GEOPHYSICAL RESEARCH LETTERS, VOL. 34, L16707, doi:10.1029/2007GL030431, 2007

A 900-year (600 to 1500 A.D.) record of the Indian summer monsoon precipitation from the core monsoon zone of India

Ashish Sinha,[1] Kevin G. Cannariato,[2] Lowell D. Stott,[2] Hai Cheng,[3]
R. Lawrence Edwards,[3] Madhusudan G. Yadava,[4] R. Ramesh,[4] and Indra B. Singh[5]

Received 18 April 2007; revised 19 June 2007; accepted 19 July 2007; published 21 August 2007.

[1] We present a near-annually resolved record of the Indian summer monsoon (ISM) rainfall variations for the core monsoon region of India that spans from 600 to 1500 A.D. from a ^{230}Th-dated stalagmite oxygen isotope record from Dandak Cave. Our rainfall reconstruction, which spans the Medieval Warm Period (MWP) and the earliest portion of the Little Ice Age (LIA), indicates that the short instrumental record of ISM underestimates the magnitude of monsoon rainfall variability. Periods of severe drought, lasting decades, occurred during the 14th and mid 15th centuries and coincided with several of India's most devastating famines. **Citation:** Sinha, A., K. G. Cannariato, L. D. Stott, H. Cheng, R. L. Edwards, M. G. Yadava, R. Ramesh, and I. B. Singh (2007), A 900-year (600 to 1500 A.D.) record of the Indian summer monsoon precipitation from the core monsoon zone of India, *Geophys. Res. Lett.*, *34*, L16707, doi:10.1029/2007GL030431.

1. Introduction

[2] The seasonal rainfall brought by the Southwest Indian summer monsoon (ISM) supplies nearly 80% of Southeast Asia's annual precipitation and is vital to sustaining the region's agriculture, which supports nearly a quarter of the world's population. The instrumental record of ISM (~150 years) reveals strong interannual to inter-decadal variability associated with the El Nino Oscillation events (ENSO). As recently as the late 1960s, El Niño related ISM failure for three consecutive years resulted in 1.5 million deaths within India [*Center for Research on the Epidemiology of Disasters*, 2005]. A longer record of ISM variability is necessary to determine whether longer intervals of monsoon failure occurred in the past and assess future risk.

[3] A growing array of evidence indicates that during the past 1500 years centennial-scale natural climate oscillations affected broad areas of the Earth. These include the Little Ice Age (LIA, ca. 1400 to 1850 A.D.) [*Mann*, 2002a] and Medieval Warm Period (MWP, ca. 900 to 1300 A.D.) [*Mann*, 2002b]. These century-scale climate oscillations were not restricted to the high latitudes and appear to have involved a significant perturbation in tropical ocean temperatures as well as monsoon rainfall [*Newton et al.*, 2006]. While pre-instrumental historical accounts of ISM variations exist during this time interval, they document ISM variability mainly in terms of extreme events such as occurrences of droughts, floods, and famines [*Pant et al.*, 1993] and therefore are of limited use in assessing the full frequency spectrum of ISM variability. Consequently, multi-century ISM reconstructions derived from proxy records of precipitation are vital to assess pre-instrumental patterns of ISM variability and examine relationships with other components of the climate system.

[4] Efforts to extend our knowledge of ISM variability beyond the instrumental and historical records have largely focused on marine proxies that reflect the extent of upwelling and wind intensity in Arabian Sea sediments [*Anderson et al.*, 2002; *Gupta et al.*, 2003; *Overpeck et al.*, 1996; *Schulz et al.*, 1998]. However, these ISM reconstructions have several disadvantages for assessing monsoon rainfall variations and its potential impact on human populations. In particular, these marine records do not sample hydrologic conditions within the core monsoon zone of India where most human populations are located and are of lower temporal resolution, so only the longer-term variations in monsoon wind intensity is monitored. It is the amount of rainfall and the inter-regional pattern of monsoon rainfall variability, particularly drought that is of greatest importance to human populations.

