

Experimental reconstruction of large-scale summer monsoon drought over India and the Tibetan Plateau using tree rings from 'High Asia'

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ABSTRACT

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We investigate the use of a network of 194 tree-ring chronologies from the 'High Asia' region of the Himalaya, Karakoram, Tien Shan, and Tibetan Plateau for the reconstruction of large-scale summer monsoon drought variability over India and the Tibetan Plateau. This experiment was carried out to see if the High Asia tree-ring network contains large-scale information related to the north-south climate gradient over this region, which is known to be associated with the development of the Indian summer monsoon. The climate field chosen for experimental reconstruction was a 735-point grid of self-calibrated Palmer Drought Severity Indices (PDSI) calculated from CRU TS 2.1 gridded temperature and precipitation data. This PDSI grid was reconstructed for the summer monsoon season from a subset of tree-ring chronologies covering the period 1600-1989 using a reduced-space multivariate regression method, and both calibration and verification tests indicate that the reconstructions contain significant hindcast skill in substantial sub-regions of the grid, especially over India. Principal components analysis carried out on the almost 400 years of data over the complete grid indicates that the leading spatial mode or EOF of summer monsoon PDSI expresses a dominant north-south alternation in drought and wetness between India and the Tibetan Plateau, a leading mode also present in the instrumental data. Wavelet analysis of the time series expression of this leading EOF also reveals a very stable 40-60 year pattern of multidecadal variability in it. This time scale of variability appears to be related to slowly changing tropical ocean sea surface temperatures. Together, these results validate the use of a geographically restricted set of tree-ring chronologies over High Asia for the reconstruction of large-scale summer monsoon variability over India and the Tibetan Plateau.

Key-words—High Asia tree rings, India, Tibetan Plateau, Summer monsoon, Drought reconstruction, ENSO.

भारत तथा तिब्बती पठार में 'उच्च एशिया' से प्राप्त वृक्ष-वलियों का प्रयोग करते हुए
बड़े पैमाने पर ग्रीष्म मानसून सूखे का प्रायोगिक पुनर्निर्माण

एडवर्ड आर. कुक एवं पाल जे. क्रुसिक

सारांश

हम भारत एवं तिब्बती पठार में बड़े पैमाने पर ग्रीष्म मानसून सूखे के पुनर्निर्माण हेतु हिमालय, कराकोरम, टिन शान एवं तिब्बती पठार के 'उच्च एशिया' क्षेत्र से प्राप्त 194 वृक्ष-वलय कालानुक्रमों के एक तंत्र का प्रयोग करते हुए अन्वेषण करते हैं। यह अनुप्रयोग यह देखने को किया गया कि इस क्षेत्र में उत्तर-दक्षिण

जलवायु प्रवणता से संबंधित उच्च एशिया वृक्ष-वलय तंत्र में बड़े पैमाने पर जानकारी सन्निहित है, जो कि भारतीय ग्रीष्म मानसून के विकास से संबंध मानी जाती है। प्रायोगिक पुनर्संरचना हेतु चुना गया जलवायु क्षेत्र सी.आर.यू.टी.एस. 2.1 ग्रिड तापमान व अवक्षेपण आधार तथ्य से परिकल्पित स्व-अंशशोधित पामर सूखा उग्रता अक्षांक (पी.डी.एस.आई.) का एक 735-बिंदु ग्रिड था। पी.डी.एस.आई. ग्रिड वृक्ष-वलय कालानुक्रमणों के उपसमुच्चय से ग्रीष्म मानसून ऋतु हेतु लघुकृत-स्थान बहुचर वक्रगति विधि का प्रयोग करते हुए 1600-1989 की अवधि तक पुनर्संरचित किया गया तथा अंशशोधन व सत्यापन दोनों परीक्षण संकेत देते हैं कि विशेषतः भारत में ग्रिड के मजबूत उप-प्रदेशों में पुनर्संरचना सार्थक पश्चदलान दक्षता निहित है। समूचे ग्रिड पर लगभग 400 वर्षों के तथ्यों के आधार पर किए गए मुख्य संघटक विश्लेषण इंगित करते हैं कि अग्रणी स्थानिक मोड अथवा ग्रीष्म मानसून पी.डी.एस.आई. का ई.ओ.एफ. भारत एवं तिब्बती पठार के मध्य सूखा व नमी (आर्द्रता) में प्रभुत्वी उत्तर-दक्षिणी परिवर्तन व्यक्त करता है, उपकरणिय आधार तथ्य में भी एक अग्रणी मोड (प्रावस्था) मौजूद है। इस अग्रणी ई.ओ.एफ. की समय श्रृंखला अभिव्यक्ति का तंत्रगत विश्लेषण इस बहुदशकीय परिवर्तनीयता का एक अत्यंत स्थायी 40-60 वर्ष प्ररूप उद्घाटित करता है। परिवर्तनीयता का यह समय मान धीमे बदलते उष्णकटिबंधीय महासागरी समुद्र पृष्ठीय तापमानों से संबंधित प्रतीत होता है। इसके साथ, भारत एवं तिब्बती पठार में बड़े पैमाने पर ग्रीष्म मानसून परिवर्तनीयता की पुनर्संरचना हेतु उच्च एशिया में वृक्ष-वलय कालानुक्रमणों के भौगोलीय रूप से प्रतिबंधित समुच्चय के प्रयोग से परिणाम अभिपुष्ट करता है।

मुख्य शब्द—उच्च एशिया वृक्ष वलय, भारत, तिब्बती पठार, ग्रीष्म मानसून, सूखा पुनर्निर्माण, ई.एन.एस.ओ.।

INTRODUCTION

THE importance of the summer monsoon to India and its agricultural economy cannot be overstated (Krishna Kumar *et al.*, 2004; Mall *et al.*, 2006). Yet, to this day it is not as well understood as it should be, nor has there been great progress in the long range prediction of Indian summer monsoon rainfall (Munot & Krishna Kumar, 2007). This has been the case even though India has some of the longest rainfall and temperature records on the Asian continent (Hingane *et al.*, 1985; Rupa Kumar *et al.*, 1992; Sontakke *et al.*, 1993; Parthasarathy *et al.*, 1995) for developing and testing forecast models of Indian summer monsoon rainfall. Even so, a large variety of empirical associations have been identified between various leading (i.e. pre-monsoon) climate variables and the Asian monsoon, especially for forecasting all-India summer monsoon rainfall (AISMR) (Parthasarathy *et al.*, 1988; Pant & Rupa Kumar, 1997; Sahai *et al.*, 2003). These empirical associations with AISMR can be roughly assigned geographically to links with the tropical Pacific El Niño/Southern Oscillation (ENSO) region (Pant & Parthasarathy, 1981; Rasmusson & Carpenter, 1983; Parthasarathy & Pant, 1985), links with the Indian Ocean (Saji *et al.*, 1999; Webster *et al.*, 1999; Clark *et al.*, 2000; Gadgil *et al.*, 2003), and links with Asian land surface thermal conditions (Hahn & Shukla, 1976; Yanai *et al.*, 1992; Sankar-Rao *et al.*, 1996). See also Meehl and Arblaster (2002) for a multivariate analysis of these three coupled regions as they affect monsoon development over the more general Asian monsoon system.

Many of the identified empirical associations used to predict the AISMR have not proved to be stable through time (Krishna Kumar *et al.*, 1999a, b; Liu & Yanai, 2002), a problem that has existed since the early days of Indian monsoon forecasting (Savur, 1931). This instability could be due to transient associations or statistical artifacts. It may also reflect true 'epochal changes' in the association between the Indian summer monsoon and large-scale processes like ENSO (Krishna Kumar *et al.*, 1999a) through a natural modulation of the linkage caused by more slowly varying interdecadal variability in the

Pacific Ocean (Torrence & Webster, 1999; Krishnamurthy & Goswami, 2000; Krishnan & Sugi, 2003). There is even a reasonable chance that the loss of predictive skill of certain large-scale associations could be reflecting a true change towards a new climate state caused by 20th Century global warming (Krishna Kumar *et al.*, 1999b). Differentiating between these various 'causes' of lost predictive skill is an important scientific challenge that is hampered by the shortness of the available climate data.

