

Stable oxygen and hydrogen isotope ratios in shallow ground waters from India and a study of the role of evapotranspiration in the Indian monsoon

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Abstract—We have measured the δD and $\delta^{18}O$ values of shallow ground waters along a traverse in the east-northwest parts of India that come under the direct influence of the Indian monsoon. On the premise that shallow ground waters represent an “averaged contemporary precipitation,” their isotopic ratios can be interpreted as proxy indicators of the δD and $\delta^{18}O$ values of the local precipitation. The isotopic ratios thus measured suggest that, as in many other parts of the world, the rain-out process basically follows a Rayleigh-type distillation with a $\delta^{18}O$ inland gradient of about -2 per mil per 1000 km between the coastal Calcutta and inland Delhi. However, the gradient appears to be smaller than what would be expected for a pure Rayleigh process where the vapor reduction factor in Delhi is more than half. This reduced gradient can arise if the original parcel of water vapor is augmented by additional vapor transport through evapotranspiration. A simple box model has been employed to take this into account. This model suggests that up to 40% of water that has precipitated has to be put back into the vapor system to explain the observed isotopic trend. This estimate of 40% is not farfetched when compared to values estimated based on other hydrological balance calculations and is also similar to the 35% obtained in European precipitations. The δD - $\delta^{18}O$ line of the shallow ground waters along this traverse is expressed by the equation $\delta D = 6.4 \delta^{18}O - 1$; significantly different from the Meteoric Water Line (MWL) of $\delta D = 8 \delta^{18}O + 10$. However, if twelve of the samples that lie close to large irrigation systems and rivers are disregarded, the relationship is found to be of the form $\delta D = 7.12 \delta^{18}O + 3$, which is similar to the relationship obtained for the direct precipitation in three IAEA stations in India. These data lend support for the argument that the shallow ground waters serve as good proxy records of the isotopic composition of local precipitation.

INTRODUCTION

AT ANY GIVEN LOCATION the distribution of stable oxygen and hydrogen isotopes in precipitation is influenced by several factors (DANSGAARD, 1964); to name a few: the “temperature effect” that causes the isotopic ratio to change proportionately with the surface air temperature; the “amount effect” in which there is a negative correlation between the δ values and the amount of precipitation; and the “altitude effect” that causes a depletion of the heavier isotope with increasing altitude where the precipitation falls (YURTSEVER, 1975; FRIEDMAN and SMITH, 1976; EHHALT *et al.*, 1963). In spite of these several local effects that have to be taken into account while interpreting the isotope ratio in precipitation, it is fortunate that on a global scale and also to a large extent on a regional scale, the rain-out process can be explained by means of a simple mechanism. This simple mechanism, as indicated by DANSGAARD (1964), turns out to be a Rayleigh-type distillation where from an initial oceanic reservoir water vapor is generated and lost through successive steps of precipitation as this parcel of vapor moves polewards. The distribution of the ox-

xygen and hydrogen isotopic ratio in such a condensation process can be expressed as

$$R_p = \alpha R_0 F^{\alpha-1} \quad (1)$$

where R_p = the isotopic ratio of precipitation at a point away from the source region, R_0 = the isotopic ratio of the initial vapor mass, F = the vapor reduction factor at the point where R_p is being calculated, and α = the fractionation factor between the vapor and the liquid phases. Note that this expression assumes constancy of α at every stage of condensation.

Since the fractionation causes the liquid phase to be higher in the concentration of the heavier isotope, one of the main consequences of a process such as the Rayleigh distillation is the “continental effect.” Because of this effect the δ values of the precipitation become progressively more negative as one moves away from the oceanic source region. That this trend has been observed in the majority of cases is a strong support for the Rayleigh distillation hypothesis of precipitation.

SAMPLES AND EXPERIMENTAL METHODS

In all, more than 80 samples encompassed by a triangular sector Ahmedabad-Calcutta-Delhi were collected and analysed (KRISHNAMURTHY, 1984; BHATTACHARYA *et al.*, 1985). However, for the present study which attempts

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to estimate the role of evapotranspiration, only 55 of these are relevant. These 55 samples cover a distance of approximately 1400 km, and form a roughly rectangular track between coastal Calcutta and inland Delhi. The sampled area falls in the "monsoon trough" and directly experiences the influence of the Indian southwestern monsoon that operates between June and September every year (SIKKA, 1977; ANANTHAKRISHNAN, 1977; PISHAROTY, 1965; RAO, 1981). Significantly, this area also experiences the "monsoon depressions" which are characterised by severe precipitation events accompanied by strong winds. The samples came from open dug wells of depth no more than ten meters from the surface. Care was taken to ensure that the wells were situated as far away as possible from irrigation systems, rivers, or lakes. Samples were collected in pre-cleaned dry glass or plastic bottles and filled to the brim and sealed to avoid evaporation. Dug well samples were collected by sending an air-filled bottle upside down to the bottom part of the water column and then tilting it. All the samples were analysed within about a month after arrival in the laboratory to minimize errors that might be caused by storage over long periods of time. Sample locations are indicated in Fig. 1.