[5] Recent work has shown that speleothems can be excellent archives of monsoon rainfall variability [*Burns et al.*, 2002; *Fleitmann et al.*, 2003, 2007; *Neff et al.*, 2001; *Sinha et al.*, 2005; *Yadava and Ramesh*, 2005; *Yadava*, 2002; *Yuan et al.*, 2004]. The oxygen isotopic composition (δ^{18}O) of speleothem calcite from tropical and monsoon locations is primarily controlled by the δ^{18}O value of precipitation, which in turn varies inversely with rainfall amount [*Dansgaard*, 1964] and/or fraction of water vapor removed from maritime air masses as they move away from their source regions [*Rozanski et al.*, 1993].

2. Cave Location and Climatology

[6] The δ^{18}O of speleothems collected from Dandak Cave, (19°00′N, 82°00′E; elevation ~400 m), is an ideal archive of rainfall variability within the core monsoon region of India (Figure 1). First, instrumental rainfall data from the core monsoon region (Figure 1), which is comprised of the following Indian meteorological sub-divisions 6–8, 18–22, 24–27, is strongly correlated with the area weighted "All India Summer Monsoon Rainfall" (AISMR) time series [*Gadgil*, 2003]. Furthermore, the instrumental rainfall data from Jagdalpur, the nearest meteorological

---

[1] Department of Earth Sciences, California State University, Dominguez Hills, Carson, California, USA.

[2] Department of Earth Sciences, University of Southern California, Los Angeles, California, USA.

[3] Department of Geology and Geophysics, University of Minnesota, Minneapolis, Minnesota, USA.

[4] Physical Research Laboratory, Navrangpura, Ahmedabad, India.

[5] Department of Geology, Lucknow University, Lucknow, India.

0094-8276/07/2007GL030431$05.00

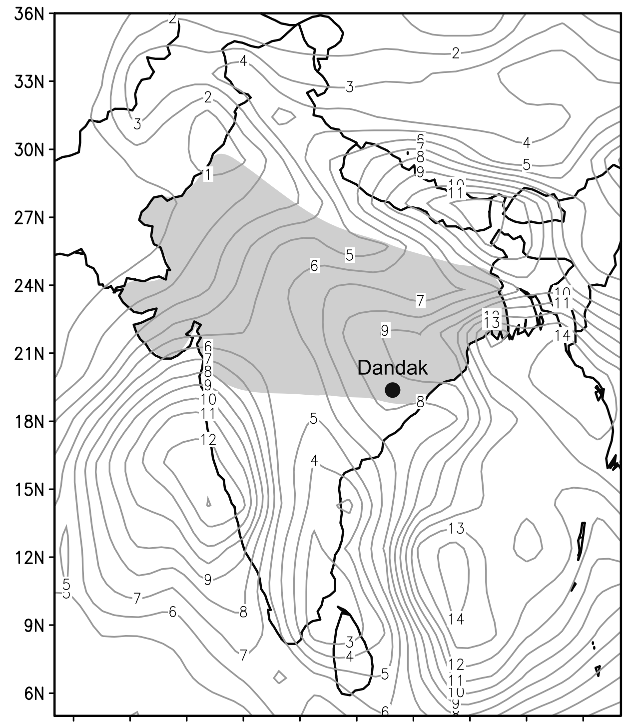

Figure 1. Location of Dandak Cave, Kanger Valley National Park, Chhattisgarh, India. Contours indicate Southwest Indian summer monsoon mean precipitation rate (mm/day) from NCEP/NCAR reanalysis dataset (June to September, 1948 to 2006). Shaded region represents the core monsoon zone of India.

station from Dandak Cave (∼30 km), exhibits a strong correlation with the core monsoon zone rainfall (Figure S1)[1] suggesting that rainfall variations at our site are broadly representative of the core monsoon region. Over the last century, nearly every major sub-decadal length interval of weaker rainfall in the core monsoon zone rainfall is also represented in local rainfall variations (Figure S1).

