Yifeng Cui

Different Approaches towards an Understanding of Runoff Generation

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Preface

Twenty-eight years ago the Hydrological Research Basins in the Kaiserstuhl area have been established. Primarly the two test catchments served as experimental basins to investigate the hydrological response on large scale terracing of the wine yards. Numerous studies have been carried out investigating the influence of large scale terracing on the water balance components and the runoff generation.

The research work of Dr. Yifeng Cui deals with the application of modern techniques in particular experimental tracer methods to adress the problem of runoff generation. It is a study to the recent research on that issue. A continuous runoff separation model is combined with the application of natural tracer techniques to investigate the process of runoff generation. The results of the investigations emphasize that graphical and isotope methods have different boundary conditions. The combination of both models and experimental methods leads to a better understanding of the hydrological system.

We wish Dr. Yifeng Cui all the best in his future work as hydrologist. His profound knowledge, his friendly and modest appearance, his high engagement and enthousiasme for hydrology will be an excellent background.

Christian Leibundgut *Siegfried Demuth*

- Editors -

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Abstract

To contribute to the understanding of runoff generation, two types of runoff separation methods were used simultaneously in this study. The first one is the application of a component model based on the traditional graphical principle. The second one is the experimental investigation, with the help of oxygen-18 and hydrochemical tracers. The aim of this work is to obtain a quantitative estimate of the water which contributed from various sources during the course of the hydrological year 1995, and to obtain more catchment information, such as residence times and storage volumes.

The study sites are two neighboring small catchments built out of silty loess. The catchment areas have similar physiographical inventories. The small-terraced Rippach catchment has an area of 1.2 km² and lies in an intensively used agricultural area. The large-terraced Loechernbach catchment has an area of 1.7 km², there man-made changes have a great influence on the topclimate, the ecology of the viticulture and the soil structure.

The hydrometeorological data have been collected from more than 15 years of daily data. The experimental investigations include weekly sampling for 14 months, of streamwater, baseflow, groundwater or spring water, drainage water and precipitation. In addition, two automatic sampling devices were installed in gauge stations, in order to collect the streamwater samples in 4 hour intervals.

The investigations of individual events indicate that new water fractions are up to 90 % of the total runoff. Even in the small altered basin the event water dominates the major phase of the event. This contradicts the results reported in previous studies. In particular, a significant macropore flow in the unsaturated zone was found in the altered catchment. Macropore flow even dominated the event water 500 minutes after the start of the event, confirming its important contribution to storm runoff.

The 14 month experimental investigation revealed that of the totals, 92 % and 83 % is old water in the unaltered and altered basins respectively. The combination of oxygen-18 and chloride allowed the separation of rapid interflow. This component is negligible in the unaltered basin, while the interflow in the altered basin, increased by macropore flow via a large number of drainage subpipes, is 8 %. The 1.7 years of groundwater residence time for both basins was determined by a dispersion model, implying a total storage capacity in the order of 1085 mm and 1235 mm as an annual average. In addition, the summer infiltration coefficient of 0.68 in the altered basin is the result of the large terracing, and is low compared with the nearly balanced one of 0.98 in the unaltered basin.

With the help of the component model DIFGA, a continuous separating hydrograph model, runoff in the research catchments was separated into three and four components. Compared with the isotope results, the indirect runoff of 84 % in the unaltered basin by DIFGA model is in the same order of magnitude. However, this delayed component by DIFGA is underestimated by 19 % in the altered basin. There are various reasons for this underestimation, but non-linear storage is the main one. It is also concluded that the major reason for the failure of all models applied in the altered basin is the incorrect use of linear storage for the second component (interflow).

The storage volume of 9 mm in the altered catchment obtained by DIFGA is smaller by two orders of magnitude than that obtained by isotope data. This remarkable difference can be explained by a concept of a energy or dynamic reservoir, implying that the active mobile reservoir of this basin is small, caused by the destruction of the soil through machinery. It should be emphasized that graphical and isotope methods have very different boundary conditions. While the former explains how quickly water flows into streams, the latter explains how long water is stored within a basin. The combination of both gives the following picture: water is stored for a long time in the basin, before it is quickly displaced through piston-flow effect.

The sensitivity analysis of the two-component model suggested that the correct weighting calculation of input data is very important for long-term investigation. The use of piecewise linear baseflow, rather than individual groundwater data, is suggested to be appropriate as old water for the long-term separation. The short-term sensitivity analysis emphasized that field investigation is very important for the determination of the electrical conductivity of event water. The direct use of the value from precipitation would result in the underestimate of new water contribution by about 13 % in the present study sites. This underestimate in other basins may be higher.

Zusammenfassung

Die vorliegende Arbeit will einen Beitrag zum Verständnis der Abflußbildung leisten. In den Untersuchungen wurden zwei Methodentypen zur Trennung des Abflusses gleichzeitig angewandt. Die erste Methode besteht in der Anwendung eines Ganglinienseparationsmodells, das auf dem traditionellen graphische Prinzip aufbaut. Den zweiten methodischen Ansatz bildet eine experimentelle Untersuchung mit Hilfe von Sauerstoff-18 und hydrochemischen Tracern. Zielsetzung war es, zu quantitativen Schätzungen über dasjenige Fließwasser zu gelangen, das aus verschiedenen Quellen im Laufe des Hydrologischen Jahres 1995 zum Abfluß kam. Darüber hinaus sollten die Kenntnisse über Verweilzeiten und Speichervolumen im Einzugsgebiet verbessert werden.

Die Untersuchungsgebiete bilden zwei kleine benachbarte Einzugsgebiete im östlichen Kaiserstuhl, der sich als vulkanischer Gebirgsstock aus der südlichen Oberrheinebene erhebt. Seine Hänge sind zu großen Teilen mit Löß bedeckt. In der landwirtschaftlichen Nutzung herrscht der Weinbau vor. Hier haben Rebflurbereinigungen in den 70er Jahren dieses Jahrhunderts die neuen Landschaftsformen der Großterrassen geschaffen. Welche Auswirkungen diese Eingriffe auf die Entwicklung des Mikroklimas, der Bodenstruktur, des Abflusses und der Weinbauökologie haben, wurde in den Folgejahren kontrovers diskutiert und aus verschiedenen Forschungsansätzen heraus betrachtet. Die hier untersuchten Einzugsgebiete zeigen bis auf ihre unterschiedlichen Terrassierungsformen einen ansonsten vergleichbaren physiographischen Bestand. Das kleinterrassierte Einzugsgebiet Rippach umfaßt eine Fläche von 1,2 km^2, das durch Großterrassen umgelegte Gebiet Löchernbach eine Fläche von 1,7 km^2. Beide Einzugsgebiete liegen im Löß und werden landwirtschaftlich intensiv genutzt.

Als hydrologische Daten wurden der tägliche Niederschlag und die Abflußwerte herangezogen, die über einen Zeitraum von 15 Jahren in den beiden Untersuchungsgebieten ermittelt worden waren. Die experimentellen Untersuchungen umfaßten wöchentliche Proben von Fließwasser, Grundwasser, Drainagen und Niederschlag, die über 14 Monate hinweg kontinuierlich gezogen wurden. In den bestehenden Pegelstationen am Rippach und Löchernbach wurden zwei zusätzliche automatische Probenahmegeräte installiert, um Bachwasserproben in vierstündigen Intervallen zu erhalten.

Die Untersuchungen zu den Einzelereignissen deuten darauf hin, daß das Ereigniswasser oder "neue Wasser" einen Anteil von bis zu 90 % am Gesamtabfluß hat. Dies steht im Widerspruch zu früheren Untersuchungen. Besonders hervorzuheben ist, daß ein bedeutender Makroporenabfluß das Bachwasser noch 500 Minuten nach Ende des Ereignisses dominierte. Dies bestätigt den beachtlichen Beitrag des Makroporenabflusses zur Abflußbildung.

Die langfristigen Untersuchungen über ein Jahr zeigen auf, daß der Gesamtabfluß im kleinterrassierten Einzugsgebiet zu 92 % und im großterrassierten zu 83 % aus "altem" bzw. Voreigniswasser gebildet wird. Die Kombination der Traceranwendung von Sauerstoff-18 und Chlorid erlaubte die Trennung des schnellen Interflow. Diese Komponente konnte im kleinterrassierten Einzugsgebiet vernachlässigt werden, während sie im großterrassierten Einzugsgebiet durch den Makroporenabfluß der zahlreichen Drainagen 8 %

betrug. Die Verweilzeiten des Grundwassers von 1,71 bzw. 1,75 Jahren für die beiden Einzugsgebiete wurden mit Hilfe eines Dispersionsmodells bestimmt und weisen auf eine Gesamtspeicherkapazität in der Größenordnung von 1085 bzw. 1235 mm im Jahresdurchschnitt hin. Das Verhältnis zwischen Sommer- und Winterinfiltrationskoeffizient liegt im großterrassierten Einzugsgebiet bei einem Wert von 0,68. Das kleinterrassierte Gebiet zeigt hingegen mit einem Wert von 0,98 eine fast ausgeglichene Situation.