The $\delta^{18}\text{O}$ values were determined by a slight modification of the standard technique first suggested by EPSTEIN and MAYEDA (1953). The δD values were determined by reduction of the water sample over hot uranium metal (BIGELEISEN *et al.*, 1952; FRIEDMAN and HARDCASTLE,

1970). Samples were analysed in batches of five, and with each batch a calibrated laboratory standard was analysed. The δ values are reported with respect to SMOW in the usual way and the overall precision determined by repeated analysis of the laboratory standard is ± 0.1 and ± 1.5 per mil for $\delta^{18}\text{O}$ and δD , respectively.

RESULTS AND DISCUSSION

Oxygen isotope ratios

The $\delta^{18}\text{O}$ values of the ground waters as a function of distance from Calcutta are plotted in Fig. 2. The data points other than the solid circles come from localities that do have large rivers and irrigation systems close by and so might have influenced the ground water samples. The "continental effect" causing the depletion in ^{18}O with distance, that is going inland from the coastal Calcutta, is clearly evident. This is to be expected in a typical Rayleigh distillation process where during the monsoon, air masses collect moisture in and around the Bay of Bengal and release as rain when they are carried by the prevailing winds inland in a north-west-west direction. The inland gradient in $\delta^{18}\text{O}$ is

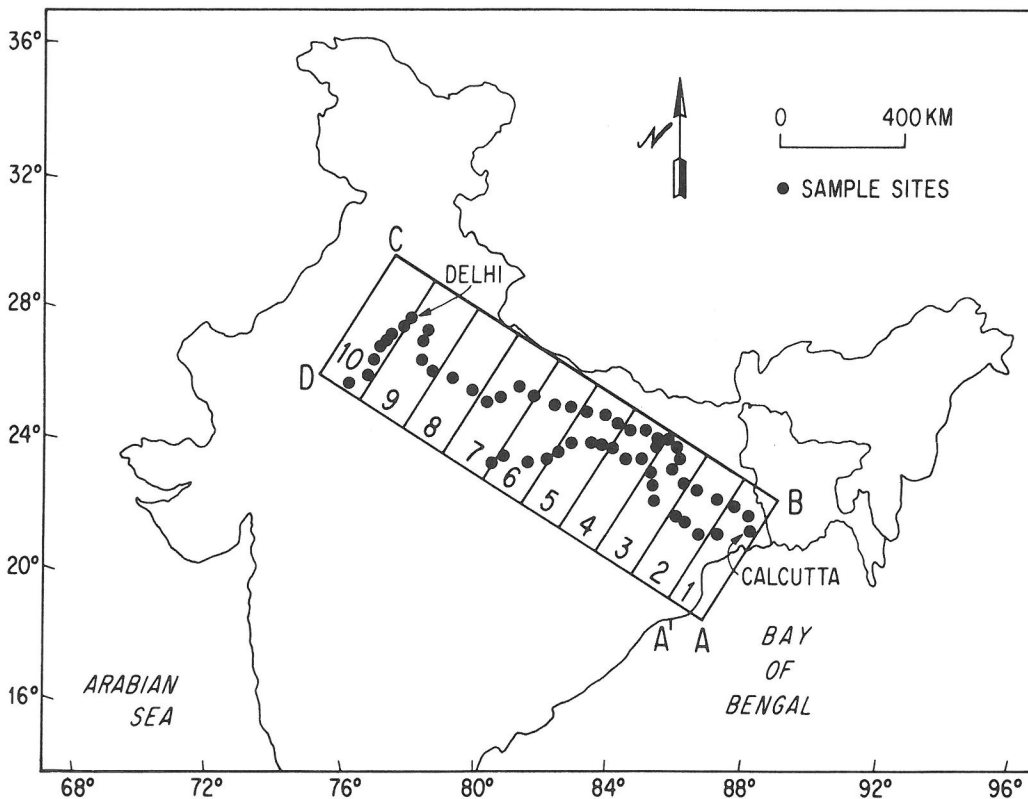


FIG. 1. Map of part of India showing the ground water sample sites. Each filled circle represents one sampling location. The rectangular box ABCD and the sub-boxes (1-10) were constructed to model the contribution of evapotranspiration. See text for details.