[7] Summer monsoon precipitation (June through September) accounts for nearly 80% of the annual rainfall (∼1528 mm) at the cave site. Although no long-term rainwater δ^{18}O measurements are available from the cave site, δ^{18}O measurements from the nearby state of Orissa [*Yadava*, 2002] demonstrate that the amount effect is the dominant influence on the isotopic variability of monsoon rainwater and that the magnitude of this effect is consistent with reported values for coastal and island stations worldwide [*Yurtsever and Gat*, 1981]. The rainfall weighted δ^{18}O value of precipitation in this region, estimated from Global Network for Isotopes in Precipitation (GNIP) stations [*Bowen and Revenaugh*, 2003] (http://waterisotopes.org), is between −4.3‰ (total annual precipitation ∼1528 mm) and −3.75‰ (monsoon months precipitation ∼1200 mm) (Vienna standard mean ocean water, VSMOW). These values are in close agreement with δ^{18}O values of local surface waters (lakes, ponds, and rivers) but is 0.5‰ to 1.0‰ more negative than groundwater and cave waters [*Yadava*, 2002] suggesting some evaporative enrichment during the groundwater recharge and cave infiltration process. Non-monsoon rainfall at the cave site comprises only a small portion (∼20%) of the annual rainfall. Post-monsoon rainfall (Oct–Nov, −8.0‰) from short-lived tropical storms and cyclones accounts for 5 to 10% of the annual precipitation. Winter rainfall (Dec–March, ∼1.0‰) results from either isolated westerly disturbances or locally recycled moisture [*Rozanski et al.*, 1993].

[8] Because ∼80% of annual precipitation at Dandak Cave is delivered during the monsoon months with a distinctively negative δ^{18}O value (not considering more negative but a minor contribution from post-monsoon rainfall during Oct–Nov), the temporal variations in the speleothem δ^{18}O record primarily reflects variations in monsoon rainfall intensity via the amount effect. This interpretation rests in part on the assumption that non-monsoonal rainfall in this region has not been a significant source of rainwater. There are other processes that can influence the δ^{18}O of speleothem calcite such as changes in the amount of water evaporation from the soil, or the amount of groundwater infiltration and drip rate. A change in relative humidity and ambient cave temperatures would also influence the stalagmite δ^{18}O values, however these factors would shift the values in the same direction as does the amount effect [*Burns et al.*, 2002; *Fleitmann et al.*, 2003].

3. Stalagmite Sample, Chronology, and Isotopic Analyses

[9] Dandak Cave consists of two chambers connected via a narrow crawl space. A stalagmite (DAN-D, 27 cm long) was collected from the second, poorly ventilated chamber ∼220 m from the cave entrance. This chamber has near constant air temperature (∼25°C) and high relative humidity (>90%) [*Yadava*, 2002], ideal conditions for equilibrium calcite precipitation. The δ^{18}O of modern calcite in Dandak Cave is in close agreement with the expected δ^{18}O value of speleothem calcite (−4.43‰, assuming precipitation of speleothem calcite at 25°C, and the δ^{18}O of drip water = −3.0‰) indicating that calcite precipitation is occurring at or near isotopic equilibrium [*Yadava*, 2002; *Yadava and Ramesh*, 2005]. Additionally, samples analyzed from several growth horizons passed the Hendy test [*Hendy*, 1971] and δ^{18}O and δ^{13}C values of DAN-D stalagmite are poorly correlated, further suggesting equilibrium precipitation condition existed (Figure S2). The stalagmite is composed entirely of crystalline calcite and shows no evidence for any major discontinuity such as clay bands or truncated calcite crystals.

[10] Thirteen ^{230}Th dates were obtained by magnetic sector inductively coupled plasma mass spectrometry (Table S1) (Figure 2) at the University of Minnesota Isotope Laboratory. Samples were collected using a hand-held dental drill. Thorium and uranium isotopes were analyzed separately on a Finnigan-MAT Element outfitted with a double focusing sector-field magnet in reversed Nier-Johnson geometry and a single MasCom multiplier. The instrument was operated in low resolution and in electrostatic peak hopping mode. Combined ionization plus trans-

---

[1] Auxiliary material data sets are available at ftp://ftp.agu.org/apend/gl/2007gl030431. Other auxiliary material files are in the HTML.