Durch Anwendung des DIFGA-Modelles, einem kontinuierlichen Ganglinenseparationsverfahren, wurde der Abfluß in drei bzw. vier Komponenten separiert. Verglichen mit den Ergebnissen der Isotopenuntersuchungen liegt der Anteil des indirekten Abflusses am Gesamtabfluß mit 84 % im kleinterrassierten Einzugsgebiet in einer plausiblen Größenordnung. Jedoch wird der entsprechende Wert im großterrassierten Einzugsgebiet um 19% unterschätzt. Dafür gibt es mehrere Gründe. Hauptgrund ist der nicht-lineare Speicher in diesem Einzugsgebiet.

Das Speichervolumen von 9 mm im großterrassierten Einzugsgebiet, das mit Hilfe des DIFGA-Modelles ermittelt wurde, ist um zwei Ordnungen kleiner als dasjenige, das die Analyse der Isotopendaten liefert. Dieser bemerkenswerte Unterschied kann mit Hilfe des Konzeptes der sogenannten Energiespeicher entsprechend des beweglichen Wassers erklärt werden, was darauf hinweist, daß das aktive bewegliche Wasser in diesem Einzugsgebiet nur gering vertreten ist. Diese Tatsache ist vor allem eine Folge der Bodenzerstörung durch Baumaschinen.

Es zeigt sich, daß graphische und isotopische Methoden unterschiedlichen Rahmenbedinungen unterliegen. Während die erste erklärt, wie schnell Wasser fließt (Antwortzeit), erklärt die zweite, wie lange Wasser im Einzugsgebiet verweilt (Verweilzeit). Die Kombination beider methodischen Ansätze ergibt folgendes Bild: Wasser wird für eine lange Zeit in den Einzugsgebieten zurückgehalten, bevor es schnell durch den Piston-Flow-Effekt verdrängt wird. In diesem Zusammenhang spielt auch der kapillare Aufstieg im Löß eine große Rolle.

Die Sensitivitätsanalyse des Zwei-Komponenten-Modells weist darauf hin, daß die korrekte Berechnung der Input-Daten für langfristige Untersuchungen von großer Bedeutung ist. Die Anwendung des Piecewise Linear Baseflow für älteres Wasser ist geeignet für die langfristige Separation. Die Sensitivitätsanalyse der kurzfristigen Untersuchung zeigte, daß die Ermittlung der elektrischen Leitfähigkeitwert für Ereigniswasser von grosser Wichtigkeit ist. Die direkte Anwendung der elektrischen Leitfähigkeitwerte des Niederschlags für Ereigniswasser kann den direkte Abflußanteil am Gesamtabfluß bis zu 13 % an den hier untersuchten Einzugsgebieten unterschätzen. Diese Unterschätzung in anderen Einzugsgebieten mag noch höher liegen.

1 Introduction

The process by which stream flow is generated from precipitation is not yet completely understood. The basic questions concerning this process are: of which components does the stream flow consist; by which path does the infiltrated water enter the stream; and what spatial and temporal distributions of turnover exist within the basin.

To contribute to the understanding of runoff generation, two types of runoff separation methods have been used simultaneously in this study: the application of a continuous separation model based on the traditional graphical principle, and experimental investigations with the help of isotope and hydrochemical tracers. These results are compared and combined in order to establish complete conceptual model of the study site of the Kaiserstuhl in southwestern Germany.

1.1 Study of runoff generation

An important field in the study of runoff generation focuses on the hydrograph separation, which not only provides a detailed picture of the contributions of different components to runoff, but also explains the physical mechanism of the hydrological process. Therefore, despite decades of increasingly intensive study in various fields, runoff separation remains an important topic in hydrology.

The major difficulty in the separation of total streamflow into surface water and groundwater flows is the inability to directly measure groundwater discharge. To estimate groundwater component, numerous empirical techniques have been developed.

The usual approach has been a graphical separation by extrapolation of the groundwater recession curves beneath the flood peak. Graphical methods of analysis vary from the very simple to the more complex (PINDER & JONES 1969). A quantitative evaluation of the above techniques is difficult because these methods of separation have been considered to be somewhat arbitrary, unless the exact amount of the base flow can be determined (PINDER & JONES 1969, SKLASH et al. 1979, EDEN 1982, HERRMANN & STICHLER 1980).

For this purpose a number of deterministic models have been developed for the runoff separation. They include physically-based models, mathematical models and black-box models. In particular, physically-based models have been largely developed in recent decades. These models use physical parameters concerning spatial variability to interpret the reality (ABBOTT et al. 1986).

However, many models are overparameterized, so the models may not function well outside the range of data used for their calibration (HOOPER et al. 1988). Also, the comparison of predicted to observed hydrographs cannot be considered a sufficient test of models that aim to simulate the internal responses of a catchment (BEVEN 1989). For that tasks, some simple black-box models are still appropriated because they use minimal amounts of data, and quickly and easily provide the order of magnitude of runoff components.

The residence times of storage volumes determined by these models are usually smaller than those obtained by field experiments. SKLASH (1990) wrote:

> *One of the major reasons of this failure is that the models often incorrectly estimate the residence time of the water in the catchment, that is, the flow paths assumed in the models are often wrong. After all, even though a hydrologist can physically measure the amount of overland flow and subsurface flow on part of a catchment, it is not always obvious where this water came from and how long it was in the watershed.*

Indeed, the models and the traditional graphical methods have some fundamental weakness. Efforts to acquire accurate information regarding runoff separation have been assisted by the use of tracer methods. Tracer techniques provide direct insight into the dynamics of surface and subsurface water bodies (LEIBUNDGUT 1982).

Tracer hydrology is understood as a generic term for certain fields within isotope hydrology and the tracer techniques (LEIBUNDGUT 1982). In the past two decades, stable isotopes have been used to obtain better insight into the mechanism of runoff generation. About 0.2 % of the water molecules in natural water contain the stable oxygen isotope. Because of their conservative characteristics they seemed to be ideal tracers for the purpose of separating hydrograph into the component displaced from the catchment (old water) and the one supplied directly from the rainfall generating the storm flow response (new water). The results of many studies using isotopes have shown that groundwater is the dominant contributor to storm and snowmelt runoff in stream in their particular study areas (FRITZ et al. 1976, SKLASH et al. 1979, RODHE 1981, HOOPER & SHOEMAKER 1986, HERRMANN et al. 1986, MCDONNELL et al. 1990, OGUNKOYA & JENKINMS 1993). Of course, in order to use stable isotopes to separate runoff, a number of assumptions or conditions have to be fulfilled. For example, isotopic method cannot be used if a difference between isotopic concentration of precipitation and that of groundwater is not obvious.

Alternative tracers are often needed to compare results and allow for the separation of three or more components. Some chemical parameters, such as electrical conductivity, chloride, sulfate, and silica have been used for this purpose (SKLASH & FARVOLDEN 1979, HOOPER & SHOEMAKER 1986, OGUNKOYA & JENKINMS 1993). However, chemical parameters are often influenced by chemical reactions and physical processes. Therefore, before a parameter is chosen for hydrograph separation it should be verified as a conservative tracer that is independent from residence time, which in turn is dependent upon such characteristics of the catchment as topography, soil type, plant cover, climate and the hydrological conditions (KENNEDY et al. 1986). On the other hand, if the tracer reacts with surrounding materials, changes of tracer concentration in water may yield information about the flow paths (WELS et al. 1991).

In addition, the use of artificial tracers is very effective for site-specific local studies. They can often provide vital additional information. For example, artificial tracers were used successfully in recent study of preferential flows (LEIBUNDGUT 1996).

Clearly, interpretations of isotope, hydrochemical and artificial tracers are incomplete in isolation. Using all three together, it is often possible not only to differentiate three or more components, but also to determine the path of infiltrated water flowing into a stream. On the other hand, the value of theoretical models can be greatly enhanced if they are developed in close cooperation with field studies. Cooperation would ensure that the physics of the problem are well understood.

1.2 Objectives of the present study

The majority of studies cited in the previous section do not report on experiments, in which the two types of investigations dealt with above are carried out in the same basin, in particular for seasonal time scales. For this reason, this study uses various methods simultaneously to examine runoff generation for long-term and short-term time scales within a basin, in order to gain some understanding of the mechanism of runoff generation for both water balance and storm events.

The DIFGA model (SCHWARZE et al. 1991), a continuous component separation model using daily hydrometeorological data, is chosen for this study. This component separation model is linked to the traditional graphical method, making it as the traditional principle.

Experimental investigations were conducted with the help of natural tracers (including isotope and chemical tracers) and artificial one. To compare the results with the DIFGA model, the long-term investigations were performed, including sampling at each four hour intervals throughout the hydrologic year 1995. These data permit comparison and combination of different methods within the basin. Short-term investigations were also performed in order to better understand runoff mechanisms.

The small-terraced Rippach catchment and the neighboring large-terraced and re-allocated Loechernbach catchment, located in the eastern part of the Kaiserstuhl in southwest Germany, seem to be highly suitable for this study. This is because the small-terraced Rippach catchment is rarely influenced by anthropogenic changes. In contrast, the large-terraced Loechernbach catchment had a large change in runoff behavior after being terraced. Particularly, a number of installed drainage subpipes facilitate the investigation of rapid interflow (throughflow) in this basin. Moreover, the extraordinary homogeneity loess sediment and constant land utilization make these areas suitable for water balance research. In addition, the hydrometric instrumentation and monitoring of data have existed for more than thirty years in these catchments.

