking the intra-seasonal persistence of wind anomalies into account.

The ensemble predictions of the TWS index in JFM were verified using the re-analysis data. In Fig. 12(a), the solid line shows the TWS index as computed from the NCEP re-analysis (for consistency with previous diagrams in this section; however, estimates from ECMWF re-analyses give almost identical results). The dashed line denotes the TWS index as simulated by the ensemble-means of the ECMWF winter ensembles. The large-scale tropical index appears strongly predictable: the correlation between the two curves is 81%. The correlation between the model and re-analysis is still significant (65% and 62% respectively) when the model index in JFM and the re-analysis index in the following spring (Fig. 12(b)) and summer (Fig. 12(c)) are compared. These results highlight once more the persistent character of the large-scale circulation anomaly which determines the evolution the Asian summer monsoon system.

7. **DISCUSSION**

In this study, we have investigated the variability of Eurasian snow depth as represented by ensemble simulations of the wintertime circulation made with the ECMWF GCM. An EOF analysis has been performed on the model snow-depth anomalies at the end of winter, and the associated principal components have been used to compute the covariance patterns between the leading snow-depth anomaly in March and SST, geopotential and wind anomalies during the preceding winter. Indices of reproducibility based on variance ratios, and correlation coefficients between projections of observed and modelled anomalies on covariance patterns have been used to quantify the predictability of the snow-depth field and of the associated circulation anomalies.

The main results of our analysis can be summarized as follows:

- The leading EOF of Eurasian snow depth in late-winter (March) has a zonally-oriented dipole structure, with a band of positive anomalies covering northern Europe and Siberia, and negative anomalies over central Europe, the Himalayas and north China.
- The positive phase of the snow-depth EOF1 is associated with a wintertime circulation characterized by a strengthening of the westerly winds over Europe and Siberia; in the upper troposphere, this westerly anomaly is accompanied by negative zonal wind anomalies over Eurasia around 30°–40°N and positive zonal wind anomalies between the equator and 25°N over Africa and south-east Asia.
A significant relationship has been found between the positive/negative phase of the snow-depth and circulation anomalies described above and warm/cold ENSO events. This sensitivity to boundary forcing is also reflected in the good degree of predictability shown by snow-related circulation anomalies. For anomalies in 500-hPa height, 850-hPa zonal wind and 200-hPa zonal wind, the interannual variations of the ensemble-mean fields show a correlation of 48%, 56% and 65% (respectively) with the corresponding re-analysis anomalies over the eastern half (0° to 180°E) of the northern hemisphere. The reproducibility of the snow-depth anomalies (as a function of SST) is comparable to that of the wind anomalies.

The tropical component of the zonal wind anomaly associated with the snow-depth EOF1 persists from winter to the early summer, and affects the large-scale monsoon circulation over south Asia (as quantified by the zonal-wind-shear index defined by Webster and Yang 1992). Indeed, the wintertime value of such an index simulated by the ensemble means has a correlation of 65% and 62% with the observed wind-shear index in the early (April-May-June) and middle (June-July-August) parts of the following monsoon season.

Overall, the results of our study are consistent with the existence of a negative correlation between Eurasian snow depth and Asian monsoon circulation, which has
been advocated by many investigators in observational and modelling studies. However, two important questions arise on the nature of such a relationship.

The first question regards the relative role of snow depth and SST anomalies in determining the persistence of the circulation anomaly over Eurasia. As far as winter is concerned, it is quite natural to assume that the EOF1-like snow anomaly is the consequence of the intensification of the westerly winds over Siberia; in turn, this wind anomaly is significantly affected by SST anomalies in the tropical Pacific. Since a direct influence of summertime El Niño conditions on the monsoon has been demonstrated by modelling studies (e.g. Palmer et al. 1992; Ju and Slingo 1995; Ferranti et al. 1997), one might be tempted to conclude that the snow-monsoon relationship is just an artefact of the influence of slowly-evolving El Niño anomalies on both the wintertime and the summertime circulation.

However, an active role of the snow anomaly in causing the persistence of the Eurasian wind anomaly cannot be ruled out a priori. In addition to radiative and thermodynamic effects, dynamical feedbacks may also be advocated to explain such a role. A large-scale, zonally-oriented snow anomaly can alter the meridional thermal gradient in the lower troposphere, and therefore shift the location of the areas of maximum baroclinicity (e.g. Walland and Simmonds 1997) during spring. A modified distribution of baroclinic eddies may in turn affect the upper-tropospheric momentum transport, justifying the presence of the strongly persistent signal at 200 hPa. An indication that the snow-related wind anomaly over Eurasia can persist from winter to summer independently from the Pacific SST anomaly comes from the case of 1994. In 1994 (see Fig. 11(c)), the negative phase of the snow–wind covariance pattern developed during winter and indeed evolved into a stronger-than-average monsoon circulation despite the presence of a weakly-positive ENSO index (e.g. Soman and Slingo 1997).

The second important issue is the relationship between the regional aspects of both the snow distribution and the monsoon rainfall. Previous observational studies (e.g. Hahn and Shukla 1976; Dickson 1984) have been mainly focussed on the existence of a negative correlation between snow cover in the Himalayan-Tibetan region and Indian summer rainfall. Here, however, an association has been found between a weak monsoon circulation (according to the Webster-Yang index) and a late-winter snow-depth anomaly which has positive values over Siberia but negative values over Tibet. This may not be so contradictory as it seems. As shown in Ju and Slingo (1995), apart from those years with strong El Niño conditions during summer, there is only a marginal correlation between the Webster–Yang index and the all-Indian rainfall (AIR). The two indices were actually of opposite sign between 1983 and 1986, just when the snow anomalies show a large amplitude in the ECMWF winter ensembles. Therefore, our results are still consistent with a negative correlation between AIR and Himalayan snow cover, at least for those periods in which AIR does not follow the behaviour of the Webster–Yang index. Only in years when a strong El Niño anomaly is present during summer the circulation anomaly may become so strong to overcome the regional effects and cause a spatially-consistent rainfall anomaly throughout south Asia.

The length of the ECMWF seasonal integrations analysed in this study (four months) does not allow us to prove (or disprove) such conjectures. However, it is intriguing that a number of contradictory aspects found in observational studies on monsoon rainfall might be explained by recognizing that the sign of the leading Eurasian snow anomaly is not coherent over the whole continent, as assumed in some modelling studies. The diagnostic study presented here has actually served as the basis for planning a further set of numerical experiments, which span a full seasonal cycle (from winter to the following autumn) and in which the role of land-surface and SST anomalies can be
assessed independently. Some preliminary results of such an experimental programme, which are supportive of an independent role of snow anomalies, have been presented by Ferranti and Molteni (1999).

ACKNOWLEDGEMENTS

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