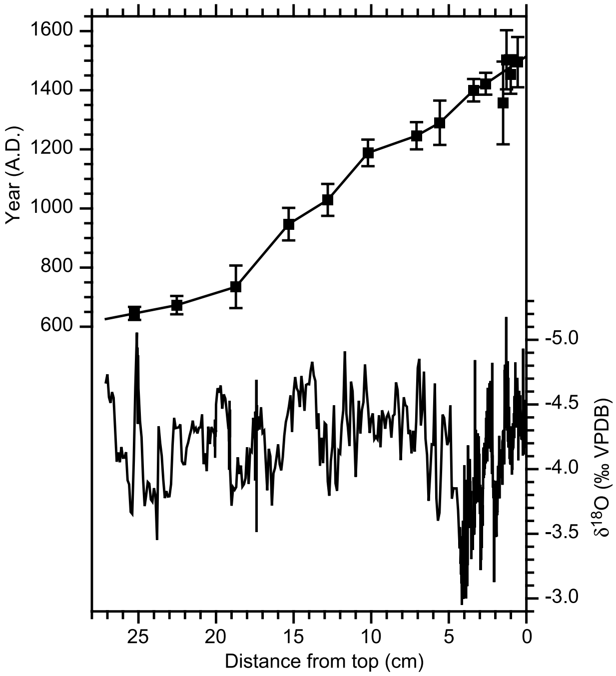

Figure 2. δ^{18}O record, ^{230}Th age dates (black squares with 2sigma error bars), and age model for Dandak Cave stalagmite.

mission efficiency of 2.5 to 3‰ has been measured for uranium and 1.5 to 2‰ has been measured for thorium. An age model was developed by linearly interpolating ages between ^{230}Th dates and indicates the stalagmite grew continuously between from 600 to 1500 A.D. at rates between ~160 and ~970 μm/yr (Figure 2).

[11] Samples for δ^{18}O analyses were milled every ~100 μm along the growth axis. A total of 684 samples were analyzed with an online automated carbonate preparation system linked to a VG Prism II isotope ratio mass spectrometer at the University of Southern California. Before 1350 A.D., every 10th milled sample was analyzed. Whereas, after 1350 A.D., every milled sample was analyzed. This sampling yields a δ^{18}O record with an average sample resolution of 2.7 yr before 1350 A.D. and 0.4 yr after 1350 A.D. However, the actual time represented by each sample may be longer because it was not always possible to follow the curved growth layers while milling.

4. Results and Discussion

[12] Between 600 and 1500 A.D., the speleothem calcite deposited in Dandak cave has δ^{18}O values of −3‰ to −5‰ (Figure 3). Today the δ^{18}O of calcite deposited in this cave and in nearby caves is −4.0‰ to −4.5‰ [*Yadava*, 2002]. For the majority of the Dandak record the speleothem δ^{18}O values are within the range of modern calcite values or only slightly more negative, indicating that ISM variability was within the range of variability represented in the modern instrumental period. However, much of the 14th and mid 15th centuries stands out as an extreme excursion in monsoon rainfall. The δ^{18}O values between 1300 to 1450 A.D. are 1.0 to 1.5‰ higher than modern values [*Yadava*, 2002]. Such an extreme positive excursion in the δ^{18}O values and the persistence of these values for over a century indicates a period of severely reduced monsoon rainfall (Figure 3).

[13] A precise transfer of the Dandak δ^{18}O values to an estimate of rainfall amount is not yet available. However, using a limited seasonal δ^{18}O data base for precipitation in this region we estimate the magnitude of the temporal changes in the monsoon rainfall that are implied by the observed δ^{18}O values [*Yadava*, 2002; *Yadava and Ramesh*, 2005]. Today there is ~ 2.2‰ depletion in the δ^{18}O of precipitation for each 100 mm increase in the monthly rainfall at the Dandak cave site [*Yadava*, 2002]. We therefore estimate up to a 30% reduction in monsoon rainfall, following the procedures outlined by *Yadava* [2002] to explain the 1.5‰ positive isotopic excursion observed in Dandak cave record during the 14th century. A 30% reduction in rainfall at this location has been observed in the modern instrumental period. However, the speleothem record documents a persistent reduction in rainfall lasting for several decades.