The aims of the present study are to investigate runoff generation within two basins defined as research sites. The specific topics are:

(1) to check the feasibility of the application of the recession analysis model DIFGA in the study sites by analyzing the runoff components and storage behaviour,

(2) to estimate the long-term average amounts of water being contributed from various sources with the help of a stable isotope (oxygen-18),

(3) to choose suitable hydrochemical tracers for runoff separation, and to differentiate runoff components for the observation year by using a two-component model,

(4) to compare the long-term results obtained by stable isotope and hydrochemical tracers, and to combine them to estimate the order of third runoff component,

(5) to determine the average residence times and storage volumes of water at the study sites, with the help of mathematical flow models, and to determine the summer and winter infiltration rates using isotope data,

(6) to investigate the contribution of event water for individual storm events, and to obtain new indicators of runoff mechanisms, in particular of the influence of subsurface flow contributing to storm runoff,

(7) to examine the differences between model and tracer methods, in particular the differences of estimated residence times and storage volumes, and to combine the results of the model and tracer methods in order to construct a conceptual model of the hydrological system in the study sites,

(8) to quantitatively estimate the uncertainty of the two-component model and to analyze the difference between rain water and event water and between baseflow and pre-event water, particularly by using hydrochemical tracers.

2 Theoretical background

The theoretical background of different methods used in this study for the analysis of runoff components will be given in this chapter; they include the traditional hydrograph concept, half-graphical component model, tracer methods, and corresponding two-component model for long-term and short-term separation. A overview of the various definitions of flow components is discussed, including a summary of the old water concept. Finally, the backgrounds of various mathematical flow models determining the residence time are given.

2.1 Conceptual model for runoff separation

2.1.1 Concept of runoff generation

The term "Hydrograph" refers to the graphical representation of the instantaneous rate of discharge of a stream plotted with respect to time. The shape of a hydrograph is influenced mainly by climatic, topographic, pedologic and geologic factors. The shape of a recession, which is a useful tool in hydrology, is largely independent of both the characteristics of the storm causing the rise and the previous soil moisture of the basin.

SHERMANN (1932) first proposed the theory of the unit hydrograph, and the concept of dividing flood water into overland flow and baseflow. HORTON (1933) then introduced the concept of overland flow, in which he presented the famous infiltration theory, i.e., infiltration capacity determines whether overland flow is formed. HERTZLER (1939) then introduced the concept of interflow. This division of runoff into three components (direct surface runoff, interflow and groundwater or base flow) is still accepted today.

Since interflow is difficult to determine, a hydrograph can be simply divided into surface runoff produced by a volume of water derived from a storm event, and base flow contributed from groundwater. The boundary between surface runoff and base flow is difficult to define and largely depends on the geological structure and composition of the catchment. The baseflow levels are also affected by the general climatic and pedological state of the area; they tend to be high after periods of wet weather and can be very low after prolonged drought (SHAW 1988).

2.1.2 Hydrograph separation by traditional graphical method

A common method of separating the runoff into surface runoff and groundwater flow is to analyze the groundwater recession or depletion curves. This analysis is based on the different velocities of direct runoff and baseflow. If all the groundwater discharge from the upstream area is intercepted at the output point, then the equation of the recession curve is:

$$Q_t = Q_0 \, e^{-t/k} \tag{2.1}$$

where Q_0 is initial discharge, Q_t is discharge at time t, and k is the recession constant whose value is dependent on the units of t.

The base flow recession curve is often required in hydrologic analysis, and one method of developing this is to piece together sections of recession from various storms until a complete composite curve has been obtained.

There are essentially three traditional graphical methods of base flow separation. The first consists of extending the base flow recession curve back under the peak of the hydrograph. The rising limb of the groundwater curve is constructed by a separation of a simple line (Fig. 2.1, curve 1). This method may have some advantages where groundwater contributions are relatively large and reach the stream quickly (CHOW 1964). The DIFGA method (SCHWARZE 1991) used in the present study employs this method. According to the second method (Fig. 2.1, curve 2), the separation of base flow is accomplished by constructing a straight line from the beginning of the surface runoff (point A in Fig. 2.1) to a point on the recession curve representing the end of the direct runoff (point B in Fig. 2.1). The third method (Fig. 2.1, curve 3) consists of extending the base flow recession

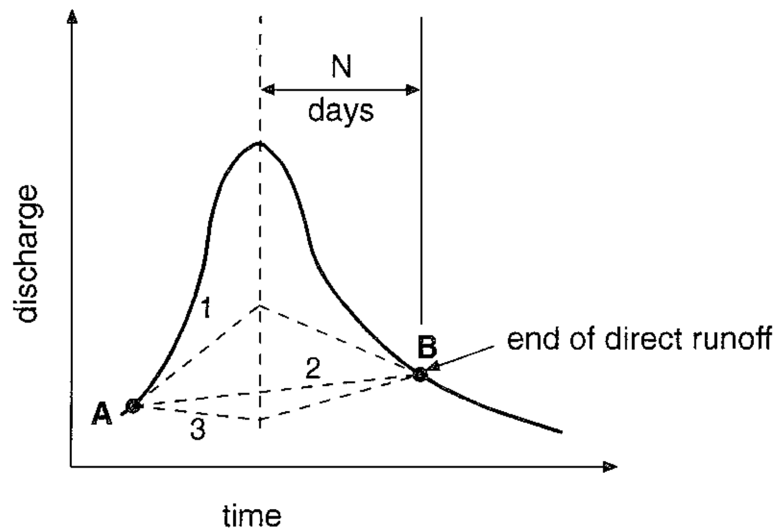


Fig. 2.1 The traditional graphical methods of base flow separation (MUTREJA 1990)

curve, which occurs before the surface runoff, past point A to a point beneath the peak and then extending a straight line to point B, representing the end of the direct runoff. This method was used by LUFT (1980) in the Rippach basin.

Clearly, the traditional graphical methods cannot consider the hydrological processes in a quantitative manner; the determination of the reaction time between the beginning of rainfall and the end of direct runoff is somewhat speculative. Hence, any separation according to these methods is hampered by a lack of physical explanation. The later tracer methods thereby introduce the concept of "age" of water (SKLASH & FARVOLDEN 1979).

The hydrograph separation discussed in the present study covers a period of one year. The traditional graphical method, applied in a component model, and experimental method have been used simultaneously. Instead of the simple hydrograph in Fig. 2.1, a more complex hydrograph resulting from more closely spaced rainfall events are separated, based on data of the hydrologic year of 1995. The corresponding methods will be discussed in the latter sections of this chapter.

2.2 Terminology

In earlier studies, various terms have been used to describe different runoff paths. To minimize the conceptual confusion in the present study, an overview of these different definitions of flow components is provided. In contrast to SKLASH & FARVOLDEN (1979), who divided runoff sources into time aspects, ultimate delivery mechanism aspects, and historical aspects, a new classification will be presented which emphasizes the differences between traditional aspects and experimental aspects. These are summarized in Table 2.1.

2.2.1 Mechanism aspects

These terms describe the ultimate delivery:

Hortonian overland flow: this term was first coined by HORTON (1933). It is generated when the rainfall rate exceeds the infiltration rate, excess rainfall then travels overland in sheet to small rivulets and then to streams. This theory implies that new water is dominant in the runoff response in well-vegetated regions (PEARCE et al. 1986).

Partial area overland flow: this term is introduced by BETSON (1964) to describe water which issues from certain areas in the watershed where surface saturation is instigated from below by a rising water table.

Variable source area-subsurface flow is introduced by HEWLETT (1961). It is assumed that quasi-uniform basin-wide generation of runoff via Hortonian overland flow cannot satisfactorily explain storm runoff generation in such environments. The variable source area subsurface flow concept states that the areas contributing to storm runoff expand and contract in response to climatic factors. The transfer of water from the hillside to the stream is accomplished through subsurface routes (SKLASH & FARVOLDEN 1979).

Base flow is the flow in the stream when all of the storm water (or snowmelt) has been routed through the system. It is commonly composed of water draining slowly from both above and below the water table. In isotope hydrology this flow is assumed to be the integrated values of deep soil water and groundwater.

The mechanism aspects mainly associate the major component of stormflow hydrographs with water of precipitation events, rather than with water that was in the basin prior to the precipitation event.

Tab. 2.1 Components of runoff

Type of Source	Components
mechanism aspects	Hortonian overland flow
	partial area overland flow
	variable source area-subsurface flow
	base flow
time aspects	rapid direct runoff
(graphical aspects)	delayed direct runoff
	short-term baseflow
	long-term baseflow
geographical aspects	surface water
	subsurface water (soil- and groundwater)
morphological aspects	overland flow
	return flow
	macropore flow
	matrix flow
	translatory flow
	groundwater ridging
mobility aspects	mobile water
	immobile water (stagnant water)
transit time aspects	event water (short-term)
	pre-event water (short-term)
	new water (long-term)
	old water (long-term)

2.2.2 Time aspects (DIFGA components)

The time components (graphical aspects) only consider the chronological order in which runoff enters the stream, neglecting the origin of the water and the physical generation process.

