

A positive ‘amount effect’ in the Sahayadri (Western Ghats) rainfall

M. G. Yadava¹, R. Ramesh^{1,*} and K. Pandarinath²

¹Planetary and Geosciences Division, Physical Research Laboratory, Navrangpura, Ahmedabad 380 009, India

²Departamento de Sistemas Energéticos, Centro de Investigación en Energía, Universidad Nacional Autónoma de México, Priv. Xochicalco S/No., Col Centro, Apartado Postal 34, Temixco, Mor. 62580, Mexico

We present stable oxygen isotopic ratios ($\delta^{18}\text{O}$) of rainfall collected over three years (July to October 2000–02) at Mangalore. We observe an apparent positive correlation between the monthly, amount-weighted mean oxygen isotopic ratio of the rainwater and monthly total rainfall, contrary to the generally expected negative trend in many island and continental stations. We offer an explanation for this observation and propose that the ‘amount effect’ in the annual rainfall still remains with a negative slope, and thus can be useful in palaeomonsoon reconstruction using oxygen isotopic variations of annually laminated speleothems.

Keywords: Amount effect, oxygen isotopic ratio, rainfall.

In tropical locations, unlike mid-latitudes, the stable oxygen or hydrogen (δD) isotopic composition of modern rainfall does not depend significantly on the local air temperatures^{1–4}. Instead, it decreases with increasing amount of rainfall^{1,4–6}. This is known as the ‘amount effect’, i.e. higher amount of rainfall is associated with lesser ^{18}O or D content of the rainwater. This effect is a manifestation of gradual saturation of air masses below the cloud base and also due to the preferential removal of $^{18}\text{O}/\text{D}$ as rain-out continues. Initially, the air mass is under-saturated and this leads to evaporation of water droplets and consequently these are enriched in $^{18}\text{O}/\text{D}$. As precipitation proceeds, air masses become saturated and, in turn, the enrichment due to evaporation gradually diminishes^{1,4,7}. In addition, if we assume the Rayleigh type³ of isotopic fractionation, the oxygen isotopic composition of the air mass is given by:

$$\delta^{18}\text{O} = \delta^{18}\text{O}^0 + \varepsilon \ln f, \quad (1)$$

where $\delta^{18}\text{O}$ and $\delta^{18}\text{O}^0$ denote the later and initial oxygen isotopic compositions of the vapour mass respectively, ε is the oxygen isotopic fractionation factor between the rain and vapour expressed in per mil (‰) units, and f is the fraction of the original vapour left. In eq. (1), $\delta^{18}\text{O}$ and $\delta^{18}\text{O}^0$ are always negative relative to the Standard Mean Ocean Water (SMOW), as during evaporation from the

ocean to form clouds, the heavier isotope ^{18}O prefers to stay in the water phase, $\ln f$ is also negative (natural logarithm of a number smaller than unity), and only ε is positive. Thus higher the rainfall, smaller the f (fraction of vapour left) and hence more depleted is the rainwater in ^{18}O . If the vapour mass entering an area is assumed to be unity and it leaves after raining a fraction Δf , then the amount of vapour leaving the area is $(1 - \Delta f)$. The oxygen isotopic composition of the vapour leaving is given by³:

$$\delta^{18}\text{O}_f = \delta^{18}\text{O}^0 + \varepsilon \ln(1 - \Delta f) \approx \delta^{18}\text{O}^0 - \varepsilon \Delta f \quad (\text{for small } \Delta f). \quad (2)$$

Therefore, the mean isotopic composition of the rainwater $\delta^{18}\text{O}_r$ is obtained by mass balance:

$$\delta^{18}\text{O}^0 = (1 - \Delta f)[\delta^{18}\text{O}^0 - \varepsilon \Delta f] + \Delta f \delta^{18}\text{O}_r. \quad (3)$$

That is,

$$\delta^{18}\text{O}_r = (\delta^{18}\text{O}^0 + \varepsilon) - \varepsilon \Delta f. \quad (4)$$

Thus higher the rainfall extracted (higher Δf), more negative is $\delta^{18}\text{O}_r$. The bracketed term on the right-hand side in eq. (4) represents the mean isotopic composition of the source, i.e. the sea water (must be around 0‰, see eq. (5) below).

