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

M. G. Yadava¹, R. Ramesh¹* 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/N, Col Centro, Apartado Postal 34, Temixco, Mor. 62580, Mexico

We present stable oxygen isotopic ratios ($\delta^{18}O$) 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 ($\delta^D$) isotopic composition of modern rainfall does not depend significantly on the local air temperatures¹–⁴. Instead, it decreases with increasing amount of rainfall⁵–⁶. 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 $^{18}O/D$ 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 $^{18}O/D$. As precipitation proceeds, air masses become saturated and, in turn, the enrichment due to evaporation gradually diminishes⁷–⁹. In addition, if we assume the Rayleigh type³ of isotopic fractionation, the oxygen isotopic composition of the air mass is given by:

$$\delta^{18}O = \delta^{18}O^o + \varepsilon \ln f,$$

where $\delta^{18}O$ and $\delta^{18}O^o$ denote the later and initial oxygen isotopic compositions of the vapour mass respectively, $\varepsilon$ 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), $\delta^{18}O$ and $\delta^{18}O^o$ 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. In $f$ is also negative (natural logarithm of a number smaller than unity), and only $\varepsilon$ 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 $\Delta f$, then the amount of vapour leaving the area is $(1 – \Delta f)$. The oxygen isotopic composition of the vapour leaving is given by:\

$$\delta^{18}O = \delta^{18}O^o + \varepsilon \ln(1 – \Delta f) = \delta^{18}O^o – \varepsilon \Delta f \quad \text{(for small } \Delta f \text{)}.$$  

(2)

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

$$\delta^{18}O = (1 – \Delta f)[\delta^{18}O^o – \varepsilon \Delta f] + \Delta f/\delta^{18}O_r.$$  

(3)

That is,

$$\delta^{18}O_r = (\delta^{18}O^o + \varepsilon) – \varepsilon \Delta f.$$  

(4)

Thus higher the rainfall extracted (higher $\Delta f$), more negative is $\delta^{18}O_r$. 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 $\delta^{18}O_m$ of precipitation and the mean monthly rainfall:

$$\delta^{18}O_m = –(0.015 \pm 0.002)P_m – (0.47 \pm 0.42),$$  

(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 SW³.⁴.⁵. 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 $\delta^{18}O$ was:

$$\delta^{18}O_r = –(0.092 \pm 0.011)S_r – (3.37 \pm 0.41),$$  

(6)

where $S_r$ 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 rainfall.

---

¹For correspondence. (e-mail: rramesh@prl.res.in)
The above data were also converted to monthly weighted mean $\delta^{18}O$ 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:

$$
\delta^{18}O = -(0.022 \pm 0.008)P_m - (2.51 \pm 1.89)
$$

$$
n = 3, r = 0.94, \text{significant at 0.15 level. (7)}
$$

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 $1.5 \pm 0.2$ shown by the IAEA stations. The amount effect may have significant spatial variation, as proposed by Ramesh and Yadava\textsuperscript{11}.

The $CO_2$ equilibration method\textsuperscript{12,13} was used. Equilibration of individual events were collected in a 20 l plastic carboy during the different spells in the period 2000–02 at Mangalagangotri, Mangalore, Karnataka. Rainwater samples from a cylinder (obtained from Vadilal Gas Company and labelled as VDCO$_2$) was let into the sample bottle. For the current international water standard, after the original 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\textsuperscript{13,19} by:

$$
\delta^{18}O_{W} = (1000 + \delta^{18}O_{CO_2}) \cdot \alpha
$$

(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 bottle. 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 reactions\textsuperscript{16}:

$$
CO_2(g) \leftrightarrow CO_2(aq), \quad (8)
$$

$$
CO_2(aq) + H_2O \Leftrightarrow HCO_3^- + H^+. \quad (9)
$$

The isotopic exchange reaction is written as:

$$
C^{16}O_2 + H_2^{18}O \Leftrightarrow C^{16}O^{18}O + H_2^{16}O. \quad (10)
$$

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

$$
(\delta^{18}O/^{16}O)_CO_2 = \alpha \cdot (\delta^{18}O/^{16}O)_{H_2O} \quad (11)
$$

or

$$
1000 + \delta^{18}O_{CO_2} = (1000 + \delta^{18}O_{H_2O}) \cdot \alpha \quad (12)
$$

The value\textsuperscript{17} of $\alpha$ at 35°C is 1.0393.