Direct runoff is defined as the rapid component, in contrast to indirect runoff. Direct runoff is the sum of rapid runoff and delayed direct runoff.

Rapid direct runoff is the fastest component entering the stream. It is usually a part of the flow which moves across the surface of the ground, facilitated by asphalt roads, saturated area and channel precipitation. The water stored at the surface can also contribute to it, but this part is very small in most cases. In some cases, even pipe flow can also contribute to this component.

Delayed direct runoff is defined as the second fast component. It is usually considered as soil-internal lateral runoff in near surface soil layers. This component can include pre-event water which is pushed out through macropore flow or subpipes.

Indirect runoff is defined as the slow component, in contrast to direct runoff. It is the sum of short-term and long-term baseflow.

Short-term baseflow represents the fast indirect runoff, and is mainly concerned with delayed hypodermic baseflow (base runoff), and enters a stream-channel from ground-water or other delayed sources.

Long-term baseflow is the slowest component contributing to runoff. It is understood as the baseflow component from groundwater, and its storage is a key component in cases of rock types with significant drained pore space. In lowland, the loose rock aquifer forms the space of origin. After a long dry period the long-term base runoff is the only continuous component and represents minimal groundwater discharge (DYCK & PESCHKE 1995).

2.2.3 Geographical aspects

Surface water is the portion of total water volumes considered which moves across the surface of the ground. Surface water includes channel precipitation.

Subsurface water is defined as the underground flow in this study. It is the complement of the surface water. Using chemical tracers such as silica and chloride, runoff can only be divided into surface and subsurface water, because the water undergoes a chemical process after infiltrating into the soil, even after a short time. Such concentration changes do not occur with isotopes. Subsurface waters can be grouped into two major categories, based on their origin and the order of contribution to stream runoff:

- *Soil water* is the subsurface water occupying the unsaturated zone. In this study, soil water is assumed for the convenience to be the same as interflow.

- *Rapid interflow* is the water from the near surface soil layer (hypodermical flow), or infiltrated water which flows rapidly through the soil mantle (throughflow), more or less directly to a stream via interconnected large pores or porous structural soil features (PEARCE et al. 1986). This flow rapidly contributes to a stream. Hence, this portion is assumed to originate from event water or new water.

- *Slow interflow* is considered to be the delayed water from the unsaturated zone (translatory flow), and from near-stream groundwater. In this study this is assumed to be old water. This can be understood as runoff from a zone of solid rock loosening in secondary mountain regions, including local valley aquifers (water-bearing bed) caused by damming wetness (DYCK et al. 1995).

Groundwater is the subsurface water occupying the saturation zone.

2.2.4 Morphological aspects

From a morphological point of view, the flows can be divided into the following components:

Overland flow is the same as that explained by the mechanism aspect (Chapter 2.2.1).

Return flow is the infiltrated water which returns to the surface after travelling a short distance beneath the surface (DUNNE 1970). Some old water may appear through return flow, so return flow is not always new water.

Macropore flow concerns the flow in readily visible macropores which are a well-established feature of soils. These can be grouped as below (BEVEN & GERMANN 1982):

(1) Cracks or planar voids in fine textured soils;

(2) Cylindrical holes or channels formed mainly by plants or animals;

(3) Voids called 'vughs' that are intermediate in shape between cracks and channels;

(4) Soil pipes which are large macropores. BEVEN & GERMANN (1982) defined soil pipes as having diameters greater than 4 cm. Some authors use the term pipe flow (e.g. SKLASH et al. 1996) because pipe flow can contribute rapidly and significantly to storm runoff in streams (NEWSON & HARRISON 1978).

Matrix flow is the water which moves through micro- and mesopores of the soil matrix. Matrix flow can appear in both unsaturated and saturated zones (SCHERRER 1996). The velocity of the water movement is dependent upon the permeability, sediment type and porosity. The permeability in a humid region is larger than that in a dry region. In this study matrix flow is considered as simply old water.

Translatory flow is described as a mechanism by which new water entering the saturated zone on a hill slope causes a displacement of old water at the base of the slope (HEWLETT & HIBBERT 1961). This is due to a rapid wave-like transmission of the pressure changes at the boundary of the saturated zone.

Groundwater ridging is a term favored by SKLASH & FARVOLDEN (1979). This mechanism is associated with areas adjacent to streams, where the capillary fringe or tension-saturated zone above the water table is close to the surface. Capillary fringe is the zone above the water table that remains saturated under negative pressure. Because this zone has little or no storage capacity, in locations where it extends up to ground surface the addition of a relatively small amount of water can cause a large and rapid rise in the water table. A water table rise could in turn result in a rapid increase in hydraulic gradient toward the stream (ABDUL & GILLHAM 1984). This capillary fringe is very obvious in the valley area of the present study sites.

Among the flows dealt with above, the macropore flow (including pipeflow) is the most important one to deliver subsurface stormflow to streams. Recently, hydrologists have recognized the importance of macropores as an issue of contaminant hydrology (MOSLEY 1979, BEVEN & GERMANN 1982, NEAL et al. 1988, MCDONNELL 1990, GERMANN 1990, SKLASH et al. 1996). Flow in the matrix can be modeled to good approximation by Richards' Equation on Darcy's Law. However, in soils with macropores, continuity of the matrix space cannot necessarily be assumed. Thus the presence of macropores may give rise to responses that differ from predictions based on Darcian principles.

2.2.5 Mobility aspects

These aspects are used in isotope hydrology to describe whether or not the water participates in flow during infiltration. There is as yet no clear definition of them.

Mobile water is usually understood as water that participates in flow during infiltration. Mobile water can flow freely without hydraulic pressure (ZOJER et al. 1996).

Immobile water (stagnant water) is water that does not participate in flow in usual terms. Immobile water is usually understood to exist in double porosity media, such as joint aquifers (HERRMAN et al. 1982), whereas in the microporous matrix the water is stagnant or quasi stagnant (MALOSZEWSKI 1991). The most reliable information about transit time of immobile water relies on tritium, since tritium can determine transit times of up to 150 years, in comparison with stable isotopes whose signals cannot determine transit times longer than 4 years (MALOSZEWSKI 1996).

2.2.6 Transit time aspects (tracer components)

Transit time components have characteristic and measurable isotopic values, which are different from the time components dealt with above from graphical aspects. Transit time components concern how long water is stored within a basin, while time components concern how fast water enters a stream. The latter chapters will discuss this difference.

Event water is the water added to the basin by a specific runoff-inducing event, either rain or snow-melt (SKLASH & FARVOLDEN 1979). It has the same content as the subsequent new water, but is limited to short-term rainfall events in this study. If, in a rainfall event, the isotopic content of the event water varies significantly over time, a traveling time for the new water to enter the stream must be assumed (HOOPER & SHOEMAKER 1986).

Pre-event water is considered to be the volume of water which exists in the basin prior to a specific event and consists of groundwater, vadose water and surface storage (SKLASH & FARVOLDEN 1979). It has the same content as old water, and is limited to the short-term rainfall events in this investigation.

New water includes at least three subcomponents; channel precipitation, overland flow and the rapid subsurface flow from frequent drainage. All three subcomponents are assumed to have a chemical and isotopic identity similar to precipitation. Here the rapid subsurface flow moves through macropores, and would therefore reflect incident precipitation chemistry contributing to runoff (PILGRIM et al. 1979). In this study, new water is limited to long-term research period. For the present seasonal and annual hydrograph separation, the isotopic concentration of new water is calculated from a weighted average of the isotopic concentrations of the rain volume of a whole week. Instantaneous introduction of new water to the stream can be assumed. The monthly isotopic new water is the weighted weekly isotopic new water.

Old water is considered to be the water which exists in the basin prior to a particular week or month and consists of groundwater, vadose water and surface storage. In this study, old water is limited to long-term period.

In previous studies there are different definitions of old water, which are summarized below:

(1) Streamwater isotopic content taken during low-flow periods (HOOPER & SHOEMAKER 1986). This is the easiest method because the streamwater samples are easy to get.

(2) Piezometer samples. Under low-flow conditions the $\delta^{18}O$ values for the piezometer samples are indistinguishable from the streamwater values (HOOPER & SHOEMAKER 1986).

(3) Direct use of groundwater sampling data. The direct use of groundwater sampling data is dependent on the catchment characteristics, the available conditions and on how one defines old water in the catchment.

Generally speaking, groundwater and old water are two different entities. It is questionable to use one groundwater sample to represent the old water of the whole basin, because a watershed is a complex system. There is a spatial variation of the isotopic content of $\delta^{18}O$, depending upon the altitude and the depth of the groundwater. It is considered impossible to calculate accurate δ-values of the groundwater contributing to streamflow on the basis of groundwater sampling, even with a very dense network of groundwater sampling points (RODHE 1987). After all, baseflow $\delta^{18}O$ as the isotopic signature of old water should be more suitable for the hydrograph separation. In the present study the first of the three definitions is used. Baseflow $\delta^{18}O$ is assumed to be the isotopic signature of old water.

Three options for calculation arise for the old water defined above:

(1) *Constant.* Old water is considered to be only the water present in the basin prior to any rainfall events (RODHE 1981, CHRISTOPHERSEN et al. 1984).