Worldwide modern precipitation data available from island stations in the equatorial belt (collected by IAEA) have shown a linear relationship² between the mean monthly $\delta^{18}\text{O}_m$ of precipitation and the mean monthly rainfall:

$$\delta^{18}\text{O}_m = -(0.015 \pm 0.002)P_m - (0.47 \pm 0.42), \quad (5)$$

where P_m is the mean monthly rainfall in mm, with a linear correlation coefficient (r) of 0.87 for 14 island stations (each has at least 40 monthly observations). As these are island stations, seasonal temperature variations are small. Average rate of depletion is found to be $-1.5 \pm 0.2\%$ for a 100 mm increase in the mean monthly rainfall.

The major source of precipitation over India is the SW monsoon, where the hot, dry air prevailing over India in April and May is replaced by moist, oceanic air coming from the SW^{8,9}. The mean rainfall over the Indian plains in the SW monsoon period is 925 mm as against 145 mm for the rest of the year. Yadava and Ramesh¹⁰ have reported the amount effect for the Indian monsoon rainfall based on samples collected from Jharsuguda, Orissa. This station receives most of its monsoon precipitation from storms and depressions that form in the Head-Bay of Bengal and progressively move northwestwards. The relation between the amount of rainfall and $\delta^{18}\text{O}$ was:

$$\delta^{18}\text{O}_i = -(0.092 \pm 0.011)S_i - (3.37 \pm 0.41), \quad (6)$$

where S_i is the amount (mm) of each rainfall spell, $r = 0.87$ (number of samples, $n = 24$). The average rate of ^{18}O depletion was: $-9.2 \pm 1.1\%$ per 100 mm increase in rain.

*For correspondence. (e-mail: rramesh@prl.res.in)

The above data were also converted to monthly weighted mean $\delta^{18}\text{O}$ and monthly total amount by simple isotopic mass balance of all the collections during each month. Expression for the best-fit line obtained from three data points for July–September was:

$$\delta^{18}\text{O} = -(0.022 \pm 0.008)P_m - (2.51 \pm 1.89) \\ n = 3, r = 0.94, \text{ significant at } 0.15 \text{ level.} \quad (7)$$

This shows that there is a depletion of $2.2 \pm 0.8\%$ per 100 mm increase in the monthly monsoon rainfall amount. This value agrees well with that of 1.5 ± 0.2 shown by the IAEA stations. The amount effect may have significant spatial variation, as proposed by Ramesh and Yadava¹¹. The intercept is significantly negative, as expected, since the surface waters in the Bay of Bengal are depleted in ^{18}O because of freshwater discharge from rivers.

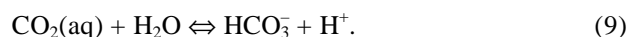
In the Sahayadri (Western Ghats) region, the source of monsoonal precipitation is the orographic ascent of air masses. Therefore, the amount effect needs to be calibrated independently. To this end we collected rainwater during the different spells in the period 2000–02 at Mangalagangothri, Mangalore, Karnataka. Rainwater samples of individual events were collected in a 20 l plastic carboy with a known diameter funnel fixed at the mouth and kept at the rooftop. Care was taken to avoid any evaporation or loss of water, to avoid any isotopic fractionation.

The CO_2 equilibration method^{12,13} was used. Equilibration was carried out on Water Equilibration System (WES) attached with the GEO 20-20 mass spectrometer. The operation of the WES was fully automated and controlled by software. Water samples (each 1 ml) were filled in a 6.8 ml standard glass bottle. A threaded cap having a neoprene septum sealed the bottle, preventing any leakage. About 40 such bottles filled with water samples were kept in a sequence in a rack. A secondary laboratory reference 'NRM' (taken from the river Narmada) was also filled in similar glass bottles and one each placed after a batch of ten samples. Its oxygen isotopic ratio with respect to VSMOW (Vienna Standard Mean Ocean Water, the current international water standard, after the original SMOW was exhausted) was used as a primary reference for the $\delta^{18}\text{O}$ and δD measurements of water samples^{14,15}. The measurement of oxygen isotopes was carried out in three steps.

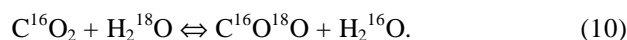
(i) Flushing of the air: A sharp Gilson needle with double holes which can move up and down into the sample bottle pierced the neoprene septum from the top of the cap, so that both holes are inside the bottle. Carbon dioxide from a cylinder (obtained from Vadilal Gas Company and labelled as VDCO_2) was let into the sample bottle. For 1 min duration, the cylinder CO_2 flushed air through the vent. After this, relatively pure CO_2 remained inside the sample bottles. During flushing water vapour was also flushed out, the amount too small to affect the isotope determinations significantly.

