4 OBSERVED ISOTOPE EFFECTS IN PRECIPITATION

In this chapter data are reported on the isotopic composition of precipitation. In principle variations in $^{18}\delta$ and $^2\delta$ are coupled, one way or the other: under conditions of isotopic (and thus also physical) equilibrium according to the meteoric water line with slope 8 (the GMWL), if the conditions are non-equilibrium by a more complicated kinetic process. For less detailed hydrological surveys it is generally assumed that $^{18}\delta$ and $^2\delta$ values are coupled as if in equilibrium. In the next sections we are therefore also reporting data from literature that are concerned with only $^{18}\delta$ or $^2\delta$.

However, it should be emphasised that measuring the oxygen as well as hydrogen isotopes often presents additional information; the combination of $^{18}\delta$ and $^2\delta$ is to be preferred.

4.1 THE LATITUDE / ANNUAL TEMPERATURE EFFECT

To explain in brief and numerically the stable isotopic composition of precipitation on a global scale, we have applied the Rayleigh model in Chapter 3, including two processes:

1) the formation of atmospheric vapour by evaporation in regions with the highest surface ocean temperatures

2) the progressive condensation of the vapour during transport to higher latitudes with lower temperatures.

The first step has been discussed in Sect.3.2. Atmospheric $^{18}\delta$ values were reported in Sect.3.4. The progressive rainout process based on the Rayleigh fractionation/condensation model, as shown in Fig.3.8 and discussed in Sect.3.5.1 results in a relation between the observed annually averaged $^{18}\delta$ and $^2\delta$ values of the precipitation and the mean surface temperatures in reasonable agreement with the observed values from the world-wide GNIP data network (Fig.4.1). The latitude effect is about

$$\Delta^{18}\delta \approx -0.6\%/\text{degree of latitude} \quad (4.1)$$

for coastal and continental stations in Europe and the USA, and up to $-2\%/\text{degree of latitude}$ in the colder Antarctic continent.

The observed relation between monthly temperature and isotopic composition shows much scatter and is not linear, except in the far north. The correlation improves by taking the
amount-weighted means. Based on north Atlantic and European stations from the GNIP network (Sect.3.1), Yurtsever (1975) reported the relation (t is the surface temperature in °C):

\[
^{18}\delta = (0.521 \pm 0.014)t - (14.96 \pm 0.21) \, \% \tag{4.2}
\]

This compares reasonably well with the relation originally reported by Dansgaard (1964) for the north Atlantic stations:

\[
^{18}\delta = 0.695t - 13.6 \, \% .
\]

The simple model calculation for both \(^2\delta\) and \(^{18}\delta\) leads to the equation for the MWL as discussed earlier and is also reasonably consistent with Fig.4.1. The relation is further improved if the ground temperature is replaced by the cloud-base temperature (Rindsbergen and Magaritz, 1983). It should be emphasised again that a global precipitation model can only give reliable results, if seawater evaporation is included over the entire range of seawater temperatures instead of being limited to the "thermal equator".

\[
^{18}\delta = 0.52t - 15\, \% \tag{acc. Fig.3.9}
\]

**Fig.4.1** The *latitude or annual-temperature effect* on \(^{18}\delta\) of precipitation (pcpt): weighted \(^{18}\delta\) of total precipitation over periods of at least one decade from marine/coastal, continental and (Ant)arctic stations are shown as a function of the mean measured surface air temperature (data from the GNIP network). The (t,\(^{18}\delta\)) relation as reported by Yurtsever (1975) (Eq.4.2) and that according to Fig.3.9 are given for reference.

On the average global scale the (\(^{18}\delta\), \(^2\delta\)) relation turns out to be satisfactorily described by the *Global Meteoric Water Line (GMWL)*, as is mentioned in Sect.3.1 (Fig.4.2). Regionally and
for certain periods (such as seasons) *Local Meteoric Water Lines* (LMWL) may be found, depending on the conditions for forming the local water source of each region (see next Section).