(2) *Stepwise.* Throughout the event, old water is defined as a constant value equal to the streamwater value before the start of the event and maintaining the same isotopic content throughout the event (HOOPER & SHOEMAKER 1986).

(3) *Piecewise linear.* Old water isotopic content is defined by connecting the data points of isotopic content of low-flow samples and linearly interpolating between these points (HOOPER & SHOEMAKER 1986).

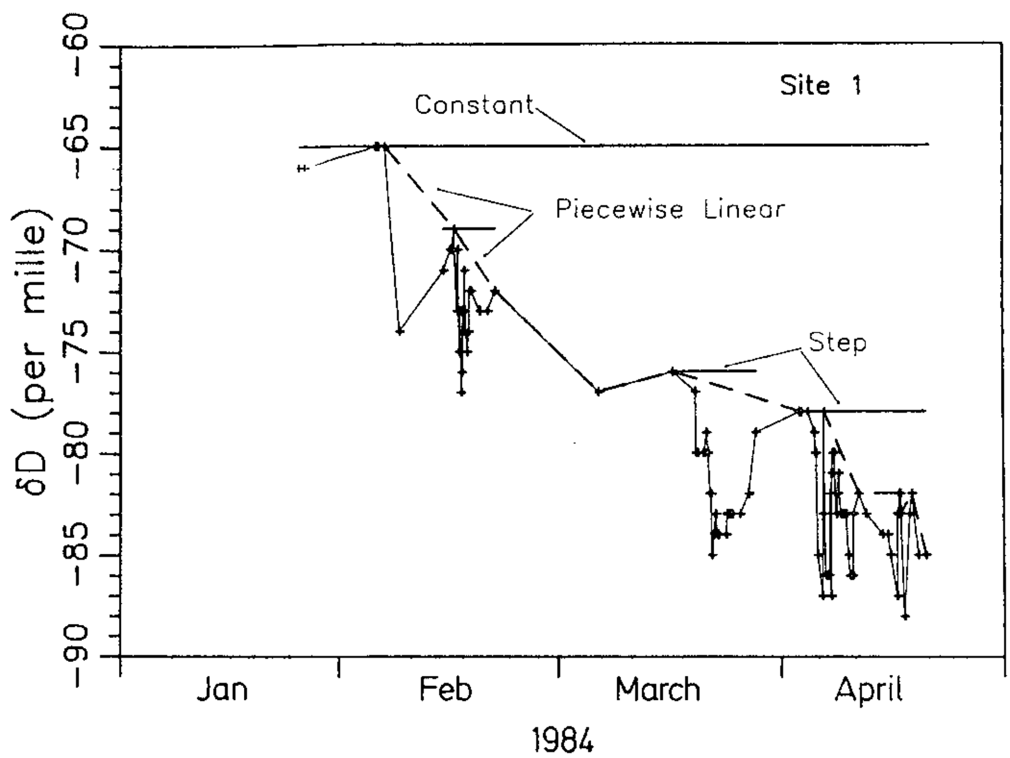


Fig. 2.2 **Different isotopic definitions of old water** (HOOPER et al. 1986)
Solid line: a constant value throughout the study
Step solid line: a step definition with a constant value throughout the event
Dashed line: a piecewise linear definition connecting low-flow values
Solid line with crosses: streamwater isotopic content

In the literature, the $\delta^{18}O$ values of old water have often been considered to be comparatively constant during the year. In the present study, the third method, piecewise linear, is used (Chapter 5). Generally speaking, this approach is less dependent on one value being designated as "old" and can incorporate seasonal trends more reliably.

2.3 Conceptual models and the DIFGA model

2.3.1 A brief overview of conceptual models

Hydrological models are divided into two groups: deterministic models and stochastic models. The present study will be concerned with a conceptual model, one type of deterministic model.

The conceptual model approach to rainfall-runoff modelling lies between physically based models and black box models. These models usually rely on a simple arrangement of interlinked conceptual elements, such as the common storage component, each representing a segment of the land phase of the hydrological cycle (JAIN 1993). The conceptual model originated from the theory of the unit hydrograph (SHERMANN 1932). The disadvantages of the unit hydrograph were soon found. These include the difficulties of separating surface and baseflow and deriving of the unit hydrograph, as well as the problem of determining effective rainfall (TODINI 1988). To overcome the non-particularly linear behavior of the real hydrological system, a theoretical cascade of reservoirs was developed (NASH 1958).

After the sixties, a series of other approaches toward rainfall-runoff modelling, such as physically-based models, were developed. These models use a number of interconnected conceptual elements, each of which represents the response of a particular subsystem (TODINI 1988). In particular, physically-based models use parameters which have physical interpretation to represent spatial variability (ABBOTT et al. 1986).

However, there is a great danger of overparameterization. Also, a single parameter value cannot reproduce the heterogeneity of responses engendered by the variable catchment characteristics. Most models are not suitable for application to spatial complex basins. On the other hand, comparison of predicted and observed hydrographs cannot be considered a sufficient test of models that support to simulate the internal responses of a catchment. Thus BEVEN (1989) indicated that the physically-based models are themselves conceptual models.

Continuous component models have been developed during the last two decades. They seek to describe realistically the multi-component processes of the hydrological cycle such as overland flow and baseflow. This scientific approach is a laudable ideal (SHAW 1988), but there exist so many complexities in the hydrological cycle that the calibration of these models becomes necessary.

Clearly, the development of models in the future must be carried out in combination with the field experiments or tracer methods, to ensure consistency between model predictions and real world processes.

2.3.2 The DIFGA model

The DIFGA model, developed in Dresden, Germany (SCHWARZE 1986), is a computer-aided component model which separates hydrographs continuously using daily routine data, and which is a valuable tool for assessing runoff component and transit time with complete areal coverage on a regional scale. The model structure is shown in Fig. 2.3 (SCHWARZE et al. 1991).

DIFGA is based on the assumption that the rainfall-runoff process within a small catchment (< 300 km^2) can be represented by a small parallel circuit of single-linear-storages. With the analysis of a long-term time series (e.g. 10 years) of precipitation and runoff the number of storages, which are defined as runoff components, and their storage constants can be evaluated by an inverse procedure (KOENIG et al. 1993). DIFGA determines up to four components of different mean transit times with the associated storage constants and reservoirs. The separation result is verified by the actual monthly water balance (SCHWARZE et al. 1993). In this section a brief overview of DIFGA will be given. A complete description of the method is given in SCHWARZE (1986, et al. 1990).

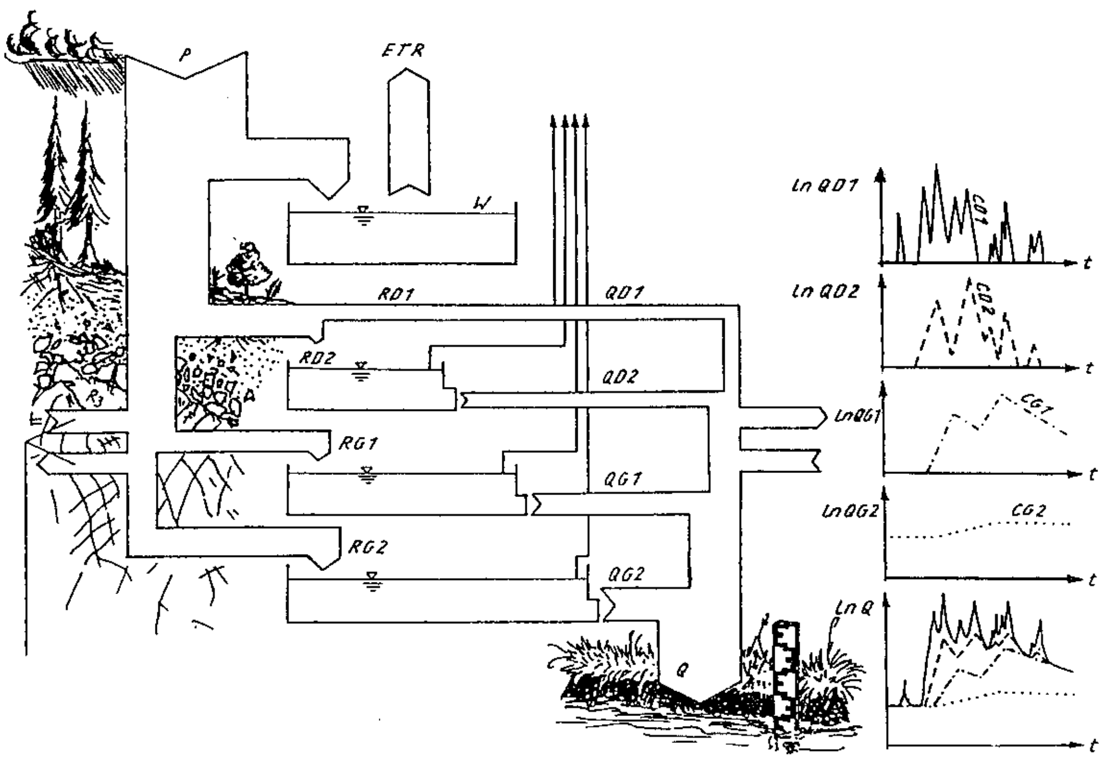


Fig. 2.3 The model structure of the DIFGA (SCHWARZE 1991)