(ii) CO_2 equilibration: The needle moved up and the holes in the neoprene septum got immediately closed by its propensity. Carbon dioxide and water were left overnight (~ 12 h) in the fully sealed and isolated condition for complete equilibration to occur. The rack carrying all the sample glass bottles was kept at 35°C to enhance the rate of isotopic exchange and to reduce the equilibration time.

Isotopic exchange took place due to the following reactions¹⁶:



The isotopic exchange reaction is written as:



After sometime (e.g. 12 h at 35°C in the present case), the reaction reached an equilibrium and the ratios ($^{18}\text{O}/^{16}\text{O}$) in water and carbon dioxide were related by a temperature-dependent fractionation factor α :

$$(^{18}\text{O}/^{16}\text{O})_{\text{CO}_2} = \alpha \cdot (^{18}\text{O}/^{16}\text{O})_{\text{H}_2\text{O}}, \quad (11)$$

or

$$1000 + \delta^{18}\text{O}_{\text{CO}_2} = (1000 + \delta^{18}\text{O}_{\text{H}_2\text{O}}) \cdot \alpha \quad (12)$$

The value¹⁷ of α at 35°C is 1.0393.

(iii) $\delta^{18}\text{O}$ measurement: The equilibrated CO_2 was passed through a moisture trap at -100°C , where all the water vapour was trapped and pure CO_2 was passed into the sample reservoir of the mass spectrometer. The reference reservoir was filled with VDCO_2 gas. The flushing and measurements were carried out in the automatic mode¹⁸.

The $\delta^{18}\text{O}_{\text{CO}_2}^o$, the oxygen isotope composition of CO_2 gas isotopically equilibrated to the original water sample (assuming that the amount of water was infinite and hence it did not alter the original isotopic composition of water during the equilibration) is given^{13,19} by:

$$\delta^{18}\text{O}_{\text{CO}_2}^o = (1 + \alpha \cdot (n/N)) \cdot \delta^{18}\text{O}_{\text{CO}_2} \\ - \alpha \cdot (n/N) \cdot \delta^{18}\text{O}_{\text{CO}_2}^i, \quad (13)$$

where $\delta^{18}\text{O}_{\text{CO}_2}$ is the oxygen isotope composition of equilibrated CO_2 and $\delta^{18}\text{O}_{\text{CO}_2}^i$ the initial oxygen isotope composition before equilibration, which is 0‰ as the same gas (VDCO_2) was used as the reference and N the gram-atoms of oxygen present in the water sample and n that in CO_2 .

As the amount of oxygen in CO_2 gas (5.8 ml at 1 atm pressure) in the 6.8 ml glass bottle was insignificant compared to that in the water phase (1 ml), hence, $n/N \approx 0$. For practical purposes, $\delta^{18}\text{O}_{\text{CO}_2}^o$ and $\delta^{18}\text{O}_{\text{CO}_2}$ were the same and the oxygen isotope composition of the water ($\delta^{18}\text{O}_w$) sample could be obtained using eq. (11).

During measurements NRM water, a secondary laboratory standard, was placed after each batch of ten samples and also equilibrated with the same CO₂ (VDCO₂). If CO₂ measurements for NRM were represented as δ¹⁸O_{NRM-VDCO₂} and for the sample as δ¹⁸O_{SAM-VDCO₂}, then oxygen isotope composition of the water sample relative to NRM water, i.e. δ¹⁸O_{SAM-NRM} was:

$$1000 + \delta^{18}\text{O}_{\text{SAM-NRM}} = (1000 + \delta_{\text{SAM-VDCO}_2}) / (1000 + \delta_{\text{NRM-VDCO}_2}) \quad (14)$$

As the temperature of equilibration for the water samples as well as the NRM standard was the same (35°C), the value of α was the same in both the cases and cancelled out in the above expression.

Values with respect to SMOW standard were given as:

$$\delta^{18}\text{O}_{\text{SAM-SMOW}} = \delta^{18}\text{O}_{\text{SAM-NRM}} + \delta^{18}\text{O}_{\text{NRM-SMOW}} + \delta^{18}\text{O}_{\text{SAM-NRM}} \cdot \delta^{18}\text{O}_{\text{NRM-SMOW}} \cdot 10^{-3} \quad (15)$$

The overall precision of the δ¹⁸C measurements was within ±0.15‰.