![Graph showing relation between weighted average $^{18}\delta$ and $^2\delta$ values of total precipitation over periods of at least one decade from the same stations of the GNIP network as represented in Fig.4.1. The data confirm the general validity of the Global Meteoric Water Line (GMWL) reasonably well.]

**Fig.4.2**  Relation between weighted average $^{18}\delta$ and $^2\delta$ values of total precipitation over periods of at least one decade from the same stations of the GNIP network as represented in Fig.4.1. The data confirm the general validity of the Global Meteoric Water Line (GMWL) reasonably well.

### 4.2 SEASONAL EFFECT

A seasonal temperature pattern is clearly followed by all but some marine stations (Fig.4.3). The dependence of the isotope variations on the local temperature (or the closely related parameter of the precipitable water content (Sonntag et al., 1983)) appears as the overriding parameter (Yurtsever, 1975; Fricke and O'Neil, 1999). Generally the temperature dependence of both $\delta$ values is smaller than shown by the latitude effect, varying from about $0.5\%/$°C for some higher-latitude stations to ultimately $0\%/$°C for tropical ocean islands (Fig.4.3 and 4.4).

From the simple model presented in the preceding section follows that the seasonal variations of $^2\delta$ and $^{18}\delta$ values of precipitation are bound to be smaller than the variations of yearly averages with latitude. The essence is in the $(1/T - 1/T_0)$ factor in Eq.3.16. During winter, the condensation temperature $T_L$ is lower than during summer. However, the same is true for $T_0$. 

\[
y = 7.8973x + 9.0133
\]
the prevailing average temperatures in the region of evaporation. This partly masks the local temperature effect.

![Graph showing the relationship between temperature (t) and isotopic composition (δ18O)](image)

**Fig. 4.3** Comparison of the latitudinal and the seasonal temperature effect of $^{18}$O in precipitation. In both cases the data show the averages over a large number of years. The dashed line refers to the latitudinal effect given by Eq. 4.2, heavy dots represent the long-term annual averages, smaller dots the monthly averages. Numbers indicate the following stations:

1. Midway, Pacific Ocean
2. Pretoria, South Africa
3. Valentia, Ireland
5. Groningen, The Netherlands
6. Vienna, Austria
7. Ottawa, Canada

The opposite seasonal effect for Pretoria may be correlated with high precipitation intensity with often low $^{18}$δ values during summer (cf. Sect. 4.5).

For instance, during winter in the northern hemisphere the highest temperatures are found south of the equator. Water vapour from this region is not easily transported to the northern hemisphere, however, so the effective evaporation region for the northern hemisphere has a lower $T_0$ value than during summer.
Observed Isotope Effects

A

Northern Hemisphere

Midway
Valentia
Tokyo
Groningen
Vienna
Ottawa
Wynyard

Monthly average pcpt

B

Southern Hemisphere

St.Helena
Brisbane
Pretoria
Stanley
Vernadsky

Monthly average pcpt
Stations situated in a mid-continental setting typically portray a seasonal change in the isotopic composition of the precipitation. These variations are correlated with the temperature, in most cases. At tropical islands, on the other hand, where the vapour source region essentially coincides with the region of precipitation ($T=T_0$) the temperature dependence almost disappears, as is shown in Fig.4.3 and 4.4.

If the weighted averages of monthly precipitation over a large number of years is considered, the ($^{18}\delta, ^{2}\delta$) relation is very close to that of the Global Meteoric Water Line (Fig.4.5). However, looking in detail deviations exist.
The most pronounced factors which determine the offset of monthly data at any station from a classical meteoric water relationship are:

1) different source characteristics of the moisture, either due to the seasonal change of the meteorological conditions over the ocean, or different location of the source regions. This basically fixes a series of parallel meteoric water lines, for each of the seasons.

2) evaporative enrichment in the falling droplets beneath the cloud base, effective during warm and dry months when rain amounts are small. This partially evaporated rain is characterised by relatively higher $^{18}\delta$ values and small to negative d-excess values.