[14] Variations in the cave temperature could have imparted a small influence on the speleothem δ^{18}O values and could account for a portion of the anomalously high δ^{18}O values we observe during the earliest portion of the LIA. However, changes in the mean annual temperature at the cave site would have been small given its proximity to the tropical Indian Ocean. Proxy reconstructions of tropical Indian SSTs since the LIA indicate there have been changes of only ~0.6°C [*Damassa et al.*, 2006]. Even an unexpectedly large shift in cave temperature of 1°C would result in a δ^{18}O shift of only 0.25‰ and therefore could not explain the full magnitude of δ^{18}O changes recorded at DAN cave. The larger δ^{18}O changes must therefore reflect substantial changes in the amount of monsoon rainfall via the amount effect.

[15] The multi-decadal period of weakened monsoon in our reconstruction, during the 14 and 15th centuries, would have imposed significant stress on the human societies living in this part of India. There are historic accounts of the major famine and drought that can be compared to our monsoon reconstruction [*Dando*, 1980; *Maharatna*, 1996; *Pant et al.*, 1993]. Indeed, nearly every major famine in India coincided (within the dating uncertainty) with a period of reduced monsoon rainfall as reflected in the Dandak δ^{18}O record (Figure 3). Most of these famines lasted for only a single to several years. However, two famines lasted a decade or longer, including the devastating Durga Devi famine (1396 A.D. to 1409 A.D.) [*Government of Maharashtra*, 1973]. Both of these famines occurred at the beginning of the LIA during the longest duration and most severe ISM weakening of our reconstruction.

[16] There are few high-resolution records of ISM variability available for comparison to our stalagmite δ^{18}O record. ISM reconstructions from Arabian Sea marine sediments [*Agnihotri et al.*, 2002; *Gupta et al.*, 2003; *von Rad et al.*, 1999] stalagmite δ^{18}O records from Oman and Yemen [*Burns et al.*, 2002; *Fleitmann et al.*, 2007] and a pollen record from the western Himalaya [*Phadtare and Pant*, 2006] also indicate a weaker monsoon during the LIA and a relatively stronger monsoon during the MWP. The timing of higher frequency variations in these records is sometimes offset from one another probably because of imprecise chronologies.