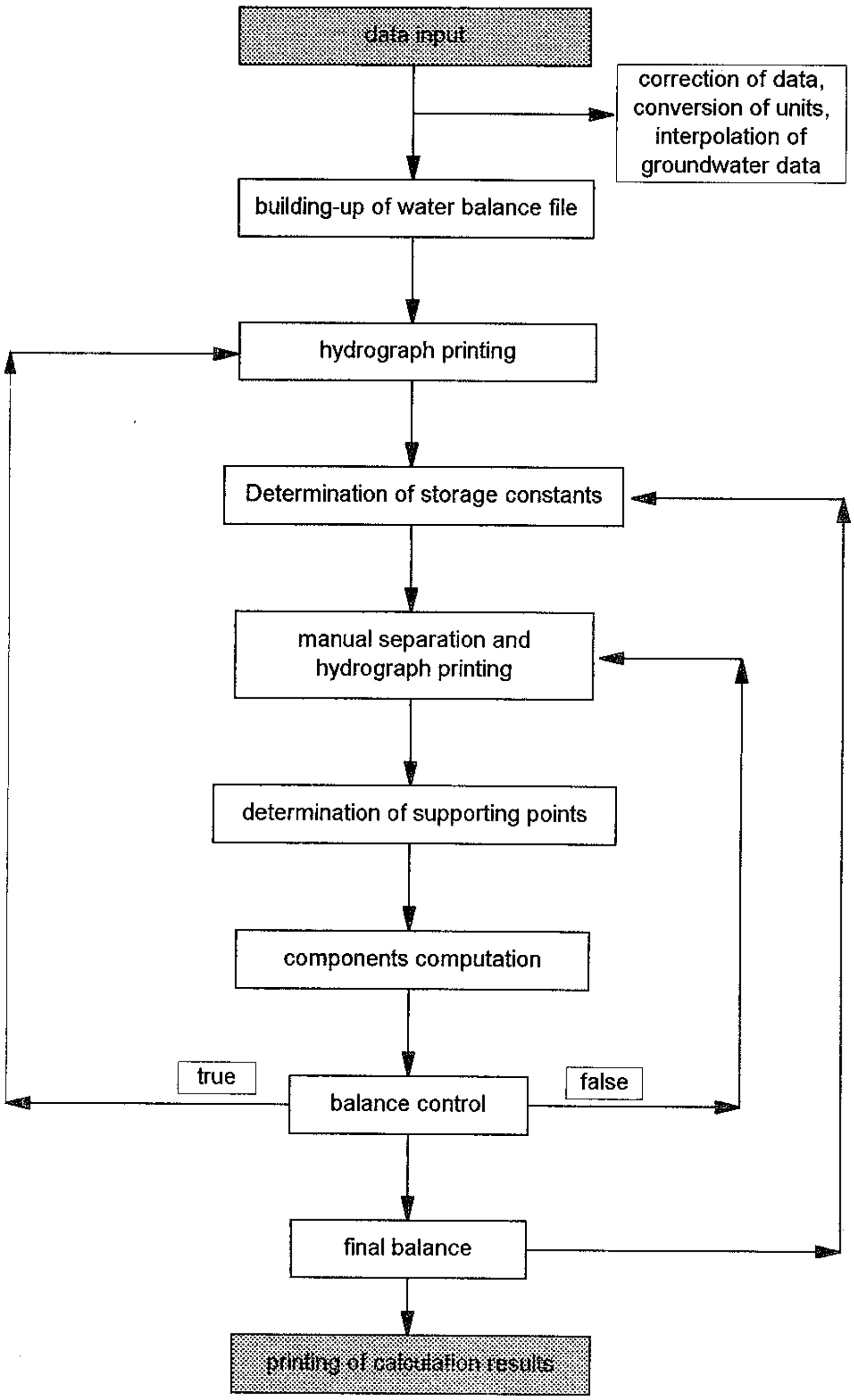


Fig. 2.4 Structure diagram of the DIFGA model (SCHWARZE 1986, modified)

In the DIFGA program, the storage-discharge relation is described by the simplest linear reservoir with the storage coefficient C:

$$V_g = C \cdot Q \tag{2.2}$$

Where V_g is the storage volume and Q is the mean discharge. The discharge of a single reservoir is represented by the equation of the exponential flow recession:

$$Q(t) = Q(t_0)\, exp(-(t-t_0)/C) \tag{2.3}$$

or $$lnQ(t) = lnQ(t_0) - (t-t_0)/C \tag{2.4}$$

Four flow components can be distinguished: fast direct runoff, delayed direct runoff, short-term baseflow and long-term baseflow. Correspondingly, *RD1, RD2, RG1* and *RG2* deal with runoff formation, and *QD1, QD2, QG1* and *QG2* deal with runoff concentration. For each component a specific storage coefficient must be estimated.

The total discharge is the sum of the discharge of *i* parallel reservoirs:

$$Q(t) = \sum_{i=1}^{N} Q_i(t_0) e^{-\frac{t}{C_i}} \tag{2.5}$$

or $$\ln q(t) = \ln \sum_{i=1}^{N} Q_i(t) = \ln \sum_{i=1}^{N} Q_i(t_0) e^{-\frac{t}{C_i}} \tag{2.6}$$

where $Q(t)$ is the total flow at time t, $Q_i(t)$ is the flow of the component i at time t; $Q_i(t_0)$ is the starting value for Q_i.

The first and the most important step is the separation of the runoff concentration of long-term base flow (*QG2*) by determining the corresponding storage constant *CG2*. Its accuracy directly affects the accuracy of the other components. To determine this constant manually, it is necessary to find long dry recession periods with durations of several months. Then the discharge recession is approximated by a straight line in a semilogarithmic scale as the lower envelope of the hydrograph, with the number of intersections, kept as low as possible. The slope of the line determines *CG2*. With a stable value of *CG2* it is possible to separate the recession sections of *QG2* incrementally. Such a procedure can be carried out automatically by computer. After the separation of this component, a record of the differences between total discharge and long-term baseflow discharge (*Q-QG2*) is given.

Similarly, the storage constants of short-term baseflow (*CG1*) and delayed direct runoff (*CD2*) can be determined. Their time scales are on the order of several days. The corresponding separations yield the differences *Q-QG1-QG2* and *Q-QG1-QG2-QD2*. The fast direct runoff (*QD1*) is the rest of the differences, and all other parts can only estimated roughly.

The separation of *QG2* (long-term baseflow) from the other components during the strong recession phase and the following rebound phase is often impossible. In order to solve this problem, a vegetation theory was developed in DIFGA. The hydrograph during the vegetation period should be considered separately.

During long, dry periods the separation is critical due to the long distances between the graphical supporting points used in the construction of *QG2* (long-term baseflow). Therefore, it is favorable to determine *QG2* by using measurements of the groundwater level *L*. *QG2* is expected to be as below:

$$QG2 = a \cdot L^b \tag{2.7}$$

where *a* and *b* are parameters (SCHWARZE et al. 1993).

After then, a continuous water balance evaluation on a monthly basis is used for each flow component, to avoid errors in the separation process. The water balance function is presented in the following form:

$$P\text{-}RG1\text{-}RG2\text{-}RD\text{-}(W+ETR) = 0 \tag{2.8}$$

where *P* = sum of precipitation

RG1 = sum of fast base flow formation

RG2 = sum of delayed base flow formation

RD = sum of flow formation of the direct components (*RD1* + *RD2*)

W+ETR = sum of residuals, containing real evapotranspiration and storage, characterized by a depletion of *ETR*

By means of the following relation, the component runoff formations, e.g. $RG2_i$ (long-term baseflow) can be determined as:

$$RG2_i = CG2 \cdot \Delta QG2_i \cdot B/A \tag{2.9}$$

where $\Delta QG2_i$ = increase of *QG2* during step *i* in m^3/s

CG2 = storage coefficient of the slow baseflow in days

A = surface area of the river basin in km^2

B = time conversion constant: 86.4

Similarly $RG1_i$ (runoff formation of short-term baseflow) can be determined from the difference between total discharge and long-term baseflow discharge (Q_i-$QG2_i$). RD_i (runoff formation of delayed direct runoff) is obtained as residual from the total observed discharge, and from the *QG1* that has been separated and the *QG2* that has been integrated over the appropriate number of months. If (2.8) is negative, then corrections of the separation are necessary until positive quantities are calculated. Changes in water storage due to evapotranspiration, or subsurface inflows and outflows across watershed must be taken into account in the application of the equation.

2.4 Isotopic tracers

2.4.1 Isotopes in hydrology

The isotopes of tritium (^3H), deuterium (^2H) and oxygen-18 ($\delta^{18}O$) are suitable for tracing water in the hydrological cycle because they are part of the water molecule. Both stable isotopes $\delta^{18}O$ and ^2H are regarded as conservative in water traveling through a basin. Any change in the isotopic ratio is assumed to be the result of mixing water with different isotopic concentrations (MOLDEN & CERNY 1994). Therefore it is possible to use stable isotopes to determine the mixing, and the movement of different water types, or to determine the residence times of the water in a basin.