3) high values of the d-excess parameter (Fig.4.4) associated with snow or hail (Jouzel and Merlivat, 1984). These events are also associated with very depleted isotopic values.

Fig.4.5  Seasonal influence on the ($^{18}\delta$, $^{2}\delta$) relation for average monthly precipitation at a number of stations, arctic, tropical, coastal and continental (data from the same series as in Fig.4.4).

The changes in $^{18}\delta$ are correlated with those of $^{2}\delta$. However, a local best fit for the ($^{18}\delta$, $^{2}\delta$) line is in most cases not a meteoric water line in the sense described before, namely one produced by varying degrees of rainout from an air mass of prescribed isotopic character. By and large, most of these effects combine to produce a low-slope ($^{18}\delta$, $^{2}\delta$) line (i.e. with a slope less than 8) which is not only indicative of the genetic and synoptic history of the rain events,
but also reflects the local conditions at the time precipitation occurs. Notably when the slope \( \Delta^2\delta/\Delta^{18}\delta \) differs from the value of 8, then a simple linear fit is often not a satisfactory description of the relationships, because a variety of processes is at play, each process with its own set of rules concerning the isotope fractionation involved.

### 4.3 OCEANIC AND CONTINENTAL PRECIPITATION

Precipitation over the ocean, collected at island stations or weatherships, has the characteristics of a first condensate of the vapour. The range of most \( ^{18}\delta \) values is relatively small, between \(-0\%\) to \(-5\%\) with but little seasonal change in many cases (see Sect.4.2), and a lack of a clear correlation with temperature (Dansgaard, 1964; Rozanski et al., 1993). The same is true for the \( ^2\delta \) values. There is a relatively large variability in the value of the Deuterium-Excess especially notable close to the major source regions of the atmospheric moisture. This is obtained by imagining MWL's (having slope 8!) drawn through the data in Fig.4.4).

For comparison, Fig.4.4 also contains \((^{18}\delta, ^2\delta)\) data from inland or continental stations. In the case of oceanic precipitation, the scatter in d-excess predominates, whereas the isotopic data are fairly well aligned along meteoric water lines further removed from the coast, so that, even though the range of \( ^{18}\delta \) is much larger than in the marine domain, the values of \( \sigma(d) \) are actually reduced. This characteristic can be quantified by the ratio of the spread in the d-excess, \( \sigma(d) \), to the range of \( ^{18}\delta \) values, \( \sigma(d) / <^{18}\delta> \). This ratio changes, for example over the European continent, from 2.0, 2.3, 1.8 and 2.8 at the coastal and island stations of Valentia (Ireland), Reykjavik, Faro (Portugal) and Weathership E in the northern Atlantic to values of 0.4, 0.35 and 0.7 at continental stations such as Berlin, Krakow and Vienna, respectively.

This observed pattern fits quite well a simplified scheme for the isotopes in the hydrologic cycle, wherein one views the d-excess value to be established at the site of the air-sea interaction. The offset from equilibrium conditions is then determined primarily by the humidity deficit above the sea surface, i.e. the value of \((1-h_N)\) (Merlivat and Jouzel, 1979). As will be discussed below, the d-excess value is basically conserved during the rainout over the continents.

Some apparently anomalous features of the marine data set (namely some relatively depleted isotopic values) can be explained, on the one hand, by the intense vertical mixing in the air column of tropical clouds in the presence of the ITCZ, as shown above. The amount effect, i.e. a correlation of the depletion of heavy isotopes with the amount of rain is explained on the other hand by the preferential isotopic exchange of the smaller droplets, which are predominant in light rains and drizzle, with the near-surface moisture. Heavier rains on the other hand maintain the depleted isotopic values from within the clouds.
Observed Isotope Effects

\[ y = 6.7997x + 6.418 \]

\[ y = 6.7976x + 1.4027 \]

\[ y = 6.8652x + 4.7017 \]

\[ y = 7.3572x + 6.1175 \]

\[ y = 7.2078x - 0.3353 \]