(iii) $\delta^{18}O$ 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 mode\textsuperscript{18}.

The $\delta^{18}O_{CO_2}$, 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\textsuperscript{13,19} by:

$$
\delta^{18}O_{CO_2} = (1 + \alpha \cdot (n/N)) \cdot \delta^{18}O_{CO_2}
$$

$$
- \alpha \cdot (n/N) \cdot \delta^{18}O_{CO_2} \quad (13)
$$

where $\delta^{18}O_{CO_2}$ is the oxygen isotope composition of equilibrated $CO_2$ and $\delta^{18}O_{CO_2}$, 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, $n/N = 0$. For practical purposes, $\delta^{18}O_{CO_2}$ and $\delta^{18}O_{CO_2}$ were the same and the oxygen isotope composition of the water ($\delta^{18}O_{w}$) 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:

\[ 1000 + \delta^{18}O_{\text{SAM-NRM}} = \frac{1000 + \delta^{18}O_{\text{SAM-VDCO₂}}}{1000 + \delta^{18}O_{\text{NRM-VDCO₂}}}. \]  

(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:

\[ \delta^{18}O_{\text{SAM-SMOW}} = \delta^{18}O_{\text{SAM-NRM}} + \delta^{18}O_{\text{NRM-SMOW}} + \delta^{18}O_{\text{SAM-NRM}} \cdot \delta^{18}O_{\text{NRM-SMOW}} \cdot 10^{-3}. \]  

(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 δ²⁸O 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):

\[ \delta^{28}O = (-0.019 \pm 0.008) \times \text{event rain (mm)} - (1.29 \pm 0.25). \]  

(16)

The amount weighted monthly mean δ²⁸O, however, shows a relationship with positive slope (Figure 3), i.e.

\[ \delta^{28}O_m = (0.008 \pm 0.002) \times \text{monthly rain (mm)} - (5.85 \pm 0.91), \]  

(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):

\[ \delta^{28}O = -(0.027 \pm 0.004) \times \text{event rain (mm)} - (0.60 \pm 0.14), \]  

(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
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 2c). 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 subdecadal, 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/subdecadal rainfall in the locality. 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 δ18O is generally an imprint of the δ18O of the local rainwater, unless the cave is underground, where groundwater contamination is possible. A fast-growing speleothem 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 δ18O. This is in accordance with the isotopic behaviour of the individual rain events shown in Figure 2a. 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 δ18O 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 δ18O 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 δ18O may be relatively enriched. Therefore, this result has a special significance for the reconstruction of the Indian monsoon rainfall using speleothem δ18O 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 δ18O 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 region10,22,23.

RESEARCH COMMUNICATIONS

6. Bar-Matthews, M. and Ayalon, A., Late Quaternary paleoclimate in the eastern Mediterranean region from stable isotope analysis of speleothems and tree rings from this climatically important region.
Late Quaternary bedrock incision in the Narmada river at Dardi Falls

Avijit Gupta¹, Vishwas S. Kale²*, 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 Indravani river². We report here a ¹⁰Be 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 Dhadaaon 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 2 (a)), 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 1 (c)). The river flows through the large tectonic feature of the Son–Narmada–Tapi (SONATA) lineament zone characterized by neotectonism, moderate seismicity⁶–⁷, and several longitudinal fault-bound blocks with an episodic history of vertical and lateral movements⁸. The resulting river morphology demonstrates tectonic and structural control as illustrated in Figure 1 (c) and (d), which is a digital elevation model (DEM) of the Dardi Falls area.

RESEARCH COMMUNICATIONS


Received 23 June 2006; revised accepted 11 May 2007

Late Quaternary bedrock incision in the Narmada river at Dardi Falls

Avijit Gupta¹, Vishwas S. Kale²*, 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

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