The present study is only concerned with the stable isotope ^{18}O. The $\delta^{18}O$ input is given by the δ-values of precipitation, and the $\delta^{18}O$ output is given by the δ-values of streamwater. In order to interpret the δ-values in terms of hydrological processes, any changes in the δ-values other than by the mixing of different waters, i.e. any changes by fractionation, must be taken into consideration and either be estimated or treated as errors.

Oxygen-18 in precipitation

The following effects may result in changes of ^{18}O composition of precipitation samples:

- The amount effect: a negative tendency between the amount of rainfall and the isotope content (-1.0 to -1.5‰ per 100 mm, DANSGAARD 1964)

- The altitude effect: existence of a negative linear correlation of heavy isotope content of the rain with altitude (0.10-0.50‰ per 100 m, MOSER & RAUERT 1980)

- Distance to the source of vapor: continental precipitation is depleted in ^{18}O as compared to marine and coastal rains (0.30‰ per 100 km from coast, SONNTAG et al. 1978)

- Seasonal and short-term variations: winter precipitation is depleted in ^{18}O with respect to summer rains. The temperature effect can be put into the following equation (FRITZ et al. 1980):

$$\delta^{18}O = (0.521 \pm 0.014)\ T_a - (14.96 \pm 0.21) \tag{2.10}$$

where T_a is air temperature in °C.

Besides the above variations, precipitation, in contact with the atmosphere, tends to become enriched in heavy isotopes due to fractionation occurring during evaporation and molecular exchange with atmospheric vapor. GAT et al. (1967) concluded that the mean $\delta^{18}O$ enrichment of the throughfall never exceeded 0.5‰. SAXENA (1987) performed a sampling experiment in a dense pine forest in Uppsala and concluded that the weighted average $\delta^{18}O$ of throughfall was enriched by 0.3‰ as compared to the precipitation.

Oxygen-18 in surface water

For the upper parts of the basin, surface waters are a complex mixing of groundwater, soil water, and precipitation. Evaporation from permanent stagnant surface waters often results in an enrichment of heavy isotopes as compared to normal groundwaters. Enrichments are also possible in rivers and in any kind of surface waters even with a short exposure time to the free atmosphere (FRITZ et al. 1980). These effects have to be taken into consideration when using the isotope method in analytical hydrology.

Oxygen-18 in unsaturated zone

The isotopic composition of water in an unsaturated zone may differ from groundwater, modification by evaporation, exchange with the atmosphere, or transpiration. From the investigation of water movement in unsaturated zones using isotope techniques ZIMMERMANN et al. (1967) concluded that with respect to the vegetation cover and the climate of Central Europe (Rhine valley), evaporation tends to increase the heavy isotope content of the upper layers of bare soils (or of the vegetation) down to a depth of several centimeters, depending on the relation between the diffusion coefficient and the effective upward velocity of water pumped out by evaporation. Their investigation suggests that dispersion is very active during downward movement, and homogenizes the isotope content after a few centimeters, or a few decimeters of vertical movement. The enrichment distribution function with depth is exponential.

In a wet soil evaporation occurs from the uppermost wet layer, which is supplied by water from below. The isotopic enrichment of infiltrating water is suggested to be negligible in soils with high infiltration capacities. Similarly, enrichment by transpiration would be of little significance in humid areas.

Oxygen-18 in groundwater

The long-term average $\delta^{18}O$ of groundwater should be close to that of precipitation. The groundwater close to the soil surface may be subjected to transpiration and direct evaporation, and exchanges with the atmosphere (RODHE 1987). Isotopic changes in the groundwater lying a few meters below the soil surface is the only isotopic exchange with the rock material, generally resulting in $\delta^{18}O$ enrichment of the water. On the other hand, the $\delta^{18}O$ of groundwater may change due to transformation of soil water into groundwater by percolation. If the isotope content does not change within the aquifer, it will reflect the origin of the water. If the isotope content changes along the groundwater paths, it will reflect the history of the water. Origin deals with location, period and processes. History deals with mixing, salinization and discharge processes (FRITZ et al. 1980).

Oxygen-18 in stream water

Stream water is the mixture of event water with water that was already in various reservoirs before an event. During the peak stage of a rainfall event, the new water fraction can increase largely. After a sufficiently long period of streamflow recession, it is probable that stream water originates from groundwater only. It is this difference that facilitates separation into two or more components. Streamflow enrichment is negligible ($< 0.1\,^{0}/_{00}$) at moderate or high flows and can rise to a probable 0.5-1.0‰ during low flow in summer (RHODE 1987).

2.4.2 Isotopic hydrograph separation

Isotopic hydrograph separation has been performed since the early 1970s, predominantly in mountainous areas in Central Europe and Canada. The investigative studies can be divided into two different time scales: short-term measurement of rainfall events such as the experiments by FRITZ (1976), SKLASH & FARVOLDEN (1979), RODHE (1987); and long-term measurement such as the experiments by MATINEC et al. (1974), HERRMANN & STICHLER (1980), STICHLER (1982), MALOSZEWSKI et al. (1992). They will be discussed separately below.

The separation of short-term runoff components

According to FONTES (1980) and KENNEDY (1986), research on the use of hydrogen isotopes to estimate sources of water during storm runoff began with HUBERT et al. (1969), who used a mass balance of tritium to estimate the relative proportions of rainfall and baseflow in storm runoff. Similar studies were later performed by MOOK et al. (1974) and DINCER et al. (1970). Later studies (SKLASH 1976, FRITZ 1976, SKLASH et al. 1979) made the important discovery that stream flow generated during snowmelt and rainfall is supplied largely by water stored in the catchment prior to the event.

Most hydrograph separations have been made using two sources of water, ignoring the possibility that soil (vadose) waters may be significant sources of water and that these soil waters may show major spatial and temporal variations in chemical and isotopic content (KENNDY et al. 1986, DEWALLE et al. 1987).

The two-component model (FRITZ 1976, SKLASH & FARVOLDEN 1979, RODHE 1987) used in the present study is as below:

$$c_d Q_d + c_g Q_g = c_t Q_t \qquad (2.11)$$

$$Q_d + Q_g = Q_t \qquad (2.12)$$

$$\frac{Q_d}{Q_t} = \frac{c_g - c_t}{c_g - c_d} \qquad (2.13)$$

where
Q_t = total discharge
Q_d = discharge of new water
Q_g = discharge of old water
c_t = tracer concentration of streamwater
c_d = tracer concentration of new water
c_g = tracer concentration of old water

For the two-component model the following assumptions are made:

(1) Stream runoff has two end-members: new water and old water. Groundwater and soil water, which together constitute old water, are chemically equivalent.

(2) Each source has its own identity.

(3) Rainfall is characterized by a relatively unique isotopic signal.

(4) Underground water and baseflow have a unique isotopic content.

(5) The contribution of stored water in surface retention, either natural or artificial, is negligible.

Limitations to these assumptions are recognized. For example, assumption (1) is difficult to satisfy because it is a simplification of catchment processes. Water samples from the soil show an isotopic content that is intermediate between values of those of the water table and of rainfall, even when differences between the water table and the soil were small. Although the contribution of undergroundwater may be underestimated, the errors should not be very significant and do not affect conclusions in many basins (JORDAN 1994).

The separation of long-term runoff components

In contrast to the short-term research, the long-term runoff separation focuses on the study of seasonal and annual variations of the runoff components. In this case the seasonal variations of the isotope input (precipitation or snowmelt) must be considered. The early studies of seasonal or annual runoff component separations are listed in Table 2.2. Generally, the early studies show that over 50 % of the runoff has a residence time of more than one year (STICHLER 1982).

For a long-term separation, the above equations 2.11 - 2.13 should be changed as follows:

$$\overline{c_d}\,\overline{Q_d} + \overline{c_g}\,\overline{Q_g} = \overline{c_t}\,\overline{Q_t} \qquad (2.14)$$

$$Q_d + Q_g = Q_t \qquad (2.15)$$

$$\overline{\left(\frac{Q_d}{Q_t}\right)} = \frac{\overline{c_g} - \overline{c_t}}{\overline{c_g} - \overline{c_d}} \qquad (2.16)$$

where $\overline{C_g}$ = isotopic content of the average base flow in the whole observation period

$\overline{C_d}$ = isotopic contents of the weighted average precipitation during the whole observation period

$\overline{C_t}$ = isotopic contents of the weighted average streamwater in output during the whole observation period

Tab. 2.2 Previous investigations for long-term period

Authors	Basin	Area (km^2)	Height (m)	Tracer	New water	Transit time (years)
Martinec et al. 1974	Dischma	43.3	1668-3140	^3H	38-48%	
Herrmann et al. 1980	Lainbachtal	18.7	670-1801	^2H	14-47%	0.7
Eden et al. 1982	Kreidenbach	1.80	1020-1360	^{18}O, EC*	50%	
Maloszewski et al. 1986	Danube River	0.30		^{18}O		0.1-0.3
Pearce et al. 1986	Maimai	<4	230 - 370	^{18}O		0.4
Ambach et al. 1988	Hintereisbach	20.0	2425-3739	^2H, ^3H	60%	
Herrmann et al. 1990	Lange Bramke	0.76	543-700	^{18}O,^2H ^3H	11%	1.9
Buzek et al. 1991	Jiifetin	≅10	240 - 750	^{18}O, ^3H	40%	0.4
Maloszewski et al. 1992	Wimbachtal	33.4	636-2713	^{18}O. ^3H		4.1
Matsutani et al. 1993	Kawakami	0.14	1500-1680	^3H	67%	2.5
Schwarze et al. 1996	Wernerbach	4.57	323-424	^{18}O. ^3H	13%	4.6

* EC: electrical conductivity

Their weight factors are considered as follows:

$$\overline{C_d} = \sum_i^n P_i C_{d_i} \Big/ \sum_i^n P_i \tag{2.17}$$

$$\overline{C_t} = \sum_i^n Q_{t_i} C_{t_i} \Big/ \sum_i^n Q_{t_i} \tag{2.18}$$

where P_i and C_{d_i} denote the amount of precipitation and the concentration of $\delta^{18}O$ fractionately collected for a certain time, e.g. one week or one month, respectively; Q_t and C_{t_i} fractionately denote the average discharge value and the concentration of $\delta^{18}O$ for a specific time.