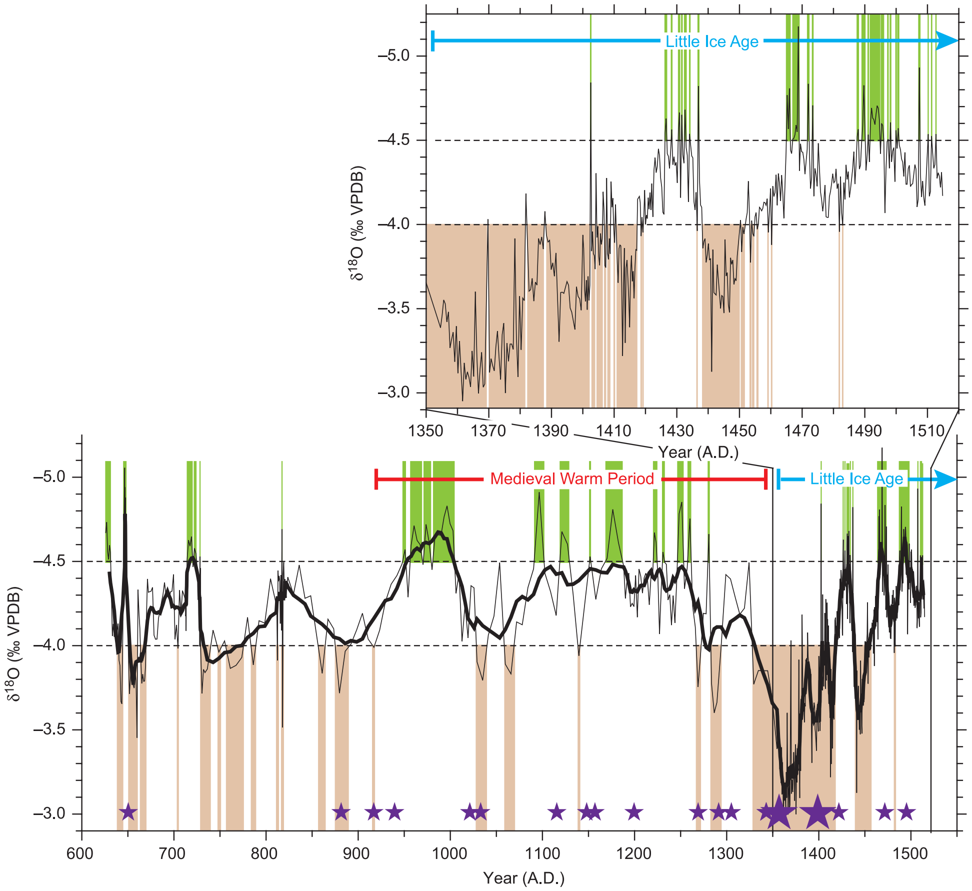

Figure 3. $\delta^{18}O$ record of Dandak Cave stalagmite. Thick line is 9-point running mean. Dashed lines indicate $\delta^{18}O$ range of calcite from Dandak Cave stalagmites deposited during the modern instrumental period. Intervals of inferred stronger monsoon shaded green and intervals of weaker monsoon shaded brown. Purple stars indicate timing of historical famines in India (two large stars indicate decadal length famines).

[17] ISM failures lasting 15 to 30 yr occur every ~150 yr in some climate simulations as extreme events of naturally occurring multidecadal variability associated with the Indian and Pacific Oceans [*Meehl and Hu*, 2006]. A number of recent studies demonstrate a close correspondence between Holocene monsoon variability and proxies for solar activity. Our reconstruction of ISM variability, on its independent chronology, exhibits a close resemblance to the solar proxies (Figure S3) only during the MWP. Such a correlation suggests that a direct solar influence on the location of the Intertropical Convergence Zone [*Newton et al.*, 2006] and thus monsoon precipitation is a plausible explanation for the ISM variations during the LIA and MWP [*Kodera*, 2004].

[18] Alternatively, newer generation GCMs with fully interactive atmospheric chemistry indicate irradiance related variations in tropopause ozone can also explain the MWP/LIA pattern of changes in the hydrological cycle [*Shindell et al.*, 2006]. However, other more complex climate teleconnections may be important. For example, substantial reduction of North Atlantic thermohaline circulation could weaken the Asian monsoons through air-sea interactions [*Zhang and Delworth*, 2005] although there is little evidence for such a decline during the LIA [*Keigwin and Boyle*, 2000]. If the LIA and MWP are related to solar forced changes in the state of the Arctic Oscillation [*Shindell et al.*, 2001], then multidecadal oscillations in North Atlantic SST may control ISM variations [*Goswami et al.*, 2006]. Alternatively, solar forced changes in the state of ENSO may have had a pronounced affect on ISM precipitation [*Emile-Geay et al.*, 2007].

[19] Since the end of the LIA, ca 1850 A.D., the human population in the Indian monsoon region has increased from about 200 million to over 1 billion. A recurrence of weaker interval of ISM comparable to those inferred in our record would have serious implications to human health and economic sustainability in the region. At present, water-resource management as well as agricultural and land-use planning do not account for the possibility of large-sustained drought in India and SE Asia and underscores the importance of refining our knowledge of how various factors influence the behaviour of the ISM.

[20] **Acknowledgments.** We thank M. Rincon for sampling and stable isotope analysis. This work was supported by grants from the National Science Foundation (ATM-0502615 to L.D.S., R.L.E., and H.C.).

References

Agnihotri, R., K. Dutta, R. Bhushan, and B. L. K. Somayajulu (2002), Evidence for solar forcing on the Indian monsoon during the last millennium, *Earth Planet. Sci. Lett.*, *198*, 521–527.

Anderson, D. M., J. T. Overpeck, and A. K. Gupta (2002), Increase in the Asian Southwest Monsoon during the past four centuries, *Science*, *297*, 596–599.

Bowen, G. J., and J. Revenaugh (2003), Interpolating the isotopic composition of modern meteoric precipitation, *Water Resour. Res.*, *39*(10), 1299, doi:10.1029/2003WR002086.

Burns, S. J., D. Fleitmann, M. Mudelsee, U. Neff, A. Matter, and A. Mangini (2002), A 780-year annually resolved record of Indian Ocean monsoon precipitation from a speleothem from south Oman, *J. Geophys. Res.*, *107*(D20), 4434, doi:10.1029/2001JD001281.