The possibility that Indian summer monsoon rainfall is being modulated by interdecadal processes with return times as long as 50-60 years (Krishnamurthy & Goswami, 2000) means that even the long instrumental climate records from India are not sufficient to study the long-term properties of such variability. For that reason, we must resort to paleoclimate proxies of past monsoon variability to provide enough realizations of such variability for study. High-resolution sediment records have been used successfully to investigate monsoon variability over the past several centuries (Anderson *et al.*, 2002), but the annual time resolution and precise dating control of tree-ring chronologies make them nearly ideal as land-based archives of Indian summer monsoon variability if they are located in monsoon-sensitive regions. This is a problem for the core summer monsoon region of lowland India south of the Himalaya. Only a few tree-ring chronologies exist there, mainly teak (*Tectona grandis*) (Yadav & Bhattacharyya, 1996; Wood, 1996; Borgaonkar *et al.*, 2001; Sikder, 2003; Shah & Bhattacharyya, 2005; Shah *et al.*, 2007), but they are rarely much longer than the instrumental climate records (the AISMR series begins as early as 1844; Sontakke *et al.*, 1993). However, in the 'High Asia' region of the Himalaya, Karakoram, Tien Shan, and Tibetan Plateau, there is an abundant supply of long annual tree-ring chronologies, many of which extend back 400 or more years into the past. These chronologies have been developed and used by many tree-ring scientists to reconstruct relatively local temperature and precipitation histories at various locations in India (Pant & Borgaonkar, 1984; Bhattacharyya *et al.*, 1988; Hughes, 1992; Borgaonkar *et al.*, 1994, 1996; Yadav *et al.*, 1997; Bhattacharyya & Yadav,

1999; Chaudhary *et al.*, 1999; Pant *et al.*, 2000; Yadav & Park, 2000; Yadav & Singh, 2002; Bhattacharyya & Chaudhary, 2003; Singh & Yadav, 2005; Singh *et al.*, 2006; Shah *et al.*, 2007), Nepal (Bhattacharyya *et al.*, 1992; Cook *et al.*, 2003), Bhutan (Buckley *et al.*, 2005), Tibetan Plateau (Braeuning, 2001; Zhang *et al.*, 2003; Braeuning & Mantwill, 2004; Sheppard *et al.*, 2004; Gou *et al.*, 2007), and Pakistan/west central Asia (Esper *et al.*, 1995; Esper *et al.*, 2002, 2003). This list of published results is not intended to be complete. Rather, it serves as a general acknowledgement to the large amount of tree-ring research that has been conducted in High Asia to date.

As indicated earlier, the link between Indian summer monsoon rainfall, climate conditions over the Himalaya and Tibetan Plateau, and the thermal contrast between the regions north and south of the Himalaya is well established now (Fu & Fletcher, 1985; Li & Yanai, 1996; Liu & Yanai, 2001, 2002; Meehl & Arblaster, 2002; Bansod *et al.*, 2003; Feng & Hu, 2005; Kennett & Toumi, 2005). Therefore, the existing network of long tree-ring chronologies distributed over High Asia may be useful for reconstructing long-term, large-scale, hydroclimatic variability over India even in the absence of any tree-ring chronologies from lowland India. To date, no attempt has been made to collectively use the High Asia tree-ring network to reconstruct large-scale summer monsoon variability over the region. It is the purpose of this paper to investigate that potential through the experiment described below.

The Hydroclimate Metric

Prior to reconstructing summer monsoon variability over the the Indian Subcontinent and bordering areas of South Asia and the Tibetan Plateau, the decision was made to use the Palmer Drought Severity Index (PDSI; Palmer, 1965) as the metric of hydroclimatic variability to use. This choice was motivated by a number of factors. First, it was necessary to have a gridded climate field for reconstruction that covered the complete region of interest in such a way that it would hopefully contain information about the link between Indian summer monsoon variability and climatic conditions over the Tibetan Plateau. The choice made here was to use the CRU TS 2.1 interpolated data set of monthly temperature and precipitation available for global land areas on a 0.5 degree grid (New *et al.*, 2002; Mitchell *et al.*, 2004). The monthly data cover the period 1901–2002 everywhere, but in data sparse areas and periods the values ‘relax’ towards climatology making them less reliable. This is a particular problem in western China and the Tibetan Plateau, where very little climate data exist prior to 1950. In India south of the Himalaya, the quality of the pre-1950 gridded climate data is likely to be much better because the quantity of data available for gridding is much higher.

While it would have been possible to attempt a reconstruction of summer monsoon rainfall from the CRU TS 2.1 data directly from tree rings, the decision was made to base

the reconstructions on the Palmer Drought Severity Index (PDSI; Palmer 1965), a well-known and widely used measure of drought and wetness. Previous efforts at reconstructing large-scale PDSI from tree rings have been highly successful (Cook *et al.*, 1999, 2004; Zhang *et al.*, 2004; Brewer *et al.*, 2006; Nicault *et al.*, 2006), and there is little reason in principle to expect it not to work here as well. That being the case, the original PDSI algorithm was not configured with the statistical properties of monsoon climates in mind. This could conceivably cause problems in its calculation over the study region here. To reduce that possibility, the PDSIs were calculated from the CRU TS 2.1 data using an improved ‘self-calibrating’ algorithm described by Wells *et al.* (2004). Dr Gerard van der Schrier of the Royal Netherlands Meteorological Institute performed these calculations and kindly supplied the PDSI data for this study. See van der Schrier *et al.* (2006) for methodological details. Doing so produced a 0.5 degree grid of monthly self-calibrated PDSI over India and the Tibetan Plateau that served as the basis for the calibration data set used for our tree-ring reconstruction experiment.

The PDSI grid used here is shown in Fig. 1. For computational efficiency, the grid spacing was coarsened to 1.0 degree resolution, resulting in 735 grid points of coverage over India and bordering areas of South Asia and the Tibetan Plateau. At this point, the decision was made to use June–July–August (JJA) as the definition of the summer monsoon season and seasonalize the PDSI accordingly. September was viewed as being too late to materially affect tree growth in the High Asia region.

The High Asia Tree-Ring Network

The High Asia tree-ring network available for use here is comprised of 194 annual tree-ring chronologies shown in Fig. 1. It is based on collections made over the years by the Lamont Tree-Ring Lab in Nepal, Bhutan, and China and on many kind contributions from other tree-ring scientists active in the Asian monsoon region as well (see the **Introduction** for a partial review of that work). Of the 194 chronologies, a subset of 65 series covering the common period 1600–1990 was selected as candidate predictors of gridded June–August PDSI. The choice of 1600–1990 for calibration and reconstruction had the unfortunate effect of deleting all of the available tree-ring chronologies from northwest India (see Fig. 1). However, the overall lack of quality climate data prior to 1950 over the Himalaya and Tibetan Plateau necessitated using an end date that provided enough quality data for calibration purposes. The 1600 date was also chosen to enable the development of long enough reconstructions to look for multidecadal variability in a statistically robust way.

Prior to use for reconstruction, the 65 retained tree-ring series were prewhitened to remove positive autocorrelation or persistence that is commonly described as ‘red noise’. This was accomplished using low-order autoregressive (AR) models fitted by the minimum Akaike Information Criterion

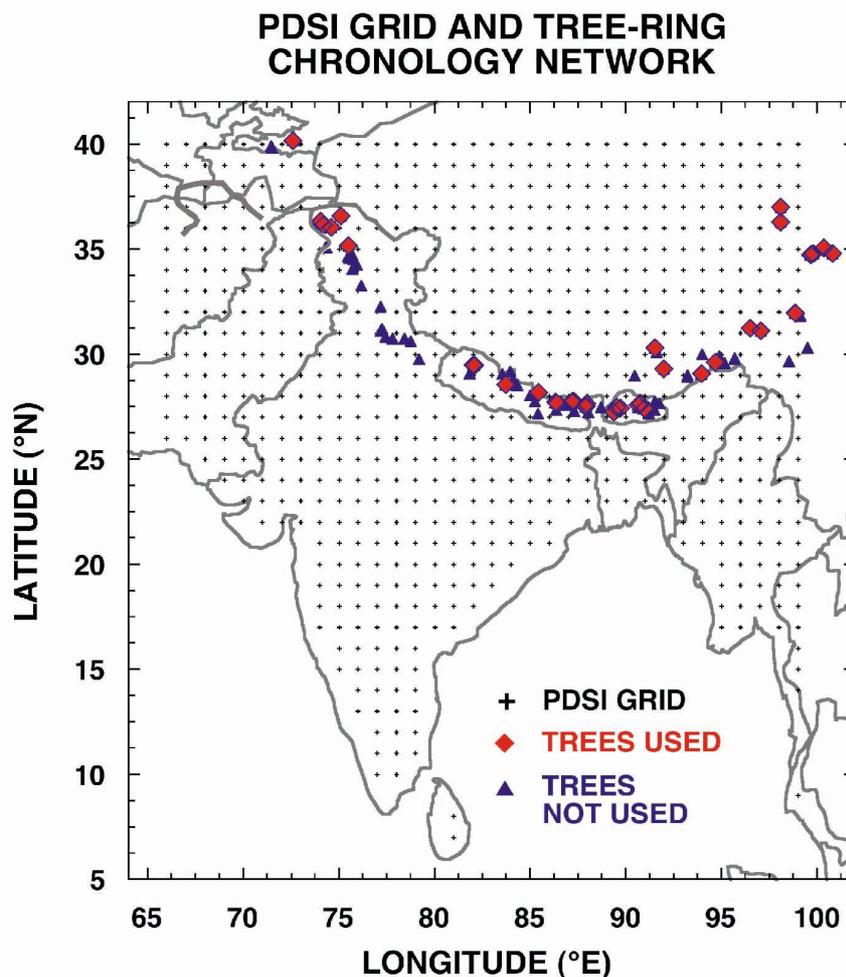


Fig. 1—Map of India and High Asia showing the 735 $1^\circ \times 1^\circ$ grid points of monsoon season (June-August) PDSI (+) and the 194 chronology tree-ring network (▲) used for reconstruction of the PDSI field. The 44 tree-ring chronology subset (◆) ultimately used in the reconstruction models after screening is indicated as well.