A total of 109 samples were analysed and the results are shown in Figure 1 and discussed in the context of our earlier results from Orissa.

The data exhibit a linear correlation (*r* = 0.23, significant at 0.05 level) between δ¹⁸O and the amount of event rainfall (we avoid using the word ‘daily’ rainfall because it implies at least one rain event everyday, which is seldom true; see Figure 2 a):

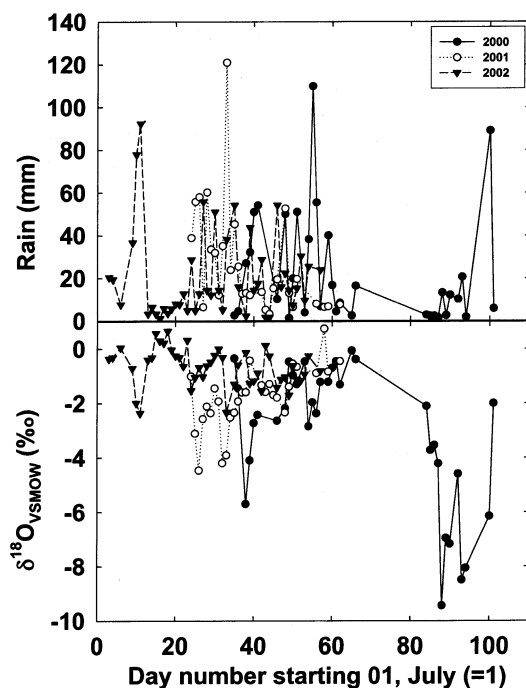


Figure 1. Plot of daily rainfall amount at Mangalore (top panel), and its oxygen isotopic composition (bottom panel).

$$\delta^{18}\text{O} = (-0.019 \pm 0.008) \times \text{event rain (mm)} - (1.29 \pm 0.25) \quad (16)$$

The amount weighted monthly mean δ¹⁸O, however, shows a relationship with positive slope (Figure 3), i.e.

$$\delta^{18}\text{O}_m = (0.008 \pm 0.002) \times \text{monthly rain (mm)} - (5.85 \pm 0.91) \quad (17)$$

with *r* = 0.85, significant at 0.02 level (number of weighted samples = 7). We explain this apparently ‘unusual’ positive slope as follows: If we consider only July and August precipitation (individual rain events), there is a significant negative slope (Figure 2 b):

$$\delta^{18}\text{O} = -(0.027 \pm 0.004) \times \text{event rain (mm)} - (0.60 \pm 0.14) \quad (18)$$

with *r* = 0.56, significant at 0.01 level (*n* = 95). This is similar to the relationship obtained earlier for the Indo-Gangetic belt (eq. (6)), albeit with a different but still negative slope. The intercept is somewhat below zero as

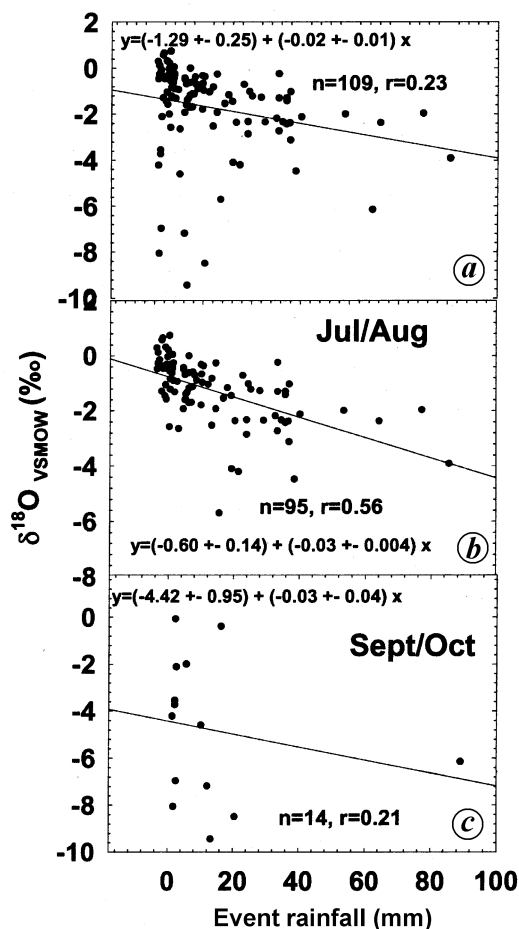


Figure 2 a-c. Stable oxygen isotopic composition of Mangalore rainfall vs the amount.