Center for Research on the Epidemiology of Disasters (CRED) (2005), EM-DAT: The International Disaster Database, http://www.em-dat.net, Brussels, Belgium.

Damassa, T. D., J. E. Cole, H. R. Barnett, T. R. Ault, and T. R. McClanahan (2006), Enhanced multidecadal climate variability in the seventeenth century from coral isotope records in the western Indian Ocean, *Paleoceanography*, *21*, PA2016, doi:10.1029/2005PA001217.

Dando, W. A. (1980), *The Geography of Famine*, 209 pp., John Wiley, New York.

Dansgaard, W. (1964), Stable isotopes in precipitation, *Tellus*, *16*, 438–468.

Emile-Geay, J., M. Cane, R. Seager, A. Kaplan, and P. Almasi (2007), El Niño as a mediator for the solar influence on climate, *Paleoceanography*, *22*, PA3210, doi:10.1029/2006PA001304.

Fleitmann, D., S. J. Burns, M. Mudelsee, J. Kramers, U. Neff, J. Kramers, A. Mangini, and A. Matter (2003), Holocene forcing of the Indian Monsoon recorded in stalagmite from Southern Oman, *Science*, *300*, 1737–1739.

Fleitmann, D., S. J. Burns, A. Mangini, M. Mudelsee, J. Kramers, U. Neff, A. A. Al-Subbary, A. Buettner, D. Hippler, and A. Matter (2007), Holocene ITCZ and Indian monsoon dynamics recorded in stalagmites from Oman and Yemen (Socotra), *Quat. Sci. Rev.*, *26*, 170–188.

Gadgil, S. (2003), The Indian Monsoon and its variability, *Annu. Rev. Earth Planet. Sci.*, *31*, 429–467.

Goswami, B. N., M. S. Madhusoodanan, C. P. Neema, and D. Sengupta (2006), A physical mechanism for North Atlantic SST influence on the Indian summer monsoon, *Geophys. Res. Lett.*, *33*, L02706, doi:10.1029/2005GL024803.

Government of Maharashtra (1973), Report of the fact finding committee for survey of scarcity areas Maharashtra State, 310 pp., Mumbai, India.

Gupta, A. K., D. M. Anderson, and J. T. Overpeck (2003), Abrupt changes in the Asian southwest monsoon during the Holocene and their links to the North Atlantic Ocean, *Nature*, *421*, 354–356.

Hendy, C. H. (1971), The isotopic geochemistry of speleothems I. The calculation of the different modes of formation on the isotopic composition of speleothems and their applicability as paleoclimatic indicators, *Geochim. Cosmochim. Acta*, *35*, 801–824.

Keigwin, L. D., and E. A. Boyle (2000), Detecting Holocene changes in thermohaline circulation, *Proc. Natl. Acad. Sci. U.S.A.*, *97*, 1343–1346.

Kodera, K. (2004), Solar influence on the Indian Ocean monsoon through dynamical processes, *Geophys. Res. Lett.*, *31*, L24209, doi:10.1029/2004GL020928.

Maharatna, A. (1996), *The Demography of Famines: An Indian Historical Perspective*, 317 pp., Oxford Univ. Press, Delhi, India.

Mann, M. E. (2002a), Little Ice Age, in *Encyclopedia of Global Environmental Change*, edited by M. C. MacCracken and J. S. Perry, pp. 504–509, John Wiley, Chichester, U. K.

Mann, M. E. (2002b), Medieval climatic optimum, in *Encyclopedia of Global Environmental Change*, edited by M. C. MacCracken and J. S. Perry, pp. 514–516, John Wiley, Chichester, U. K.

Meehl, G. A., and A. Hu (2006), Megadroughts in the Indian monsoon region and southwest North America and a mechanism for associated multidecadal Pacific sea surface temperature anomalies, *J. Clim.*, *19*, 1605–1623.

Neff, U., S. J. Burns, A. Mangini, M. Mudelsee, D. Fleitmann, and A. Matter (2001), Strong coherence between solar variability and the monsoon in Oman between 9 and 6 kyr ago, *Nature*, *411*, 290–293.