\[ y = 7.4903x + 5.6249 \]

\[ y = 5.548x + 6.3181 \]
Fig. 4.5  Regression lines for (\(^{2}\delta, {^{18}}\delta\)) relations of monthly precipitation samples from marine stations, continental tropic and (Ant)Arctic stations (data from GNIP); d-excess values to be estimated from lines with slope 8 through the data points, are highly variable, in agreement with Fig.4.4.
The \textit{continental effect}, also referred to as the \textit{distance-from-coast effect}, i.e. a progressive $^{18}$O depletion in precipitation with increasing distance from the ocean, varies considerably from area to area and from season to season, even over a low-relief profile. It is also strongly correlated with the temperature gradient and depends both on the topography and the climate regime.

During the passage over Europe, from the Irish coast to the Ural mountains, an average depletion of 7‰ in $^{18}\delta$ is observed. However, the effect in summer is only about one fourth of the effect in winter. Eichler (1964) attributed this to the re-evaporation of summer rain.

An extreme case of the absence of an inland effect over thousands of kilometres, in spite of strong rainfalls en route, was reported over the Amazon (Salati et al., 1979). This is also attributed mainly to the return flux of the moisture by (non-fractionating) transpiration and thus invalidates the effect of the rainout. However, some of the return flux apparently occurs by evaporation from open waters. This process then results in some change in the isotopic composition, and in particular, an increase of the d-excess. Such an increase is indeed noted in the comparison of the precipitation from inland to coastal stations.

A similar effect of an increase of the d-excess will also result from the partial re-evaporation of rain droplets beneath the cloud base as these fall to the ground surface, as described by Dansgaard (1964) in the case of precipitation at Adis-Abeba in north-east Africa.

The continental effect in $^{2}\delta$ is nicely shown by the iso-$^{2}\delta$ contours reported by Taylor (1972) for the United States. However, from an evaluation of the IAEA data, it turns out that there are continental stations without a \textit{continental effect}. Many results for coastal stations appear to deviate from the lines in Fig.4.1, whereas several continental stations have precipitation in agreement with these ($t$, $^{18}\delta$) relations. The extent to which a continental effect occurs probably depends on the prevailing direction of the movement of air masses, rather than simply on distances from the ocean.

4.4 \textbf{ALTITUDE EFFECT}

As a rule the isotopic composition of precipitation changes with the altitude of the terrain and becomes more and more depleted in $^{18}$O and $^2$H at higher elevations. This has enabled one of the most useful applications in isotope hydrology, namely the identification of the elevation at which groundwater recharge takes place.

This \textit{altitude effect} is temperature-related, because the condensation is caused by the temperature drop due to the increasing altitude. Due to the decreasing pressure with increasing altitude (-1.2‰/100 m), a larger temperature decrease is required to reach the saturated water vapour pressure than for isobaric condensation. Therefore, $dN_V/N_V$ per °C and thus $d\delta_V/dT$ is smaller than for the isobaric condensation process which produces the latitudinal effect. The molar amount of vapour is proportional to the barometric pressure, $b$. The relative decrease in vapour content is then given by:
\[ \frac{dN_V}{N_V} \approx \frac{dp}{p} - \frac{db}{b} \]

Siegenthaler and Oeschger (1980) calculated this decrease to be -3.6%/100 m (directly caused by a temperature drop of -0.53°C/100 m and +1.2%/100 m, respectively. The value of \((dN_V/N_V)\)dT is then 2.4/0.53 = 4.5%/°C. From Eqs.3.16 and 3.17 we can then deduce the temperature effect:

\[ \frac{d^{18}\delta}{dT} +0.4‰°C \approx -0.2‰/100m \]

and

\[ \frac{d^2\delta}{dT} +3‰°C \approx -1.5‰/100m \]

However, other factors need to be considered that change the isotopic composition, besides the basic Rayleigh effect. One is the evaporative enrichment of \(^{18}O\) and \(^2\)H in raindrops during their fall beneath the cloud base, which is larger at low altitudes where the cloud base is typically high above ground level. This so-called pseudo-altitude effect (Moser and Stichler, 1974) is observed in inter-mountain valleys and on the lee side of a mountain range. This evaporative enrichment, unlike the primary Rayleigh rainout effect, also results in a decrease of the d-excess and thus marks these situations clearly. This effect is illustrated for the case of a traverse across the Judean mountain range and in the distribution of the d-excess parameter throughout the area, closely following the topography (Gat and Dansgaard, 1972).