the Arabian Sea near Mangalore also receives freshwater similar to the Bay of Bengal. However, most rain events in September and October are lower in amount and also isotopically depleted relative to those of the preceding months (Figures 1 and 2 c). Thus when monthly values are plotted (Figure 3), there is an apparent positive slope, with low rainfall, depleted isotope ratios plotting on the lower left of the graph and the high rainfall, enriched values plotting on the top right, with a distribution in between.

Speleothems, a collective name for the calcium carbonate precipitates occurring in natural limestone caves, can be used as a high-resolution long-term rainfall proxy²⁰. Usually, they record the mean annual (or mean sub-decadal, depending on their growth rates) oxygen isotope ratios of the local rainfall, and thus these values can be interpreted in terms of the total amount of annual/sub-decadal rainfall in the locality²⁰. It is well known that the monsoon rainfall over the Sahayadri (known as the Konkan meteorological subdivision of the India Meteorological Department), is significantly correlated with the all-India monsoon rainfall²¹. Speleothem $\delta^{18}\text{O}$ is generally an imprint of the $\delta^{18}\text{O}$ of the local rainwater, unless the cave is underground, where groundwater contamination is possible. A fast-growing speleothem²⁰ from Ulvi District, Karnataka has registered extreme rainfall events of the region. Heavy rainfall years appear as depleted signals, and drought years are seen as enriched signals in the speleothem $\delta^{18}\text{O}$. This is in accordance with the isotopic behaviour of the individual rain events shown in Figure 2 a. It seems that oxygen isotopic values of speleothems, due to fast growth, are influenced by the isotopic signature of individual rain events. This implies that speleothem $\delta^{18}\text{O}$ does not show average (e.g. monthly) behaviour of several rain events, as observed in Figure 2, unless another (following) process is envisaged: Higher annual rainfall implies increased rains in September and October as well, in addition to July and August (i.e. an 'extended' monsoon season), and hence the mean annual value of the precipitation $\delta^{18}\text{O}$ will be lower. In drought years, rainfall may be restricted to (i) less number of rainfall events, and

also less rainfall and (ii) most rain events are likely in June and July, and hence the annual mean $\delta^{18}\text{O}$ may be relatively enriched. Therefore, this result has a special significance for the reconstruction of the Indian monsoon rainfall using speleothem $\delta^{18}\text{O}$ values, as this region has many limestone caves. Long-term monitoring of the annual mean isotopic composition of rainfall in this region would be important to verify the above conclusions.

Systematic measurements of the stable oxygen isotopic composition of the Arabian Sea branch of the Indian monsoon rainfall at Mangalore revealed an apparent positive amount effect as far as monthly rainfall is concerned. This is explained as due to the significant differences in the early and late monsoon rains in terms of both the amount and isotopic compositions of rainfall (the latter is less in amount and also in the isotopic composition). Overall, while considering the annual average $\delta^{18}\text{O}$ values, as recorded by the speleothems, the amount effect remains with a negative slope. This relationship can be useful in quantifying the proxy climatic records from speleothems and tree rings from this climatically important region^{10,22,23}.

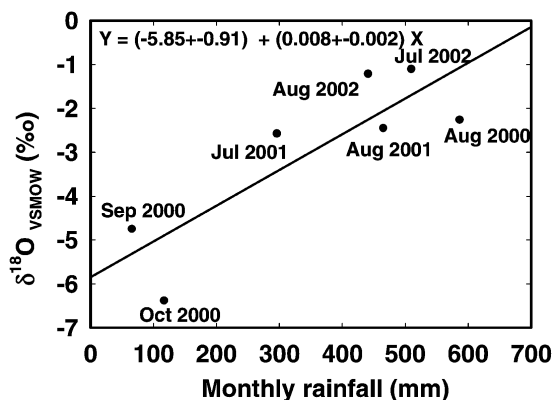


Figure 3. Amount effect in Mangalore rainfall.