found to be -2 per mil per 1000 km. This is comparable to an inland gradient in $\delta^{18}O$ of -2.8 per mil per 1000 km observed in the analysis of over 900 samples of municipal water supplies of 480 stations in West Germany (FORSTEL and HUTZEN, 1983). An inland gradient of -3 per mil per 1000 km has been reported by SONNTAG *et al.* (1976) for the precipitation in the European continent. However, this $\delta^{18}O$ change with respect to distance is much less than the -6 per mil per 1000 km seen for the shallow ground water samples along a traverse from Ahmedabad to Delhi in the northwestern part of India (BHATTACHARYA *et al.*, 1985). One factor that can reduce the inland gradient is the recycling of the vapor mass through processes such as evapotranspiration. This has the effect of restoring the vapor content of the airmasses and increasing its isotopic composition over what it would normally be. Apparently, recycling of water vapor within the basin is responsible for the low $\delta^{18}O$ inland gradient of -0.75 per mil per 1000 km in the Amazon basin (SALATI *et al.*, 1979).

δD - $\delta^{18}O$ relationship

Figure 3 shows the δD - $\delta^{18}O$ relationship for the ground water samples analysed. The relationship when *all* the samples are considered is

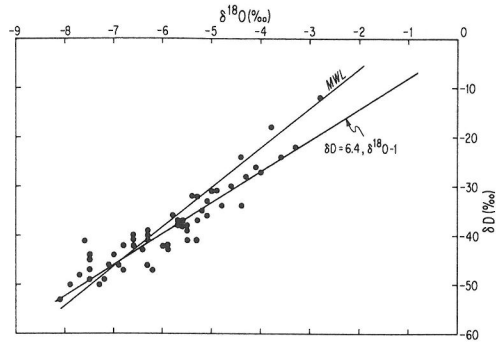


FIG. 3. The δD - $\delta^{18}O$ relationship in the ground water samples. The meteoric water line (MWL) is also shown for comparison.

$$\delta D = 6.4 \delta^{18}O - 1 \quad (r = 0.95, n = 55).$$

BHATTACHARYA *et al.* (1985) have found a δD - $\delta^{18}O$ relationship given by

$$\delta D = (6.8 \pm 0.1) \delta^{18}O + (2.2 \pm 0.4)$$

by taking into consideration the analysis of ground waters from several other stations, including the ones reported here. However, if only the samples