(AIC) method (Akaike, 1974), which objectively selects the best order AR model that explains the observed persistence in the series. Doing so produced a set of serially random, or ‘white noise’, series, consistent with each fitted AR model, which simplified the testing of statistical associations between tree rings and PDSI because no correction for loss of degrees of freedom due to autocorrelation in the time series was necessary. Following the method described in Cook *et al.* (1999), after prewhitening the tree-ring series were lagged on themselves one year to produce two candidate predictors (t , $t+1$) for each chronology. This is generally necessary because some tree species discretely encode climate information from one year into the growth ring of the following year.

The Method Of PDSI Reconstruction

Reconstructing spatial patterns of PDSI from tree rings can be done using a variety of statistical methods. Cook *et al.* (1999, 2004) pioneered the use of ‘point-by-point regression’

(PPR), which involves the sequential fitting of single point reconstruction models over the grid. This method works well when there is a reasonable match between the spatial distributions of PDSI and tree rings and the tree-ring network is also reasonably dense. As can be clearly seen in Fig. 1, such is not the case here. For this reason, only large-scale patterns of drought will be reconstructed here, with an emphasis on reconstructing some aspect of the gradient between the high Tibetan Plateau and lowland India. To accomplish this, a simple, but powerful, method of reduced space multivariate regression called ‘Orthogonal Spatial Regression’ (OSR; Briffa *et al.*, 1986; Cook *et al.*, 1994) was used.

Generically speaking, OSR begins by applying principal components analysis (PCA) separately to the predictand field of climate data and the predictor field of tree rings over a specified calibration period. The correlation matrix is typically chosen in each case for this purpose. Each field has its own

NUMBER [†]	EIGENVALUE	% VARIANCE	CUMULATIVE %	#TR p<0.20 [‡]	NPC [*]	R ²
1	5.687	14.582	14.582	40	7	0.725
2	3.936	10.092	24.674	25	7	0.651
3	3.509	8.997	33.671	27	5	0.693
4	2.856	7.323	40.995	27	5	0.581
5	2.391	6.131	47.125	41	7	0.683
6	1.780	4.563	51.689	29	4	0.582
7	1.669	4.279	55.968	26	6	0.695
8	1.411	3.618	59.586	23	4	0.579
9	1.254	3.215	62.800	35	7	0.723
10	1.170	3.000	65.800	30	1	0.349
11	1.086	2.784	68.584	29	6	0.589

[†]EOFs of PDSI predictands with eigenvalues greater than 1.0

[‡]Number of t or t+1 tree-ring variables (p<0.20) used as predictors

^{*}Number of tree-ring PCs entered as predictors in minimum AIC model

Fig. 2—The OSR calibration results. Eleven EOFs of PDSI with eigenvalues >1.0 were retained for reconstruction. They cumulatively explained 68.6% of the total variance. The number of tree-ring chronologies used to reconstruct each EOF ranges from 23 to 41 and the respective explained variance (R²) of each model ranges from 35 to 73%. The range in the number of chronologies used reflects the predictor variables screening applied independently to each model.

characteristic set of empirical orthogonal functions (EOFs) that reflect orthogonal modes of spatial variability within each data set. These are sorted by their eigenvalues from highest to lowest variance accounted for. If there is a statistically significant association between the EOFs of climate and tree rings, this indicates that the trees contain spatial climate information that may be reconstructed back in time. In practice, the tests of association between the climate and tree-ring EOFs are accomplished through comparisons of their principal components (PCs), which express the temporal importance of each EOF in the data over the calibration period. Once the PCs of climate have been regressed on the PCs of tree rings, it is a straightforward process to take the regression-weighted climate estimates from tree rings in 'EOF space' and back-transform them to 'climate space' to produce estimates of climate for each year at each grid point. See Briffa *et al.* (1986) and Cook *et al.* (1994) for details.

EOF analysis was first applied to the 735 grid points of monsoon season PDSI over the chosen 1950-1989 calibration period, the period of highest data quality over the domain. This was done after the PDSI grid point data were also prewhitened with low-order AR models as done for the tree rings. Given that the number of variables (m=735) is greater than the number observations (n=40), there are only n-1 or 39 EOFs with positive eigenvalues. Of those, the first eleven EOFs with eigenvalues exceeding 1.0 were retained for reconstruction. Cumulatively, they explained 68.6% of the total variance in the PDSI field (see Fig. 2). On the tree-ring side of OSR, traditionally only one set of tree-ring EOFs based on all of the available data have been used to reconstruct each

of the retained climate EOFs (Briffa *et al.*, 1986; Cook *et al.*, 1994). While this has worked well in the past, experience in reconstructing PDSI from tree rings using PPR indicates that some level of predictor variable screening can improve the results (Cook *et al.*, 1999). Consequently, each EOF of PDSI was reconstructed using a varying subset of the 65 available tree-ring chronologies (a total of 130 t and t+1 prewhitened candidate variables), with each subset being based on those tree-ring series that correlated best with each PDSI PC at the 80% significance level. This somewhat liberal screening criterion was a compromise between retaining enough chronologies to capture the spatial details of the PDSI EOFs and eliminating those chronologies that were not likely to have any meaningful information about large-scale PDSI variability over the region. As it turned out, only 44 of the 65 candidate tree-ring chronologies were used in varying numbers as actual predictors of the eleven retained PDSI EOFs. Those chronologies are indicated by red diamonds in Fig. 1. This illustrates the relatively high level of 'noise' in the High Asia network for this purpose, i.e. those chronologies not clearly related (p<0.20) to large-scale summer monsoon drought variability accounted for almost one-third of the total. However, the chronologies used for reconstruction still cover the geographic range of the complete High Asia tree-ring network, which indicates that the large-scale summer monsoon signal in the retained tree-ring network is spatially general.

Fig. 2 shows the OSR results based on what has been described. Note that the number of tree-ring variables used to reconstruct each PDSI EOF varied considerably across the eleven modes. This illustrates the importance of pre-calibration