- Dansgaard, W., Stable isotopes in precipitation. *Tellus*, 1964, **16**, 436–468.
- Yurtsever, Y. and Gat, J. R., Atmospheric waters. In *Stable Isotope Hydrology Deuterium and Oxygen-18 in the Water Cycle* (eds Gat, J. R. and Gonfiantini, R.), Technical Report Series No. 210, IAEA, Vienna, 1981, pp. 103–142.
- Clark, I. D. and Fritz, P., In *Environmental Isotopes in Hydrogeology*, Lewis Publisher, 1997, p. 328.
- Fricke, H. C. and O'Neil, J. R., The correlations between $^{18}\text{O}/^{16}\text{O}$ ratios of meteoric water and surface temperature: Its use in investigating terrestrial climate change over geologic time. *Earth Planet. Sci. Lett.*, 1999, **170**, 181–196.
- Bar-Matthews, M., Ayalon, A., Matthews, A., Sass, E. and Halicz, L., Carbon and oxygen isotope study of the active water-carbonate system in a karstic Mediterranean cave: Implications for palaeoclimatic research in semiarid regions. *Geochim. Cosmochim. Acta*, 1996, **60**, 337–347.
- Bar-Matthews, M. and Ayalon, A., Late Quaternary paleoclimate in the eastern Mediterranean region from stable isotope analysis of speleothems at Soreq Cave, Israel. *Geochim. Cosmochim. Acta*, 1977, **47**, 155–168.
- Rozanski, K., Araguas-Araguas, L. and Gonfiantini, R., Isotopic patterns in modern global precipitation. In *Climate Change in Continental Climate Records* (eds Swart, P. K. et al.), Am. Geophys. Union, Geophys. Monogr., 1993, vol. 78, pp. 1–36.
- Pant, G. B. and Rupa Kumar, K., *Climates of South Asia, Belhaven Studies in Climatology* (ed. Gregory, S.), Wiley, Chichester, 1997, p. 320.
- Ramesh, R., High resolution Holocene monsoon records from different proxies, an assessment of their consistency. *Curr. Sci.*, 2001, **81**, 1432–1436.
- Yadava, M. G. and Ramesh, R., Monsoon reconstruction from radiocarbon dated tropical Indian speleothems. *Holocene*, 2005, **15**, 48–59.
- Ramesh, R. and Yadava, M. G., Climate and water resources of India. *Curr. Sci.*, 2005, **89**, 818–824.
- Epstein, S. and Mayeda, T., Variation of ^{18}O content of waters from natural sources. *Geochim. Cosmochim. Acta*, 1953, **4**, 213–221.

13. Gonfiantini, R., The δ -notation and the mass-spectrometric measurement techniques. In *Stable Isotope Hydrology Deuterium and Oxygen-18 in the Water Cycle* (eds Gat, J. R. and Gonfiantini, R.), Technical Report Series No. 210, IAEA, Vienna, 1981, pp. 35–84.
14. Baertschi, P., Absolute ^{18}O content of Standard Mean Ocean Water. *Earth Planet. Sci. Lett.*, 1976, **31**, 341–344.
15. Hageman, R., Nief, G. and Roth, E., Absolute isotopic scale for deuterium analysis of natural waters. Absolute D/H ratio for SMOW. *Tellus*, 1970, **22**, 712–715.
16. Mook, W. G., In *Handbook of Environmental Isotope Geochemistry* (eds Fritz, P. and Fontes, J. Ch.), 1980, vol. 1, pp. 49–74.
17. Criss, R. E., *Principles of Stable Isotope Distribution*, Oxford University Press, New York, 1999, 1st edn, p. 254.
18. Ramesh, R. and Yadava, M. G., Significance of stable isotopes and radiocarbon dating in fluvial environments. In *Lecture Notes, DST Programme on Fluvial Systems* (ed. Chamyal, L. S.), Dept of Geology, M.S. University of Baroda, Vadodara, 2004, pp. 107–149.
19. Craig, H., Isotopic standards for carbon and oxygen and correction factors for mass-spectrometric analysis of carbon dioxide. *Geochim. Cosmochim. Acta*, 1957, **12**, 133.
20. Yadava, M. G., Ramesh, R. and Pant, G. B., Past monsoon rainfall variations in peninsular India, recorded in a 331 year old speleothem. *Holocene*, 2004, **14**, 517–524.
21. Gadgil, S., Yadumani and Joshi, N. V., Coherent rainfall zones of the Indian region. *Int. J. Climatol.*, 1993, **13**, 547–566.
22. Yadava, M. G. and Ramesh, R., Decadal variability in the Indo-Gangetic monsoon rainfall during the last ~2800 years: Speleothem $\delta^{18}\text{O}$ evidence from the Sota cave, Uttar Pradesh. In *Antarctic Geoscience: Ocean–Atmosphere Interaction, and Palaeoclimatology* (eds Rajan, S. and Pandey, P. C.), Special Publication of NCAOR, Goa, 2005, pp. 184–197.
23. Korisettar, R. and Ramesh, R., The Indian Monsoon: Roots, relations and relevance. In *Archaeology and Interactive Disciplines, Indian Archaeology in Retrospect, Vol. III* (eds Settar, S. and Korisettar, R.), Indian Council of Historical Research, Manohar Publications, New Delhi, 2002, pp. 23–59.