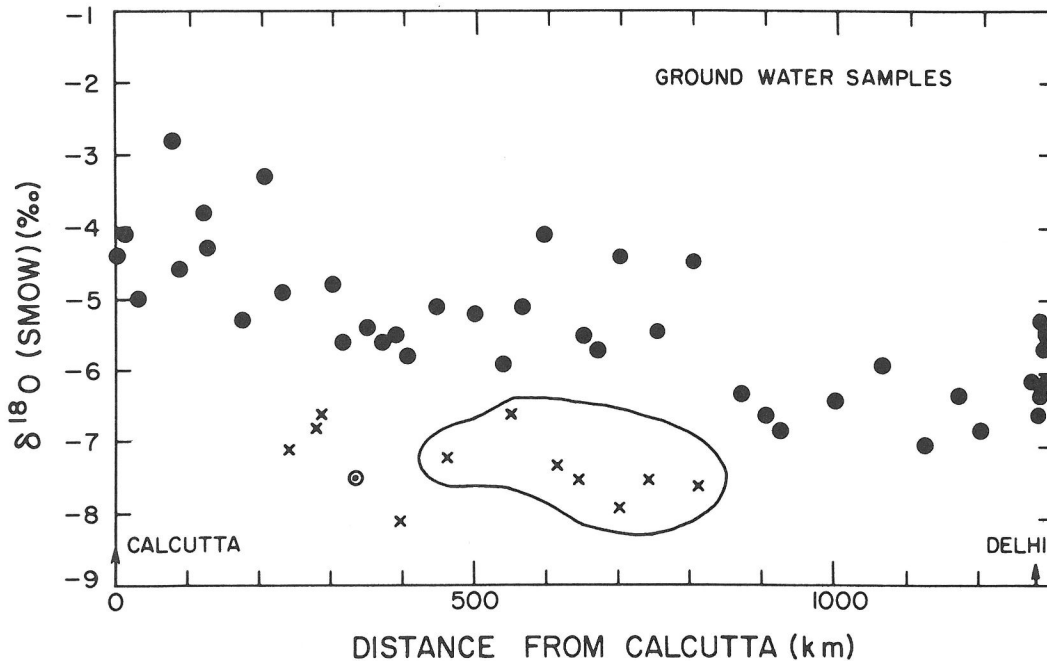


FIG. 2. The $\delta^{18}O$ values of the ground water samples along the Calcutta-Delhi traverse. The zero point is set at Calcutta on the eastern coast where the vapor mass originates and propagates inland towards Delhi. The points other than the filled circles represent sampling locations that are presumably influenced by large artificial irrigation systems, rivers, etc. They may thus have contribution other than local precipitation.

indicated in Fig. 2 by filled circles are used, because of the likely mixed sources of the other ones, the relationship is found to be (Fig. 4)

$$\delta D = 7.12 \delta^{18}O + 3 \quad (r = 0.92, n = 43).$$

It is interesting to note that this relationship is quite similar to that calculated by BHATTACHARYA *et al.* (1985) for the limited number of direct precipitations measured by the International Atomic Energy Agency (IAEA) for the stations in New Delhi, Bombay, and Shillong. They found the relationship to be

$$\delta D = (7.2 \pm 0.1) \delta^{18}O + (5.1 \pm 0.1).$$

This similarity adds strength to the argument that the δ values of the shallow ground water samples along the Calcutta-Delhi sector, ignoring those with anomalous values, quite reasonably represent an averaged precipitation in the region. Therefore, the value of ~ 7.2 can be taken as characteristic of the δD - $\delta^{18}O$ slope for the average Indian monsoonal precipitation. While this relationship deviates slightly from the well-known Meteoric Water Line (MWL) δD - $\delta^{18}O$ relationship (CRAIG, 1961), it should be noted that correlation lines with slope less than eight are not rare, especially on a regional scale where local effects play an important role. For example, GAT and DANSGAARD (1972) found a slope close to the "evaporation line" in their study of water samples from Israel and Jordan. In continental Europe the slope of the δD - $\delta^{18}O$ line has been reported to be 7.6 and 6.0 for the winter and summer precipitations, respectively (ROZANSKI *et al.*, 1982). Even in tropical island stations such as Barbados, Canton, Johnston, and Wale, the least-square fit linear relationship has a slope of 6.17.

Role of evapotranspiration in the Indian monsoon

The distribution of oxygen and hydrogen isotope ratios in precipitation along the sector Calcutta-Delhi as inferred from an analysis of shallow ground waters can be explained by treating the rain-out process in this sector as a Rayleigh-type distillation. However, the $\delta^{18}O$ change with distance or the $\delta^{18}O$ gradient inland is less than what one might expect for a pure Rayleigh-type distillation. We arrive at this by following the methods used by ROZANSKI *et al.* (1982) for interpreting the stable isotope composition of European precipitation. The sector under investigation is approximated by a rectangle and divided into ten boxes of equal size (Fig. 1), with each box measuring 350×140 km. The construction of the rectangle is mainly dictated by the disposition of the monsoon trough and that of the

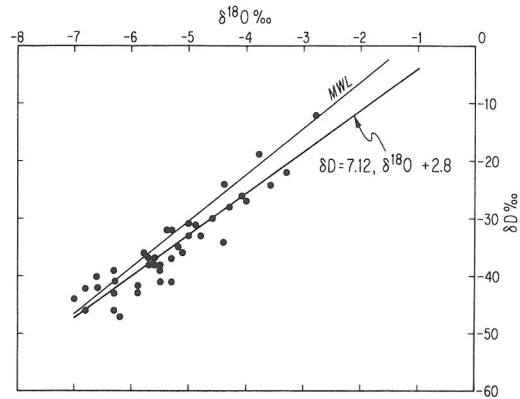


FIG. 4. The δD - $\delta^{18}O$ relationship in the ground water samples obtained after removing those with multiple contribution (irrigation systems, rivers, etc). These latter samples are shown in Fig. 2 by symbols other than filled circles.

boxes by the need to distribute the sample locations and meteorological stations as uniformly as possible. An assumption is introduced that the atmospheric circulation over the rectangular area is essentially zonal in character. This simplifies the situation by restricting the calculation of only the vapor flux entering the first box, *i.e.* along the boundaries AB and AA'. The vapor flux across the above boundaries can be calculated using the following relation (DEWAN and DATTA, 1976):

$$W = t/g \int_{p_{1000}}^{p_{500}} (L_i U_i X_i) dp \quad (2)$$

where

W = vapor flux

t = time period for which the flux is to be calculated

g = acceleration due to gravity

L_i = the distance across which the flux enters *i.e.* AB and AA'

u_i = wind component normal to the sides AB and AA'

x_i = the humidity mixing ratio.