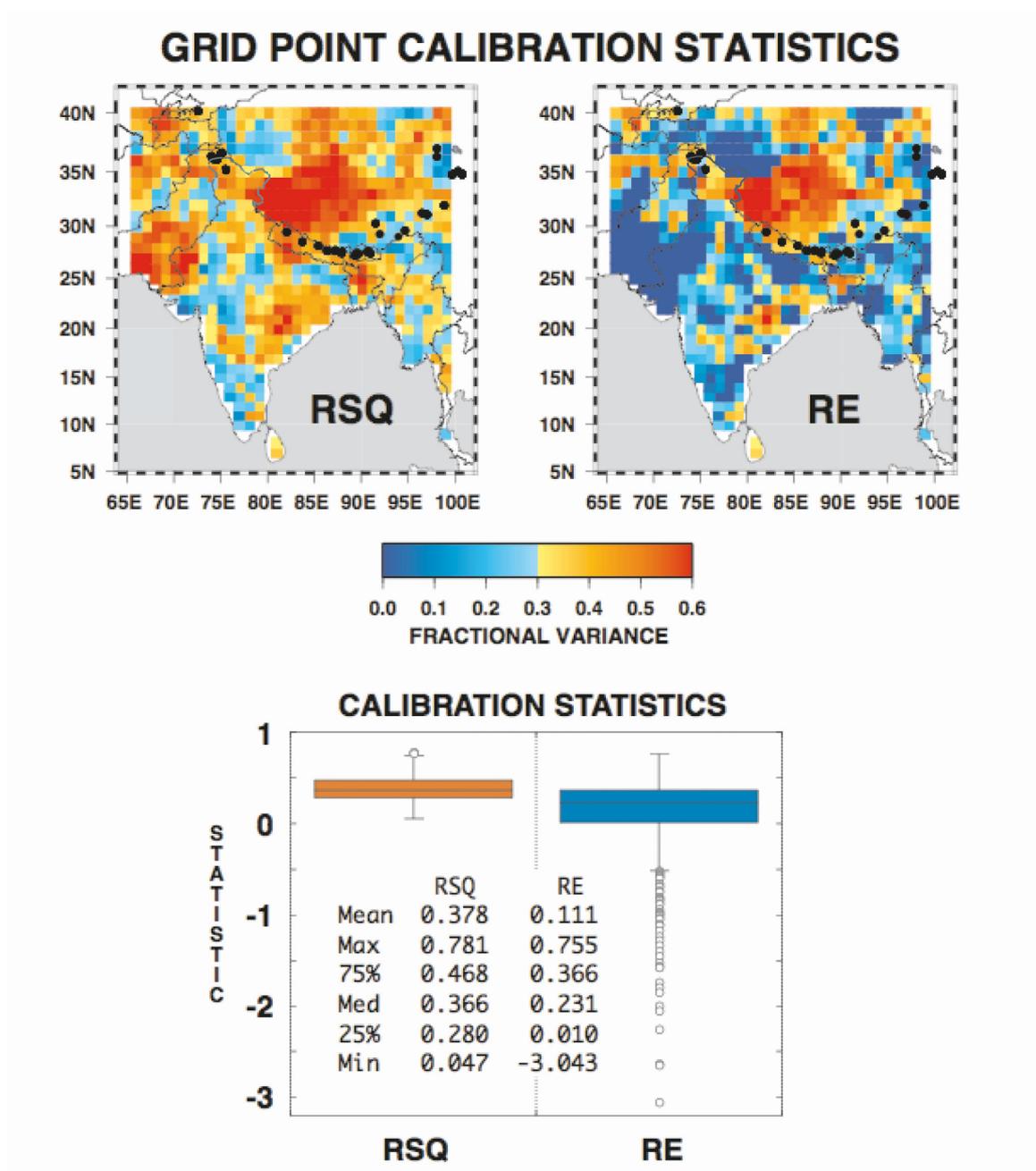


Fig. 3—The calibration period R^2 and RE statistics based on the reconstructions of eleven EOFs of PDSI over India and the Tibetan Plateau. The top maps show the spatial distribution of these statistics. Any $RE < 0$ is regarded as indicated some hindcast skill in the reconstructions. The box plots show some distributional properties of the statistics shown in the maps.

screening because different mixtures of tree rings are sensitive to different EOFs of summer monsoon PDSI. The number of tree-ring PCs used in each EOF model, determined by the minimum AIC criterion, also varied somewhat. However, all but one EOF model required at least four tree-ring PCs to satisfy the minimum AIC criterion. This probably reflects both the spatial complexity of the PDSI EOFs and the broad east-west, crescent-like, spatial structure of the tree-ring network. The

explained variance due to regression (R^2) for each EOF was in some ways surprisingly high (36–73% with 10 of 11 above 57%) given the clear mismatch between the climate and tree ring fields. While some of the R^2 is undoubtedly due to overfitting (Rencher & Pun, 1980), it still suggests that the spatially restricted High Asia tree-ring network contains significant large-scale summer monsoon climate information over India and the Tibetan Plateau for the reasons cited earlier.

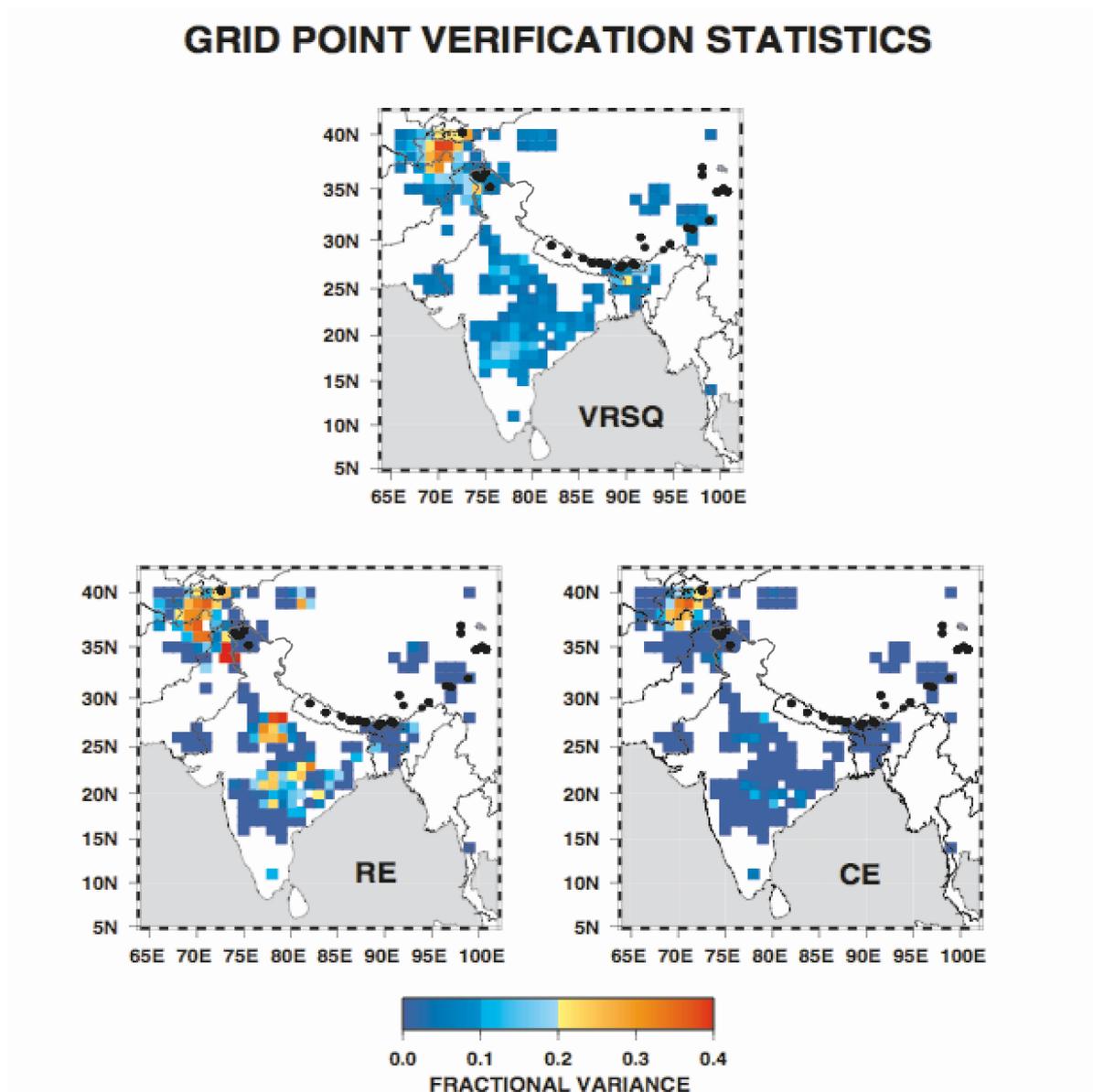


Fig. 4—Maps showing the distribution of three verification statistics: the square of the Pearson correlation (VRSQ), the reduction of error (RE), and the coefficient of efficiency (CE). VRSQ is the easiest test to pass, followed by RE and then CE. The statistics are plotted wherever VRSQ is significant ($p < 0.10$, 1-tailed test). RE and CE are regarded as significant if either is positive.

The PDSI Reconstructions

After the the PDSI EOFs were estimated using OSR, the regression-weighted tree-ring PCs were extended back to 1600 and back-transformed into ‘climate space’ as a linear combination of the eleven calibrated PDSI EOFs following the procedure described in Briffa *et al.* (1986) and Cook *et al.* (1994). At this point, the 735 grid point reconstructions, still based on prewhitened data, had the modeled instrumental data persistence added back in to recover that aspect of PDSI variability over the domain. See Cook *et al.* (1999) for procedural details. Once this was done, the individual grid

point reconstructions were recalibrated against the instrumental data over the 1950–89 period and verified over the 1901–1949 period. The latter was done with full realization that the pre-1950 verification results might be difficult to interpret in certain data-poor areas like the Tibetan Plateau.

Fig. 3 shows two recalibration maps for the full 735-point grid, one for R^2 and the other for the reduction of error (RE) (Cook *et al.*, 1999). Ordinarily, $RE = R^2$ in the calibration period. In the case of OSR, this only applies to ‘EOF space’, not ‘climate space’. Hence, RE provides a somewhat different and generally more rigorous indication of how the tree-rings used in OSR