**Fig.4.7**  Example of the altitude effect on precipitation for the eastern slopes of the Andes mountains, as deduced from samples of undepth groundwater/soilwater, collected from springs. The magnitude of the effect is increasing from -0.2 to -0.6‰/100m (Vogel et al., 1975).
The most elusive factor is where different air masses with different source characteristics affect the precipitation at the base and crest of a mountain. A prominent case is that of the western slopes of the Andes in South America: precipitation near the crest results predominantly from air from the Atlantic with a long continental trajectory, whereas air from the Pacific Ocean, with predominantly oceanic attributes, affects the precipitation in the lower elevations. Under such conditions one encounters apparently anomalously large altitude effects.

An opposite effect results when the long-term samples (e.g. monthly, seasonal or annual composites) represent different time series at the varying altitudes. As an example can be cited the case of a low-altitude and a high-altitude station in Cyprus (Gat et al., 1962). Frontal rains affect both stations and show a normal altitude effect. In the mountain, however, rain also occurs orographically at times that no rain falls in the plain, so that in the composite sample of the mountain station the frontal rains are diluted by precipitation representing a 'first condensate', with relatively enriched isotopic values.

The observed $^{18}$O effect generally varies between $-0.1\%$ and $-0.6\%/100\text{ m}$ of altitude often decreasing with increasing altitude (Vogel et al., 1975) (Fig.4.7). Values in this range have also been reported for mountain regions in Czechoslovakia (Dinçer et al., 1970), Nicaragua (Payne and Yurtsever, 1974), Greece (Stahl et al., 1974), Cameroon (Fontes and Olivry, 1977), Italy (Bortolami et al., 1978) and Switzerland (Siegenthaler and Oeschger, 1980).

Friedman et al. (1964) reported data on the altitudinal effect on $^2\delta$, showing roughly $-4\%/100\text{m}$ for the coastal region of the western United States, whereas Moser et al. (1978) report a value of $-2.5\%/100\text{m}$, observed in S.W. Germany and values ranging from $-1\%/100\text{m}$ to $-4\%/100\text{m}$ for Chile (Moser et al., 1972).

In many cases the d-excess is found to increase with altitude, possibly for a variety of reasons. This issue has not been finally resolved.

### 4.5 AMOUNT EFFECT

Dansgaard (1964) observed a relation between the amount of precipitation and $^{18}\delta$. For example, the very strong tropical rainfalls at times of the passage of the Intertropical Convergence Zone (ITCZ), characterised by towering clouds and strong downdrafts, may be extremely depleted in $^{18}\delta$ and $^2\delta$, the former by as much as $-15\%$. Similar, though smaller effects are observed in thunderstorm-engendered precipitation. In north-western Europe during convective storms changes in $^{18}\delta$ have been found of $-7\%$ within 1 hours. Examples are shown in Fig.4.8. Probably, the dip in the $^{18}\delta$ curve for seasonal precipitation at Tokyo in June (Fig.4.4), quite consistent over several years, is due to the high rain intensity during this month. The same may be true for the opposite seasonal temperature effect over Pretoria (Figs.4.3).
In other (isolated) cases amount effects are not consistent and appear to depend critically on the meteorological conditions at the time of the precipitation. In the eastern Mediterranean some of the largest rainfall events result from air masses whose origin differs from that of the usual winter precipitation and which show rather enriched isotopic values, compared to the rest of the precipitation.