The integration is performed between the heights 1000 and 500 mb since the moisture content above the 500 mb level is negligible. Adopting values of 5 m/sec for u_i and 12.3 g/kg for x_i (RANGARAJAN and MANI, 1982; DESAI and SUBRAMANIAN, 1978) the value of W, the vapor flux entering the first box turns out to be 1.57×10^{12} tons per season (June to September). It must be noted that the value of the flux is strongly dependent on the values used for u_i and x_i . We feel nevertheless that the value obtained by us is quite reasonable. This quantity

of vapor flux entering the first box is allowed to undergo a Rayleigh distillation so that condensation takes place in each box by successive removal of the initial vapor mass. Equation (1) will then govern the isotopic ratio of the system. We can slightly modify Eqn. (1) to accommodate a varying α, the fractionation factor from box to box. The modified expression for the isotopic composition of the condensate, *i.e.* rain at a distance from the coast, can be written as

$$R_{p(i)} = \alpha_i R_{(i-1)} F_i^{n_i-1} \quad (3)$$

where

$R_{p(i)}$ = isotopic ratio of the precipitation in the *i*th box

$R_{(i-1)}$ = isotopic ratio of the vapor entering the *i*th box

F_i = The vapor reduction factor within a single box

α_i = the fractionation factor in the *i*th box as applicable at temperatures at the average cloud base at 880 mb level.

Equation (3) also allows a step-wise development of *R* from box to box. In a step-wise treatment of *R* from box to box, proper amount of vapor can be put back by evapotranspiration before it goes into the next box so that the R_{2-1} term includes the contribution of evapotranspired water as well. A simplifying assumption is to be made that the evapotranspired component is isotopically identical to the fraction left behind as precipitation (ZIMMERMAN *et al.*, 1967).

The remaining water vapor fraction $F(n)$ in box number *n* is calculated using the equation

$$F(n) = \frac{W - \sum_{i=1}^n (P_i - E_{i-1})}{W} \quad (4)$$

where

W = water vapor flux into the first box

P_i = the precipitation flux in the *i*th box

E_{i-1} = the evapotranspiration flux in the (*i* - 1)th box.

The value of P_i was estimated for each box using the rainfall data recorded by the meteorological stations lying in the *i*th box. Table 1 shows the F_n , P_i , and α_i values for each of the ten boxes without taking into account contribution by evapotranspiration, *i.e.*, setting the F_i terms in Eqn. (4) to zero. The initial isotopic composition of the vapor in box 1 was set so that the condensate in the first box will have a δ¹⁸O value of -4.4, the value for the ground

Table 1. The precipitation (P_i), fraction remaining in vapor (F_n) and the fractionation factor for each of the ten boxes. P_i and F_n are for conditions of no evapotranspiration

Box No.	P_i (10 ¹² tons/season)	F_n	α_i
1	0.05	0.968	1.00970
2	0.05	0.936	1.00966
3	0.049	0.905	1.00962
4	0.047	0.875	1.00958
5	0.047	0.845	1.00954
6	0.048	0.814	1.00950
7	0.042	0.788	1.00946
8	0.035	0.766	1.00942
9	0.038	0.741	1.00938
10	0.031	0.721	1.00934

water in Calcutta and its neighborhood. Figure 5 shows the isotopic composition in precipitation, due to a Rayleigh process, as a function of distance from Calcutta for three initial vapor fluxes entering the first box. The three fluxes—1.19, 1.57, and 1.95 × 10¹² tons per season—allow for an error in the estimate of the initial flux by ±25%. Also shown by the straight line is the best fit line of the experimental δ¹⁸O data of ground water samples. Clearly, a pure Rayleigh-type distillation does not allow for the observed trend. The line based on experimental δ¹⁸O