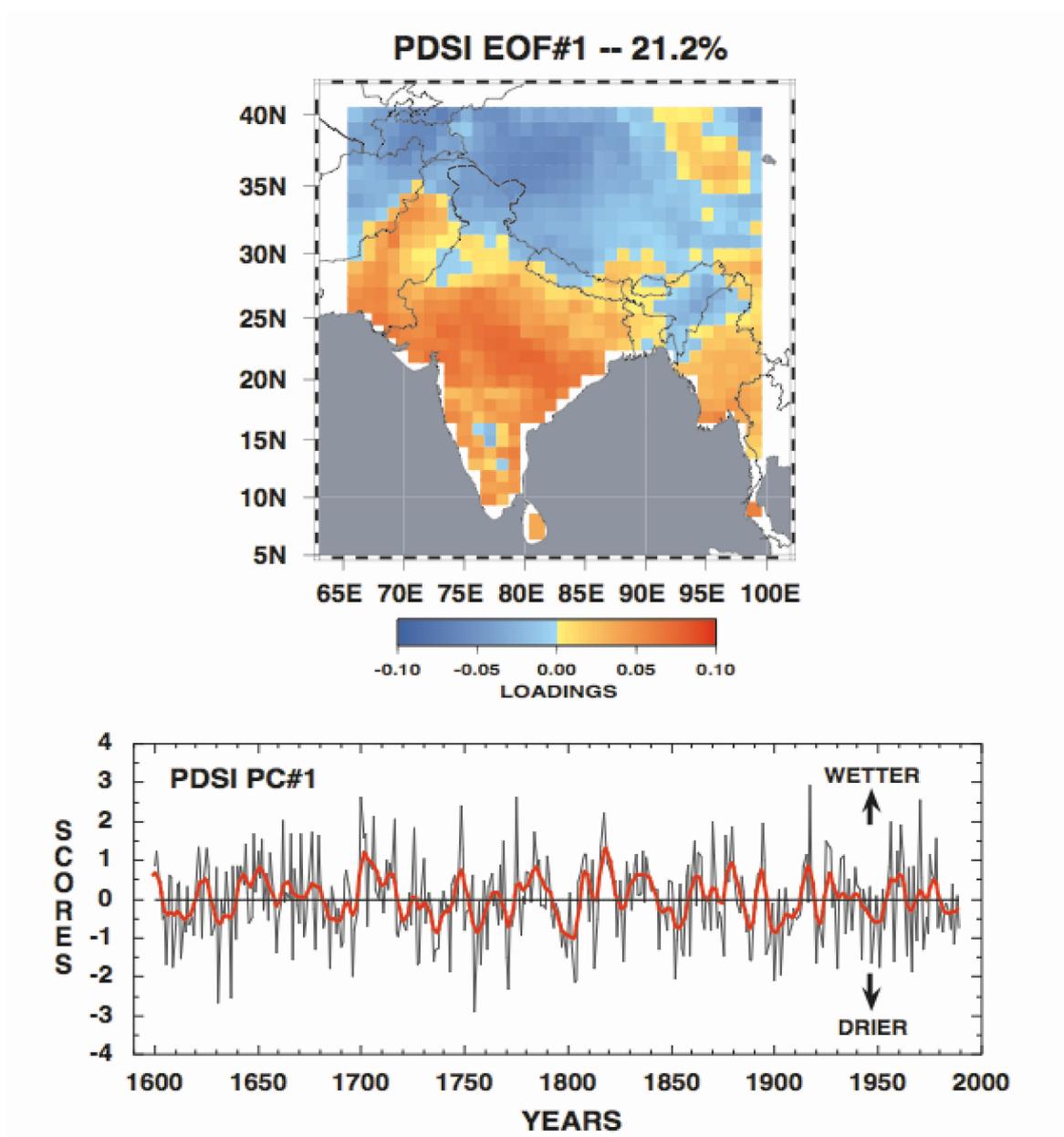


Fig. 5—The leading EOF of the tree-ring reconstructions of summer monsoon PDSI (top map) and its corresponding PC (bottom time series). The EOF loadings and PC scores are signed to indicate wetter (drier) conditions in India when positive (negative) and vice versa for the Tibetan Plateau.

have done in the calibration period after back-transformation, with $RE=0$ indicative of no meaningful model skill. The difference between R^2 and RE is especially apparent in the southern half of Pakistan and western India, where R^2 is highly significant, but RE near zero or strongly negative. Beneath the maps are box plots that provide some distributional statistics of the R^2 and RE statistics in the maps. The median R^2 and RE are 0.37 and 0.23. Each is indicative of a significant amount of calibrated model skill over the domain. In fact, the

lower hinge, or 25th quantile, of the RE distribution is slightly above zero, which indicates that 75% of the grid point estimates have some level of model skill. These results again indicate that the High Asia tree-ring network contains useful information about the summer monsoon over India and the Tibetan Plateau. However, to feel more confident that this is truly the case, some level of reconstruction verification in the 1901-49 pre-calibration period is desirable.

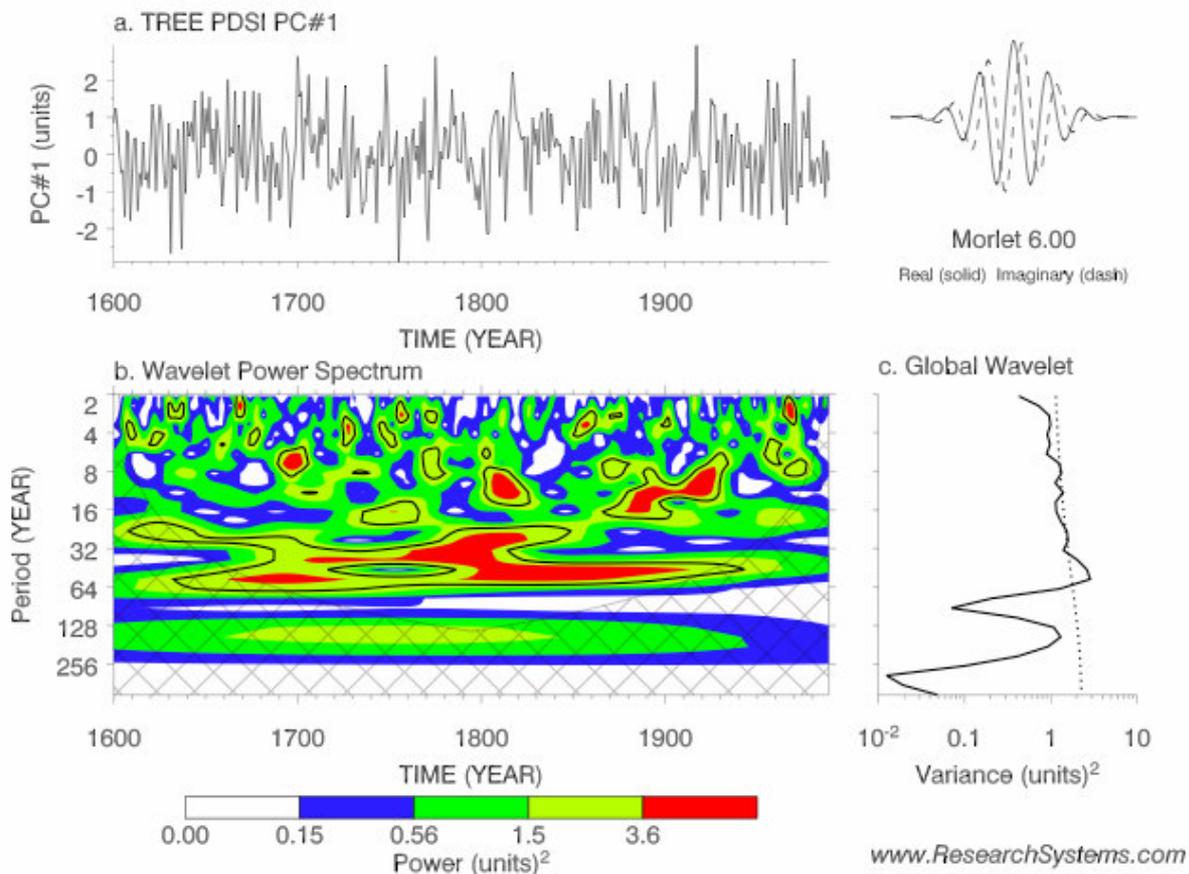


Fig. 6—The wavelet spectrum of the PCs of the leading EOF mode in Fig. 4. The heavy black lines in the wavelet spectrum plot indicate intervals and periods where band-limited variance is statistically significant ($p < 0.10$).

Fig. 4 shows maps of three verification statistics for the 1901–49 period: the square of the Pearson correlation (VRSQ), the RE (as before), and the coefficient of efficiency (CE; Nash & Sutcliffe, 1971). The relative merits of each statistic as a measure of model verification are discussed in Cook *et al.* (1994, 1999). The significance of VRSQ is testable in a theoretical sense, while only positive values of RE and CE indicate of some level of hindcast skill, with CE being the most difficult to pass. The values shown in each map correspond to only those grid points where VRSQ was significant at the 90% level or better (1-tailed test). These results are decidedly weaker than the calibration R^2 and RE statistics in Fig. 3. This result is not too surprising, but it is difficult to know how much of the indicated decline in fidelity was caused by tree rings or reduced instrumental data quality prior to 1950. Certainly, the lack of verification over the Tibetan Plateau and most of western China must be partly due to the diminished quantity of instrumental data available for use in the CRU TS 2.1 interpolated estimates. South of the Himalaya, where the quantity of the pre-1950 climate data is much greater, the generally positive VRSQ results for much of central India are

highly encouraging. The RE and CE results are noticeably and predictably weaker, but there is still some level of skill (RE or $CE > 0$) indicated at some grid points. Thus, the reconstructions of past summer monsoon drought over India appear to contain some useful information based on independent time period verification statistics. For the Tibetan Plateau, only the calibration period results can be realistically used for guidance.

Large-scale PDSI variability

The generally positive PDSI reconstruction results described above suggest that information about summer monsoon drought variability over India and the Tibetan Plateau can be examined back to 1600 now. However, given the encouraging, but somewhat modest, verification period results, it is probably unwise to do this at the grid point level. Also, given the nature of this experiment in determining the degree to which the High Asia tree rings contain large-scale monsoon signals related to the gradient between the regions north and south of the Himalaya, only this aspect will be investigated in the reconstructions for now.