Among island stations, where temperature variations are small, a dependence of $^{18}\delta$ on rain intensity is observed to the extent of $-1.5\%_o/100\text{mm of monthly precipitation}$. It turns out that if this is taken into account, the temperature correlation is improved.

On the other hand, small amounts of rain are, as a rule, enriched in the heavy isotopes along typical evaporation lines, especially in the more arid regions. This effect obviously results from the evaporation of rain droplets on their fall to the ground. However, no further consistent amount effect is noted for rain intensities in excess of about 20\text{mm/month}.

The conclusion is that one should refrain from generalisations and explore the local amount effect individually in each case, by running a special sampling programme.

![Fig.4.8](image)

**Fig.4.8** Time sequence of the isotopic composition of precipitation during showers; examples are shown for two cases of convective storms: **A)** rain intensity in mm/2 hours (Mook et al., 1974); **B)** cumulative rain over variable periods.

### 4.6 INTERANNUAL VARIATIONS

Yearly average $^{18}\delta$ values vary from year to year. In temperate climates the values generally do not vary by more than 1\%, and a large part of the spread is caused by variations in the average annual temperature. Figs.4.9 and 4.10 present annual variations for a number of stations,
showing a certain spread of the weighted annual precipitation $^{18}\delta$ as well as $^2\delta$. In semi-arid climates, with a less regular rain distribution in time, larger variations occur. In those cases only hydrological systems that pool precipitation inputs over many years can be related to the average rain input over many years.

![Graph showing weighted average annual values of $^{18}\delta$ in precipitation (pcpt) vs the mean surface air temperature, showing the variations from year to year (data from the GNIP network).](image)

**Fig.4.9** Weighted average annual values of $^{18}\delta$ in precipitation (pcpt) vs the mean surface air temperature, showing the variations from year to year (data from the GNIP network).

### 4.7 SMALL-SCALE VARIATIONS

#### 4.7.1 SMALL-SCALE SPATIAL VARIATIONS

It is important to establish, for instance for $^{18}\text{O}/^{16}\text{O}$ tracer studies in which $^{18}\delta$ or $^2\delta$ variations in rain are related to those in run-off, that short- and long-term variations of $^{18}\delta$ of precipitation with distance have not occurred over the region of interest.
In the temperate climate of north-western Europe it has regularly been observed that the $^{18}\delta$ values of rain samples collected over periods of 8 and 24 hours from three locations within 6 km$^2$ at equal elevations agree within 0.3\%/oo (Mook et al., 1974). Similar results were obtained from the semi-arid climate of Israel (Rindsberger et al., 1990).

At larger distances, especially with single convective storms, larger differences are to be expected. Over a period of a month $^{18}\delta$ in rain has a similar pattern, even over distances of a few hundred kilometres. Despite the similarities, single monthly samples might differ significantly, although the average values are only slightly ($\approx 1\%/oo$) different. The $^2\delta$ and $^{18}\delta$ values probably closely obey the MWL.

### 4.7.2 SMALL-SCALE TEMPORAL VARIATIONS

Rapid variations in $^{18}\delta$ of rain can occur. In Sect.3.1.7 we have mentioned such variations in samples collected during successive two-hour periods. The magnitude of this effect depends on
the character of the precipitation process such as convective storms or weather fronts. Fig. 4.8 showed an example, in the framework of the amount effect.

![Graph showing isotopic composition of modern and palaeowaters](image)

**Fig. 4.11** Isotopic composition of modern meteoric waters and of palaeowaters in the eastern Mediterranean Sea area. The average monthly precipitation are from Ankara, Vienna and Beit-Dagan; the numbers 1-12 refer to calendar months. Furthermore, the range is indicated of groundwater sources of the modern and palaeohydrologic cycle in the Levant; data from palaeowater of the western desert (Egypt) are shown for comparison (Gat, 1983).