Newton, A., R. Thunell, and L. Stott (2006), Climate and hydrographic variability in the Indo-Pacific Warm Pool during the last millennium, *Geophys. Res. Lett.*, *33*, L19710, doi:10.1029/2006GL027234.

Overpeck, J., D. Anderson, S. Trumbore, and W. Prell (1996), The southwest Indian monsoon over the last 18,000 years, *Clim. Dyn.*, *12*, 213–225.

Pant, G. B., K. Rupa Kumar, N. A. Sontakke, and H. P. Borgaonkar (1993), Climate variability over India on century and longer time scales, in *Tropical Meteorology*, edited by R. N. Keshavamurty and P. C. Joshi, pp. 149–158, Tata McGraw-Hill, New Delhi, India.

Phadtare, N. R., and R. K. Pant (2006), A century-scale pollen record of vegetation and climate history during the past 3500 years in the Pinder Valley, Kumaon Higher Himalaya, India, *J. Geol. Soc. India*, *68*, 495–506.

Rozanski, K., L. Araguás-Araguás, and R. Gonfiantini (1993), Isotopic patterns in modern global precipitation, in *Climate Change in Continental Isotopic Records*, *Geophys. Monogr. Ser.*, vol. 78, edited by P. K. Swart et al., pp. 1–36, AGU, Washington, D. C.

Schulz, H., U. von Rad, and H. Erlenkeuser (1998), Correlation between Arabian Sea and Greenland climate oscillations of the past 110,000 years, *Nature*, *393*, 54–57.

Shindell, D. T., G. A. Schmidt, M. E. Mann, D. Rind, and A. Waple (2001), Solar forcing of regional climate change during the Maunder Minimum, *Science*, *294*, 2149–2152.

Shindell, D. T., G. Faluvegi, R. L. Miller, G. A. Schmidt, J. E. Hansen, and S. Sun (2006), Solar and anthropogenic forcing of tropical hydrology, *Geophys. Res. Lett.*, *33*, L24706, doi:10.1029/2006GL027468.

Sinha, A., K. G. Cannariato, L. D. Stott, H.-C. Li, C.-F. You, H. Cheng, R. L. Edwards, and I. B. Singh (2005), Variability of southwest Indian summer monsoon precipitation during the Bølling-Ållerød, *Geology*, *33*, 813–816.

von Rad, U., K. H. Michels, H. Schulz, W. H. Berger, and F. Sirocko (1999), A 5000-yr record of climate change in varved sediments from the oxygen minimum zone off Pakistan, northeastern Arabian Sea, *Quat. Res.*, *51*, 39–53.

Yadava, M. G. (2002), Stable isotope systematics in cave calcites: Implications to past climatic changes in tropical India, Ph.D. thesis, 207 pp, Phys. Res. Lab., Navrangpura, India.

Yadava, M. G., and R. Ramesh (2005), Monsoon reconstruction from radiocarbon dated tropical Indian speleothems, *Holocene*, *15*, 48–59.

Yuan, D., et al. (2004), Timing, duration, and transitions of the last interglacial Asian monsoon, *Science*, *304*, 575–578.

Yurtsever, Y., and J. R. Gat (1981), Atmospheric waters, in *Stable Isotope Hydrology: Deuterium and Oxygen-18 in the Water Cycle*, edited by J. R. Gat and R. Gonfiantini, *IAEA Tech. Rep. Ser. 210*, Int. At. Energy, Agency, Vienna.

Zhang, R., and T. L. Delworth (2005), Simulated tropical response to a substantial weakening of the Atlantic thermohaline circulation, *J. Clim.*, *18*, 1853–1860.

K. G. Cannariato and L. D. Stott, Department of Earth Sciences, University of Southern California, Los Angeles, CA 90089, USA.

H. Cheng and R. L. Edwards, Department of Geology and Geophysics, University of Minnesota, Minneapolis, MN 55455, USA.

R. Ramesh and M. G. Yadava, Physical Research Laboratory, Navrangpura, Ahmedabad 380009, India.

I. B. Singh, Department of Geology, Lucknow University, Lucknow 226002, India.

A. Sinha, Department of Earth Sciences, California State University, Dominguez Hills, Carson, CA 90747, USA. (asinha@csudh.edu)