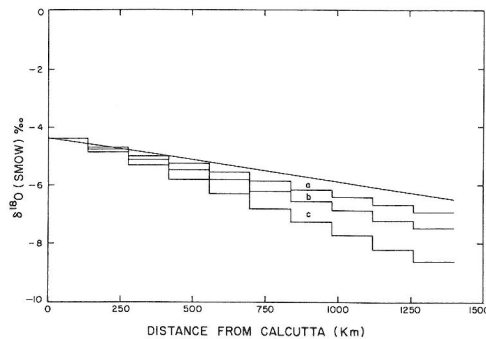


FIG. 5. The measured δ¹⁸O values in ground water samples and that estimated due to a pure Rayleigh-type distillation as the vapor mass travels from coastal Calcutta towards inland Delhi. The histograms a,b,c would be the isotopic trend in local precipitation for an initial seasonal vapor mass, respectively, of 1.95, 1.57, and 1.19 × 10¹² tons entering Calcutta and then being lost as precipitation in each of the ten boxes shown in Fig. 1. The fraction of vapor remaining in each box after removal of the condensate is given in Table 1 and was calculated using the actual rainfall data available from the meteorological stations within each box. The straight line represents the measured δ¹⁸O values in the ground water samples. If only a pure Rayleigh-type distillation operated, the experimental line would be expected to pass through one of the histograms.

values behaves as though the vapor reduction factor is offset by additional contribution to the overlying vapor mass. As stated earlier, one process that is likely to do so is evapotranspiration. Therefore, the P_i values obtained for each box may be regarded as the residual left after what has been put back by evapotranspiration.

A series of calculations can be done to estimate the fraction of the condensate in each box that needs to be put back into the overlying vapor mass so as to obtain a reasonable agreement between the calculated and observed $\delta^{18}\text{O}$ values. Figure 6 shows the results of these calculations, where it is seen that if a 40% evapotranspiration for each box is allowed, the agreement between the calculated and observed trend is pretty good. The figure of 40% compares well with a similar one quoted for the north Indian region by DATTA and DEWAN (1975) and RAMA (1980). This is also comparable with the 35% estimate for European precipitation (ROZANSKI *et al.*, 1982). It is significant, too, that the area considered here is in the fertile Ganges basin, which is abundant with vegetation and thus can be very conducive to the process of evapotranspiration.

Although an estimate of the evapotranspiration flux that is comparable with the existing estimates for this region could be made, it must be emphasized that this figure is dependent, as far our model is considered, on the two parameters W and P_i in the equations utilised. Of these, W , the initial vapor flux entering the first box, is sensitive to the wind speed that enters into the calculation. The values adopted by us are those we have been able to obtain from available information. The closer these two parameters are to the actual values, the greater will be the precision assignable to the fraction that undergoes evapotranspiration. The potential for studies of this nature in the subcontinent is clearly evident.

CONCLUSIONS

Stable oxygen and hydrogen isotope ratio of shallow ground waters along the track Calcutta-Delhi in northern India appears to be a good proxy indicator of the isotopic composition of local precipitation. Since this track is well served by the Indian monsoon, such a study can be used to understand more about the isotopic distribution in precipitation in a monsoonal system. The present study, admittedly of limited coverage, suggests that the rain-out process in this traverse can be regarded as a Rayleigh-type distillation. The Rayleigh distillation is apparently modified by additional vapor introduction to the cloud system through evapo-

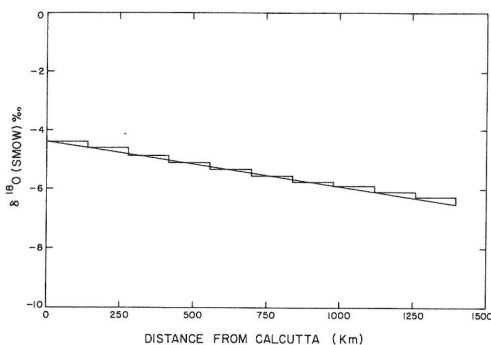


FIG. 6. The measured ground water $\delta^{18}\text{O}$ values (straight line) and the calculated $\delta^{18}\text{O}$ values (histogram) of precipitation taking into account a 40% vapor feed back through evapotranspiration. The initial vapor mass entering Calcutta is set at 1.57×10^{12} tons per season (June-September).

transpiration from the ground or vegetation cover. A simple box model can be used to estimate the fraction that is added to the overlying vapor mass by this process. If the various input parameters of the model are accepted, the evapotranspiration flux is estimated to be 40%, a figure comparable to the previous estimates.

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