PC#1 CORRELATIONS WITH KAPLAN SST
June-August Monsoon Season
1950-89, 1901-49, 1856-1900

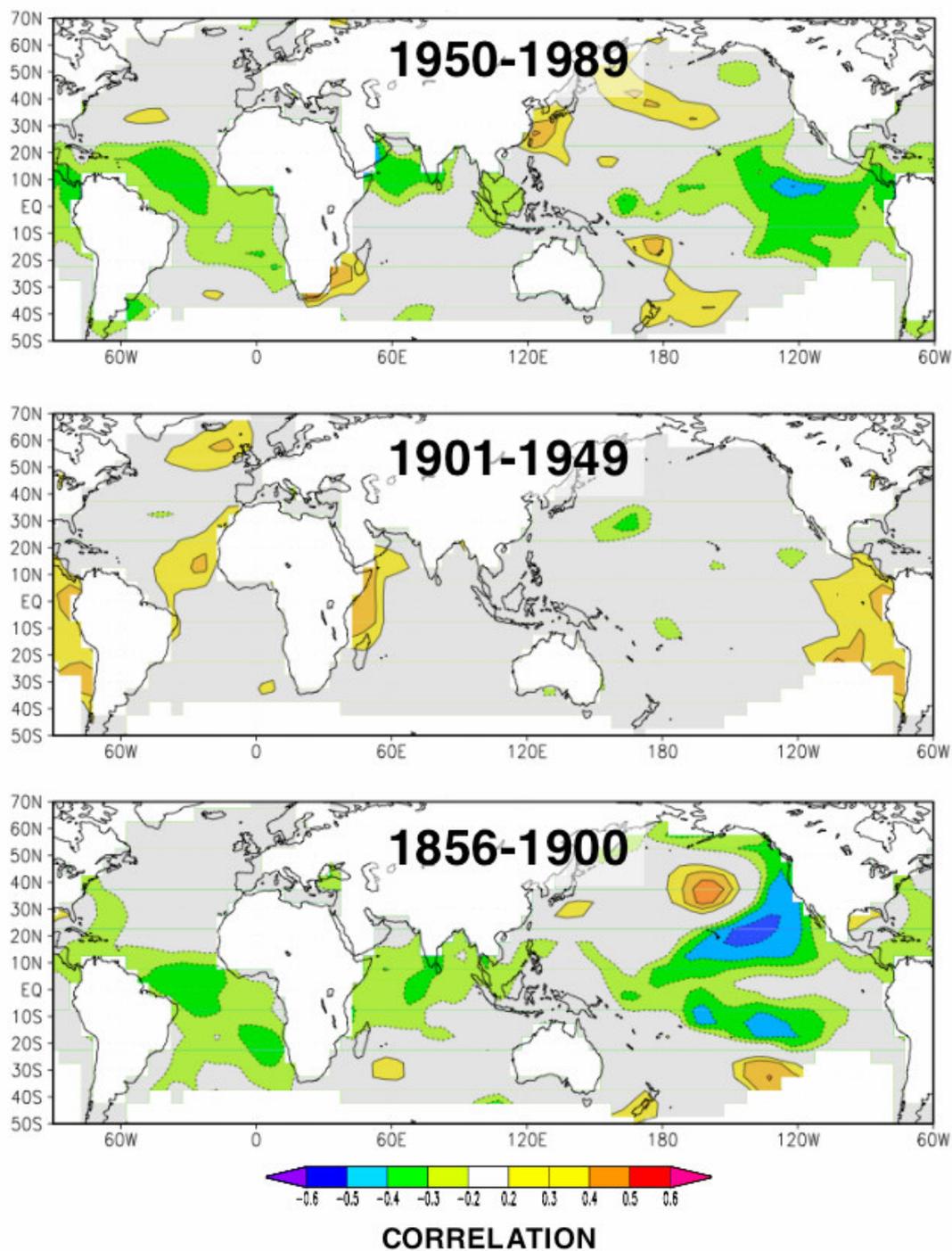


Fig. 7—Maps of correlations between the leading PC and Kaplan SSTs for three time periods: the calibration period (1950-1989), the verification period (1901-1949), and an independent period that precedes those periods used in the reconstruction phase of the study (1856-1900). Note the evidence for a ca. 50 year oscillation in the significance of the correlations with SSTs between periods, primarily in the tropical oceans.

To this end, PCA was applied to the correlation matrix of the 735 grid point PDSI reconstructions over the period 1600–1989. Fig. 5. shows the map of loadings for the leading EOF and its PC time series. The map reveals a remarkable north-south alternation in drought variability over India and the Tibetan Plateau, which accounts for 21.2% of the total variance in the field. This contrast is also present in the leading EOFs of the instrumental and reconstructed PDSI data over the 1950–89 calibration period (not shown); the latter map has a coefficient of congruence (Richman, 1986; Cook *et al.*, 1999) of 0.75 with the former, a highly significant ($p < 0.01$) result based on monte carlo significance testing. Interestingly, the north-south alternation in the reconstructions is more stable over time than that in the instrumental data over the 1901–1989 period. After applying PCA to the instrumental and reconstructed PDSIs for the 1901–49 verification period, the leading EOFs for that period were compared to those in their respective data sets based on the calibration period. The results were very informative. The leading EOFs of the reconstructions for the two time periods had a congruency of 0.90 with each other, a highly stable result that was not preordained by the reconstruction method used. In contrast, the congruence between the leading EOFs of the instrumental data was only 0.21 between time periods! In fact, the highest level of congruence found was 0.63 (0.60) between EOF#1 (EOF#2) of the calibration period and EOF#2 (EOF#1) of the verification period. The presence of this mixture of spatial structure in the instrumental data is most likely due to changing data quality between the periods.

The leading PC in Fig. 5 was subjected to wavelet analysis (Torrence & Compo, 1998) to determine the degree to which there have been persistent modes of band-limited variability over time in this large-scale mode. Fig. 6 shows the results (calculated at <http://atoc.colorado.edu/research/wavelets/>). It reveals the presence of a reasonably stable multidecadal oscillation ($p < 0.10$) in the 40–60 year range over the past 400 years and a more episodic interannual pattern of variability at around 8 years. The multitaper spectrum (Mann & Lees, 1996) of the same series (not shown) indicates similar, but more sharply defined, results with significant ($p < 0.10$) spectral peaks of 56.8 and 7.6 years. (See the wavelet ‘Frequently Asked Questions’ (FAQ) website <http://atoc.colorado.edu/research/wavelets/faq.html#bias> for why the global wavelet is not a good estimator of the overall power spectrum.) The 40–60 year variability indicated here is consistent with the 50–60 year time scale of variability described by Krishnamurthy and Goswami (2000) for the modulation of the Indian summer monsoon by low-frequency ENSO variability. Their analyses were based on a maximum of 125 years of instrumental data only. Our results based on almost 400 years of data suggest that this modulation is a long-term feature of the Asian monsoon system. The 7.6-year term also suggests a possible link with the North Atlantic Oscillation (NAO) and Eurasian

snow cover through its affects the development of the summer monsoon over India. Interesting, neither the wavelet nor the multitaper spectra contain much evidence for the 12–20 year variability like that described by Torrence and Webster (1999) in their variance and wavelet coherency analyses or the similar time scale variability described by Krishnan and Sugi (2003). This shorter-term interdecadal variability could conceivably be present in one of the higher-order EOFs not examined here.

An oceanic cause of the multidecadal modulation indicated in Fig. 6 is indicated by three maps showing correlations between our leading PC and summer monsoon season SSTs (Fig. 7). These maps were produced using Geert Jan van Oldenborgh’s KNMI Climate Explorer web program (<http://climexp.knmi.nl/>) and are based on Kaplan *et al.* (1998) SSTs. What is striking about these maps is the presence of significant correlations ($p < 0.10$) in the tropical oceans in the 1950–1989 calibration period, an almost a complete loss of significance in the 1901–1949 verification period, and a return to broad-scale significance in the tropical oceans in the 1856–1900 period. This pattern of episodic correlation on an approximate 50-year time scale is remarkably similar to the time scale of band-limited variability indicated in the wavelet and multitaper spectra. It suggests that the findings of Krishnamurthy and Goswami (2000) may be related to the alteration of climate between India and the Tibetan Plateau revealed here. Our results also indicate that this variability is not just related to ENSO and the Indian Ocean SSTs, but also to SST variability in the Atlantic Ocean. Torrence and Webster (1999) also describe a conspicuous period of low variance in both NINO3 sea surface temperatures (SSTs) in the tropical Pacific and AISMR (Torrence & Webster, 1999; their Fig. 5) during the period when the loss of correlation between our leading PC and SSTs occurs. So there may yet be a connection between our results and those of Torrence and Webster (1999).