### 4.8 PALAEOCLIMATE RECONSTRUCTION

The good correlation established between the isotopic composition of precipitation at many sites (especially in a continental setting) and climate variables, such as the ambient temperature and amount of precipitation (Yurtsever, 1975), suggested the use of the isotopic signal both as a monitor and record of climate changes, i.e. changes in the temperature and meteorological regimes.
The 30-year time series of the GNIP program has enabled one to identify a trend of increase in $^{18}O$ in parallel with a temperature increase in continental stations in Europe (Rozanski et al. 1992). Shorter term fluctuations in the isotopic composition of precipitation have been shown to coincide with changes in the synoptic pattern of the rain producing air masses, for instance for the Mediterranean, contemporaneous with a cold phase in northern Europe (Kukla and Kukla, 1974; Gat and Carmi, 1987).

Longer records of isotopes in meteoric waters are the ice accumulation on glaciers, where the record can be traced back for thousands of years (see Volume I: Sect.7.2.5).

Meteoric waters originating from precipitation in the distant past, named palaeo-waters (Fontes, 1981) are encountered as deep groundwater in those areas where the present-day replenishment of groundwaters is slow, predominantly in arid ones. In most cases their isotopic composition differs appreciably from that of the modern precipitation in the recharge zone. Fig.4.10 shows some examples.

The isotopic composition can also be preserved in materials related to the meteoric water cycle which are more ubiquitous than the water itself. Such materials, named climate proxies are:

- interstitial waters in lake sediments
- carbonates and silicates in lake sediments
- concretions in soils and stalactites
- cellulose in treerings and other plant material
- phosphates in biomass
- peat bogs.

In some of these proxies, e.g. cellulose and interstitial waters, both $^{18}O$ and $^2H$ variations can be measured. In other cases, only the oxygen isotopic composition is amenable to measurement.

Based on the changes in the isotopic composition of the proxies one seeks to infer the change in the meteoric waters, so as to relate these then to climate parameters such as temperature and humidity. In order to do so two palaeoclimate effects need to be considered. On the one hand, the changing isotope composition of the precipitation (the primary climate signal) and on the other hand, there is the effect of climate change on the relationship between the isotopic composition of precipitation and that of the proxy. The latter in itself is made up of two components, namely the relationship between the isotopic composition of the incident rain and the water body which imprints its isotopic signature on the proxy material (see for example the discussion of Gat and Lister, 1995 in the case of lake sediments) and further the transfer from the relevant water body (be it lakewater, soilwater or groundwater, etc.) to the proxy material.
As discussed in section 3.5.2., the transfer of precipitation into the terrestrial aquatic systems is neither complete nor indiscriminate. As rain falls on the land surface it is partitioned into fluxes of surface runoff, infiltration into the ground and a return flux to the atmosphere by means of evaporation and transpiration. The isotopic composition is modified as a result of these processes, both due to the isotope fractionation which accompanies evaporative processes near the surface and also due to the selective utilisation of rainfall with different isotopic composition in these different pathways. Obviously this shift in the isotope composition which accompanies the recharge–runoff processes, which were named the Isotope Transfer Function (ITF) (Gat, 1998) depend on and change with changes of the ambient conditions. These shifts are especially significant under two extreme environmental situations; namely on the one hand in lake and wetland country and, on the other hand, in an arid environment. The common denominator of these two regions is the prominent role of surface waters in their hydrological cycle.

The safest way to assess the changing isotope values in precipitation during past period appears to be through the medium of global circulation models; estimates of the possible changes in the ITF and the transfer into proxy material must depend on a in-depth understanding of the processes concerned,

Summarising, there are two palaeoclimate effects that need to be considered. On the one hand, there is the effect of the changing system on the relationship between the isotopic composition of precipitation and that of the proxy, on the other hand the changing isotopic composition of the precipitation (the primary climate signal).

As far as the latter is concerned, the most probable relation between $^{18}\delta$ (or $^2\delta$) of precipitation in the region and average temperature at the time of precipitation is similar to the present-day latitudinal temperature dependence at that specific temperature (Sect.4.1). For the ice cores in polar regions the effect of temperature may be larger than $+0.5$ to $+0.7\%o/^\circ C$ as in non-arctic regions.