Received 23 June 2006; revised accepted 11 May 2007

Late Quaternary bedrock incision in the Narmada river at Dardi Falls

Avijit Gupta¹, Vishwas S. Kale^{2,*},
Lewis A. Owen³ and A. K. Singhvi⁴

¹School of Geography, University of Leeds, Leeds, LS2 9JT, UK

²Department of Geography, University of Pune, Pune 411 007, India

³Department of Geology, University of Cincinnati, Cincinnati, OH 45221, USA

⁴Planetary and Geosciences Division, Physical Research Laboratory, Ahmedabad 380 009, India

Fluvial incision in bedrock is common in many rivers of the Indian Peninsula. We investigated a site in the gorge of the Narmada river at Dardi Falls that displays geomorphic evidence of intense bedrock erosion. We

report here a terrestrial cosmogenic radionuclide date from an eroded rock surface in Peninsular India. Terrestrial cosmogenic radionuclide dating of the rock surface adjacent to the inner gorge indicated that the minimum age of the gorge is 40 ka. We suggest that the present gorge has developed in two phases, separated by a period of large-scale aggradation that filled the gorge with alluvium. Gorge formation is most likely associated with tectonic activity in the Son–Narmada–Tapi lineament zone. Erosion at this scale also requires large palaeodischarges with high unit stream power. This study illustrates the powers of combining newly developing geomorphic, DEM and geochronological methods to elucidate the dynamics and nature of landscape evolution.

Keywords: Bedrock erosion, cosmogenic radionuclide dating, DEM, gorge, Narmada.

THE erosion-dominant landscape of the Indian Peninsula is characterized by rivers downcutting into underlying bedrock. Many large rivers of Peninsular India flow through scablands, rock gorges and knickpoint-related waterfalls. However, the dates of such downcutting or rates of erosion are not known for any of the Peninsular rivers, with the exception of the pre-Holocene Tapi Gorge¹ and the historical evidence of rapid bedrock erosion by the Indrayani river². We report here a ^{10}Be terrestrial cosmogenic radionuclide surface exposure age determination for the Narmada Gorge at Dardi Falls and discuss its significance in terms of regional geomorphology and bedrock erosion.

The 1300-km long Narmada river in Central India flows through three major bedrock gorges separated by alluvial basins. In the downstream direction these are the Marble Canyon near Jabalpur, Punasa Gorge near Khandwa, and Dhadgaon Gorge west of Badwani³. The present study is on the Punasa Gorge, where at Dardi (22°18'49"N and 76°21'11"E) the Narmada river (Figure 1) has eroded a >100 m deep and 1–7 km wide gorge with an inner canyon⁴ (Figure 2). A ca. 10-m high waterfall occurs at the head of the inner canyon. Thick (>50 m) alluvial deposits of late Quaternary age overlie the bedrock on the right bank (Figure 2a), whereas the left bank rises to a set of strath terraces with abraded and water-polished rock surfaces. The back of the terraces is covered by riverine alluvium and soil. Similar rock surfaces are likely to extend below the alluvium that occurs on the top of the right bank.

Quartzites of the Vindhyan Supergroup constitute the local bedrock. To the south, a major fault runs parallel to the Narmada river⁵ (Figure 1c). The river flows through the large tectonic feature of the Son–Narmada–Tapi (SONATA) lineament zone characterized by neotectonism, moderate seismicity^{5–7}, and several longitudinal fault-bound blocks with an episodic history of vertical and lateral movements⁶. The resulting river morphology demonstrates tectonic and structural control as illustrated in Figure 1c and d, which is a digital elevation model (DEM) of the Dardi Falls area

*For correspondence. (e-mail: vskale@unipune.ernet.in)