CONCLUSIONS

An experimental reconstruction of large-scale drought over India and the Tibetan Plateau over the past 400 years was carried out to see if a tree-ring network in a geographically restricted area of High Asia could be used to reconstruct summer monsoon PDSI over the region. The results are highly encouraging and suggest that the continued development of the tree-ring network there is justified. Indeed, one of the greatest limitations revealed by our experiment is the limited quality of the pre-1950 instrumental climate data north of the Himalaya for model verification and EOF analyses, which makes the value of tree-ring reconstructions there even greater. Nonetheless, even with the somewhat limited ability to verify the grid point reconstructions north of the Himalaya prior to 1950, there is reasonable evidence to support the validity of our tree-ring reconstructions over the region at the broad spatial scale evaluated here.

The leading EOF of reconstructed PDSI, which is probably the most reliable mode, expresses a clear alternation in wetness and dryness north and south of the Himalaya, and its PC contains a stable 40-60 year mode of band-limited multidecadal variability. This finding is consistent with previous results based on analyses of instrumental data over the past 125 years (Krishnamurthy & Goswami, 2000) and supports its longer-term existence as a modulator of the Indian summer monsoon over the past 400 years. Provisionally, the source of this modulation appears to come from slowly changing tropical ocean SSTs.

The current challenge is to extend this work to include long tree-ring chronologies in the core region of the summer monsoon in lowland India. Doing so would enable a more comprehensive reconstruction of monsoon variability there at smaller spatial scales. The current tree-ring network in lowland India is not yet sufficient either in number or in length for that purpose, but the results shown here indicate that a concerted effort to develop such a network is likely to yield great new insights into the properties and causes of summer monsoon variability over India and surrounding regions.

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REFERENCES

- Akaike H 1974. A new look at the statistical model identification. *IEEE Transactions on Automatic Control* AC-19: 716-723.
- Anderson DM, Overpeck JT & Gupta AK 2002. Increase in the Asian Southwest Monsoon during the past four centuries. *Science* 297: 596-599.
- Bansod SD, Yin Z, Lin Z & Zhang X 2003. Thermal field over Tibetan Plateau and Indian summer monsoon rainfall. *International Journal of Climatology* 23: 1589-1605.
- Bhattacharyya A, LaMarche VC & Telewski FW 1988. Dendrochronological reconnaissance of the conifers of northwest India. *Tree-Ring Bulletin* 48: 21-30.
- Bhattacharyya A, LaMarche VC & Hughes MK 1992. Tree-ring chronologies from Nepal. *Tree-Ring Bulletin* 52: 59-66.
- Bhattacharyya A & Yadav RR 1999. Climatic reconstructions using tree-ring data from tropical and temperate regions of India - A review. *IAWA Journal* 20: 311-316.
- Bhattacharyya A & Chaudhary V 2003. Late-summer temperature reconstruction of the Eastern Himalayan region based on tree-ring data of *Abies densa*. *Arctic, Antarctic, and Alpine Research* 35: 196-202.
- Borgaonkar HP, Pant GP & Rupa Kumar K 1994. Dendroclimatic reconstruction of summer precipitation of Srinagar, Kashmir, India since the late 18th Century. *The Holocene* 4: 299-306.
- Borgaonkar HP, Pant GP & Rupa Kumar K 1996. Ring-width variations in *Cedrus deodara* and its climatic response over the western Himalaya. *International Journal of Climatology* 16: 1409-1422.
- Borgaonkar HP, Rupa Kumar, K, Sikder AB, Ram S & Pant GB 2001. Tree-ring analysis of *Tectona grandis* (Teak) from central India. Poster presented at Tree Rings and People: An International Conference on the Future of Dendrochronology, Davos, Switzerland, September 22-26, 2001.
- Braeuning A 2001. Climate history of Tibetan Plateau during the last 1000 years derived from a network of juniper chronologies. *Dendrochronologia* 19: 127-137.
- Braeuning A & Mantwill B 2004. Summer temperature and summer monsoon history on the Tibetan Plateau during the last 400 years recorded by tree rings. *Geophysical Research Letters*, Vol. 31, L24205, DOI 10.1029/2004GL020793.
- Brewer S, Alleaume S, Guiot J & Nicault A 2006. Historical droughts in Mediterranean regions during the last 500 years: a data/model approach. *Climate of the Past Discussions* 2: 771-800.
- Briffa KR, Jones PD, Wigley TML, Pilcher JR & Baillie MGL 1986. Climate reconstruction from tree rings: Part 2, Spatial reconstruction of summer mean sea-level pressure patterns over Great Britain. *Journal of Climatology* 6: 1-15.
- Buckley BM, Cook BI, Bhattacharyya A, Dukpa D & Chaudhary V 2005. Global surface temperature signals in pine ring-width chronologies from southern monsoon Asia. *Geophysical Research Letters*, Vol. 32, L20704, doi:10.1029/2005GL023745.
- Chaudhary V, Bhattacharyya A & Yadav RR 1999. Tree-ring studies in the eastern Himalayan region: prospects and problems. *IAWA* 20: 317-324.
- Clark CO, Cole JE & Webster PJ 2000. Indian Ocean SST and Indian summer rainfall: predictive relationships and their decadal variability. *Journal of Climate* 13: 2503-2519.
- Cook ER, Briffa KR & Jones PD 1994. Spatial regression methods in dendroclimatology: A review and comparison of two techniques. *International Journal of Climatology* 14: 379-402.
- Cook ER, Meko DM, Stahle DW & Cleaveland MK 1999. Drought reconstructions for the continental United States. *Journal of Climate* 12: 1145-1162.
- Cook ER, Woodhouse CA, Eakin CM, Meko DM & Stahle DW 2004. Long-term aridity changes in the western United States. *Science* 306: 1015-1018.
- Cook ER, Krusic PJ & Jones PD 2003. Dendroclimatic signals in long tree-ring chronologies from the Himalayas of Nepal. *International Journal of Climatology* 23: 707-732.
- Esper J, Bosshard A, Schweingruber FH & Winiger M 1995. Tree-rings from the upper timberline in the Karakorum as climatic indicators for the last 1000 years. *Dendrochronologia* 13: 79-88.
- Esper J, Schweingruber FH & Winiger M 2002. 1300 years of climatic history for western central Asia inferred from tree rings. *The Holocene* 12: 267-277.
- Esper J, Shiyatov SG, Mazepa VS, Wilson RJS, Graybill DA & Funkhouser G 2003. Temperature-sensitive Tien Shan tree ring chronologies show multi-centennial growth trends. *Climate Dynamics* 21: 699-706.
- Feng S & Hu Q 2005. Regulation of Tibetan Plateau heating on variation of Indian summer monsoon in the last two millennia. *Geophysical Research Letters*, Vol. 32, L02702, doi:10.1029/2004GL021246.
- Fu C & Fletcher JO 1985. The relationship between Tibet-tropical ocean thermal contrast and interannual variability of Indian monsoon rainfall. *Journal of Climate and Applied Meteorology* 24: 841-847.
- Gadgil S, Vinayachandran PN & Francis PA 2003. Droughts of the Indian summer monsoon: role of clouds over the Indian Ocean. *Current Science* 85: 1713-1719.

- Gou X, Chen F, Cook E, Jacoby G & Yang M 2007. Streamflow variations of the Yellow River over the past 593 years in western China reconstructed from tree rings. *Water Resources Research* Vol. 43: W06434, doi:10.1029/2006WR005705.
- Hahn DG & Shukla J 1976. An apparent relationship between Eurasian snow cover and Indian monsoon rainfall. *Journal of the Atmospheric Sciences* 33: 2461-2462.
- Hingane LS, Rupa Kumar K & Ramana Murty BhV 1985. Long-term trends of surface air temperature in India. *International Journal of Climatology* 5: 521-528.
- Hughes MK 1992. Dendroclimatic evidence from the western Himalaya. In: Bradley RS & Jones PD (Editors)—*Climate Since A.D. 1500*: 415-431. Routledge Press, London.
- Kaplan A, Cane M, Kushnir Y, Clement A, Blumenthal M & Rajagopalan B 1998. Analyses of global sea surface temperature 1856-1991. *Journal of Geophysical Research* 103: 18,567-18,589.
- Kennett EJ & Toumi R 2005. Himalayan rainfall and vorticity generation within the Indian summer monsoon. *Geophysical Research Letters* Vol. 32, L04802, doi:10.1029/2004GL021925.
- Krishna Kumar K, Kleeman R, Cane MA & Rajagopalan B 1999a. Epochal changes in Indian monsoon-ENSO precursors. *Geophysical Research Letters* 26: 75-78.
- Krishna Kumar K, Rajagopalan B & Cane M 1999b. On the weakening relationship between the Indian monsoon and ENSO. *Science* 284: 2156-2159.
- Krishna Kumar K, Rupa Kumar K, Ashrit RG, Deshpande NR & Hansen JW 2004. Climate impacts on Indian agriculture. *International Journal of Climatology* 24: 1375-1393.
- Krishnamurthy V & Goswami BN 2000. Indian monsoon-ENSO relationship on interdecadal timescale. *Journal of Climate* 13: 579-595.
- Krishnan R & Sugi M 2003. Pacific decadal oscillation and variability of the Indian summer monsoon rainfall. *Climate Dynamics* 21: 233-242.
- Li C & Yanai M 1996. The onset and interannual variability of the Asian summer monsoon in relation to land-sea thermal contrast. *Journal of Climate* 9: 358-375.
- Liu X & Yanai M 2001. Relationship between the Indian monsoon rainfall and the tropospheric temperature over the Eurasian continent. *Quarterly Journal of the Royal Meteorological Society* 127: 909-937.
- Liu X & Yanai M 2002. Influence of Eurasian spring snow cover on Asian summer rainfall. *International Journal of Climatology* 22: 1075-1089.
- Mall RK, Singh R, Gupta A, Srinivasan G & Rathore LS 2006. Impact of climate change on Indian agriculture: a review. *Climatic Change* 78: 445-478.
- Mann ME & Lees JM 1996. Robust estimation of background noise and signal detection in climatic time series. *Climatic Change* 33: 409-445.
- Meehl GA & Arblaster JM 2002. The tropospheric biennial oscillation and Asian-Australian monsoon rainfall. *Journal of Climate* 15: 722-744.
- Mitchell TD, Carter TR, Jones PD, Hulme M & New M 2004. A Comprehensive Set of High-Resolution Grids of Monthly Climate for Europe and the Globe: the Observed Record (1901-2000) and 16 Scenarios (2001-2100). Tyndall Working Paper 55, Tyndall Centre, UEA, Norwich, UK.
- Munot AA & Krishna Kumar K 2007. Long range prediction of Indian summer monsoon rainfall. *Journal of Earth System Science* 116: 73-79.
- Nash JE & Sutcliffe JV 1971. Riverflow forecasting through conceptual models 1, A discussion of principles. *Journal of Hydrology* 10: 282-290.
- New M, Lister D, Hulme M & Makin I 2002. A high-resolution data set of surface climate over global land areas. *Climate Research* 21: 1-25.
- Nicault A, Alleaume S, Brewer S & Guiot J 2006. Mediterranean drought fluctuation during the last 650 years based on tree-ring data, *Climate Dynamics* (in review).
- Palmer WC 1965. Meteorological drought. Weather Bureau Research Paper 45, U.S. Department of Commerce, Washington, DC, 58 pp.
- Pant GB & Parthasarathy B 1981. Some aspects of an association between the southern oscillation and Indian summer monsoon. *Meteorology and Atmospheric Physics* 29: 245-251.
- Pant GB & Borgaonkar HP 1984. Growth rate of chir pine (*Pinus roxburghii*) trees in Kumaun area in relation to regional climatology. *Himalayan Research and Development* 3: 1-5.
- Pant GB & Rupa Kumar K 1997. *Climates of South Asia*. John Wiley & Sons, New York. 344 pp.
- Pant GB, Rupa Kumar K, Borgaonkar HP, Okada N, Fujiwara T & Yamashita K 2000. Climatic response of *Cedrus deodara* tree-ring parameters from two sites in the western Himalaya. *Canadian Journal of Forest Research* 30: 1127-1135.
- Parthasarathy B & Pant GB 1985. Seasonal relationship between Indian summer monsoon rainfall and Southern Oscillation. *Journal of Climatology* 5: 369-378.
- Parthasarathy B, Diaz HF & Eischeid JK 1988. Prediction of all-India summer monsoon rainfall with regional and large-scale parameters. *Journal of Geophysical Research* 93: 5341-5350.
- Parthasarathy B, Munot AA & Kothawale DR 1995. All India monthly and seasonal rainfall series: 1871-1993. *Theoretical and Applied Climatology* 49: 217-224.
- Rasmusson EM & Carpenter TH 1983. The relationship between eastern equatorial Pacific sea surface temperatures and rainfall over India and Sri Lanka. *Monthly Weather Review* 111: 517-528.
- Rupa Kumar K, Pant GB, Parthasarathy B & Sontakke NA 1992. Spatial and subseasonal patterns of the long-term trends of Indian summer monsoon rainfall. *International Journal of Climatology* 12: 257-268.
- Rencher AC & Pun FC 1980. Inflation of R^2 in best subset regression. *Technometrics* 22: 49-53.
- Richman MB 1986. Rotation of principal components. *Journal of Climatology* 6: 293-335.
- Sahai AK, Grimm AM, Satyan V & Pant GB 2003. Long-lead prediction of Indian summer monsoon rainfall from global SST evolution. *Climate Dynamics* 20: 855-863.
- Saji NH, Goswami BN, Vinayachandran PN & Yamagata T 1999. A dipole mode in the tropical Indian Ocean. *Nature* 401: 360-363.
- Sankar-Rao M, Lau KM & Yang S 1996. On the relationship between Eurasian snow cover and the Asian summer monsoon. *International Journal of Climatology* 16: 605-616.
- Savur SR 1931. The seasonal forecasting formulae used in the India Meteorological Department. *Scientific Notes* 4: 57-68. Published by India Meteorological Department, New Delhi.
- Shah SK & Bhattacharyya A 2005. Tree-ring analysis of tropical Indian tree with special emphasis on *Tectona grandis* and *Cedrela toona*. Poster presented at PAGES Open Science Meeting, Beijing, China, August 10-12, 2005.
- Shah SK, Bhattacharyya A & Chaudhary V 2007. Reconstruction of June–September precipitation based on tree-ring data of teak (*Tectona grandis* L.) from Hoshangabad, Madhya Pradesh, India. *Dendrochronologia* 25: 57-64.
- Sheppard P, Tarasov P, Graumlich L, Heussner K, Wagner M, Österle H & Thompson L 2004. Annual precipitation since 515BC reconstructed from living and fossil juniper growth of Northeast Qinghai Province, China. *Climate Dynamics* 23: 869-881.
- Sikder A 2003. Tree-Ring analysis : An indicator of Past Environment : A case study of *Tectona grandis* (teak) in Bori forest, Madhya Pradesh, India. M.Sc. thesis, Pune University, Pune, India.
- Singh J & Yadav RR 2005. Spring precipitation variations over the western Himalaya, India, since A.D. 1731 as deduced from tree rings. *Journal of Geophysical Research* 110, D01110, doi:10.1029/2004JD004855.

- Singh J, Park W-K & Yadav RR 2006. Tree-ring based hydrological records for western Himalaya, India since A.D. 1560. *Climate Dynamics* 26: 295-303.
- Sontakke NA, Pant GB & Singh N 1993. Construction of all-India summer monsoon rainfall series for the period 1844-1991. *Journal of Climate* 6: 1807-1811.
- Torrence C & Compo GP 1998. A practical guide to wavelet analysis. *Bulletin of the American Meteorology Society* 79: 61-78.
- Torrence C & Webster PJ 1999. Interdecadal changes in the ENSO-monsoon system. *Journal of Climate* 12: 2679-2690.
- van der Schrier G, Briffa KR, Osborn TJ & Cook ER 2006. Summer moisture availability across North America. *Journal of Geophysical Research* Vol. 111, D11102, doi:10.1029/2005JD006745.
- Webster PJ, Moore AM, Loschnigg JP & Leben RR 1999. Coupled ocean-atmosphere dynamics in the Indian Ocean during 1997-98. *Nature* 401: 356-360.
- Wells N, Goddard S & Hayes MJ 2004. A self-calibrating Palmer Drought Severity Index. *Journal of Climate* 17: 2335-2351.
- Wood ML 1996. Synoptic Dendroclimatology in the Upper Narmada River Basin: An Exploratory Study in central India. M.Sc. thesis, Department of Geosciences, University of Arizona, Tucson.
- Yanai M, Li C & Song Z 1992. Seasonal heating of the Tibetan Plateau and its effects on the evolution of the Asian summer monsoon. *Journal of the Meteorological Society of Japan* 70: 319-351.
- Yadav RR & Bhattacharyya A 1996. Biological inferences from the growth climate relationship in teak from India. *Proceedings of the Indian Academy of Science B* 62: 233-238.
- Yadav RR, Park W-K & Bhattacharyya A 1997. Dendroclimatic reconstruction of April-May temperature fluctuations in the western Himalaya of India since A.D. 1698. *Quaternary Research* 48: 187-191.
- Yadav RR & Park W-K 2000. Precipitation reconstruction using ring-width chronology of Himalayan cedar from western Himalaya: preliminary results. *Proceedings of the Indian Academy of Science (Earth Planet Science)* 109: 339-345.
- Yadav RR & Singh J 2002. Tree-ring-based spring temperature patterns over the past four centuries in western Himalaya. *Quaternary Research* 57: 299-305.
- Zhang QB, Cheng GD, Yao TD, Kang XC & Huang JG 2003. A 2,326-year tree-ring record of climate variability on the northeastern Qinghai-Tibetan Plateau. *Geophysical Research Letters*, Vol. 30, doi:10.1029/2003GL017425.
- Zhang, Z, Mann ME & Cook ER 2004. Alternative methods of proxy-based climate field reconstruction: application to summer drought over the conterminous United States back to AD 1700 from tree-ring data. *The Holocene* 14: 502-516.