Few studies discuss the differences between the separation calculations for hydrologic data of short and long time scales based on a two or three component model. They do have different characteristics in calculation:

(1) Short-term investigations focus on rainfall events. The isotopic contents of precipitation differ from those of groundwater. An evident variation of the isotopic contents in output (streamflow) during an event can be seen. For this reason, the separation is relatively accurate if the isotopic content in event water can be estimated accurately. Also, the corresponding pre-event water is often considered to be constant during the event. In contrast to this, long-term investigations, usually based on weekly data, must consider seasonal variations of isotopic content in both new and old water. The summer δ-values of groundwater are usually lighter than those of precipitation, and winter values are heavier. This means that there could exist at least a short period within a year during which separation using $\delta^{18}O$ is impossible due to similarity between δ-values of precipitation and those of groundwater.

(2) The calculations for old water fraction using $\delta^{18}O$ are more complicated for long-term data than are those for short-term data. For the isotopic content of both precipitation and discharge, the weighting factor must be considered. This point has often been ignored in earlier studies. The sensitivity analysis must be considered carefully, because any small error could result in unreal values of old/new water fraction. Chapter 5.3 and 7.2 will separately discuss the weighting factors and sensitivity.

2.5 Hydrochemical tracers

2.5.1 Hydrochemical tracers

Natural waters contain a complex mixture of cations and anions that include the primary cations calcium (Ca^{2+}), magnesium (Mg^{2+}), sodium (Na^{+}) and potassium (K^{+}); and the primary anions chloride (Cl^{-}), sulfate (SO_4^{2-}), carbonate (CO_3^{2-}), bicarbonate (HCO_3^{-}) and nitrate (NO_3^{-}). The exact mixture of the anions and cations in water is controlled by chemical processes, which dominate the hydrochemical response of small catchments because streamwater is largely made up of drainage water from soils.

Chemical processes can be categorized into three major groups (BREEMEN et al. 1983):

(1) Biochemical processes, including interactions between biota and the atmosphere, and interactions between biota and soil solution (e.g. assimilation and mineralization).

(2) Geochemical and soil chemical processes, including interactions between solution and the soil solid phase (e.g. cation exchange, adsorption, and chemical weathering).

(3) Chemical reactions in solution or between solution and the atmosphere (e.g. degassing of CO_2).

Rainfall bears only limited concentrations of impurities, and surface water or rapid subsurface water is low in dissolved solids. However, when water remains in contact with soils and rocks for an extended period of time or percolates into the ground, dissolved solids increase in concentration (MAIDMENT 1992). The chemistry of stream water generally reflects these reactions that occur between precipitation and watershed. The mixing of water from various sources with different chemical characteristics as it enters a stream facilitates the determination of not only the water's origins but also the pathway of the infiltrated water. Water can acquire more dissolved solids by the influence of mineral fertilizer. Also, evaporation of surface water is important in controlling the concentration of dissolved solids.

Chloride

Chloride (Cl^-) salts are highly soluble. Cl^- is also relatively free from effects of exchange, absorption, and biological activity (DAVIS 1966). Thus, Cl^- is believed to be the most chemically inert ion in the system (NEAL et al. 1988), and is seen as the conservative tracer in groundwater. Continental rain and snow may contain from 1.0 to 3.0 mg/l, but probably the average is less than 1.0 mg/l (DAVIS 1966). The concentration of Cl^- can be increased by the entrance of mineral fertilizers, liquid manure and dilute salts. The concentration of Cl^- could be increased in soil during the evaporation period (MATTHESS & UBELL 1983).

Sulfate

Sulfate (SO_4^{2-}) is formed by oxidation of pyrite and other sulfides widely distributed in igneous and sedimentary rocks. The most important SO_4^{2-} deposits are found in evaporate sediments (BOUWER 1978). Most sulfate compounds are readily soluble in water. One of the most effective natural processes for the removal of SO_4^{2-} from water appears to be the reduction of SO_4^{2-} by bacteria (DAVIS 1966). SO_4^{2-} takes part in the nutrient cycle; during the vegetation period SO_4^{2-} is partially bound in plants, later enters soil during autumn and winter with the decay of leaves, and is mineralized there, so that higher SO_4^{2-} content can be seen in groundwater during winter. Though SO_4^{2-} is not conservative to the same extent as Cl^- in salt transport, it is considered to be a good tracer in some studies. Atmospheric precipitation contains SO_4^{2-} in absolute concentrations of less than 2 mg/l (DAVIS 1966). SO_4^{2-} has strong regional and seasonal fluctuations, because most SO_2 originates from industrial gas emission.

Nitrate

Most nitrate (NO_3^-) in natural water comes from organic sources or from agricultural chemicals. An additional minor source is nitric oxides produced by lightning discharges. Some of the NO_3^- produced may be leached by percolating water and eventually reach the groundwater. However, most . NO_3^- is probably used by plants soon after it is released by bacterial action. NO_3^- compounds are highly soluble. NO_3^- is taken out of natural water only through the activity of organisms, and its concentration in rainwater is sufficiently small that they are ignored.

Bicarbonate

Most bicarbonate (HCO_3^-) ions are derived from carbon dioxide (CO_2) in the atmosphere, carbon dioxide in the soil, and from dissolving carbonate rocks. Groundwater generally contains more than 10 mg/l. Higher concentrations of HCO_3^- occur where carbon dioxide is produced within the aquifer.

Electrical Conductivity

The electrical conductivity (EC) of water is a function of temperature, the types of ions present, and the concentrations of various ions, so that variations in conductance are a function only of the concentration and the type of dissolved constituents. In dilute solutions, it is found that a linear relation exists between the concentration of dissolved solids and the conductance of the solutions. As the conductance increases, however, this relationship changes its slope depending on the quantity and quality of the salt in the solution (MATSUBAYASHI 1993). Pure rainwater will usually range in conductivity from about 5.0 μs to 30 μs.

Many studies have shown that the EC increases downstream. This has been interpreted as either the result of an increasing fraction of groundwater with long transit time (PINDER & JONES 1969, CALLES 1985), or as groundwater ion contribution dissolved from different geological formations (APPELO et al. 1983). According to calculations by CALLES (1985), there also is a rapid and large increase in deep groundwater flow toward the stream during stormflow.

2.5.2 Runoff separation by hydrochemical tracers

A wide range of chemically conservative solutes can be measured in all of the storages and transfer volumes of water (recharge and discharge zones) within a catchment system. Solutes have been used for hydrograph separation or as tracers within catchments to infer water pathways and response times. Employed were tracers such as Cl^- (SKLASH & FARVOLDEN 1979, NEAL et al. 1988, MCDONNELL et al. 1991, JENKINS et al. 1994), silica (MAULÉ & STEIN 1990, DURAND et al. 1993), EC (SKLASH et al. 1976, RODHE 1987, DE BOER & CAMPBELL 1990, MCDONNELL et al. 1991) and SO_4^{2-} (HUNTINGTON et al. 1994). The basic premise for the applicability of chemical tracers to the evaluation of hydrologic processes is that the chemical tracers are nearly conservative, and they mark the source waters, as do isotopes (MOLDAN & CERNY 1994). The interpretation of such chemical data demands care, because major sources of some constituents may change with location in the catchment and with time. In other words, the use of water chemistry as a tracer relies on the non-conservative property of water chemistry during flow through a basin. For example, SO_4^{2-} budgets may balance in heavily acidified catchments, whereas Cl^- budgets may not balance in the short term due to interaction with vegetation (MOLDAN & CERNY 1994).

In contrast to the variability of the isotopic content in precipitation, the level of EC, Cl^- and SO_4^{2-} are sufficiently small to be neglected compared to their levels in the streamwater of the study catchments, making them useful as tracers for the separation.

In the present study, Cl^- is chosen as a chemical tracer for the long-term hydrograph separation. Cl^- with EC and SO_4^{2-} is used as a tracer for short-term separation. The details of how these parameters are chosen will be discussed in a latter chapter.

The simple calculation methods for isotopes used above are assumed to be suitable for the chemical separation. Unlike stable isotopes, chemical tracers may only be used to separate water contributing to streamflow from surface flow and subsurface flow, because dissolution of soil minerals occurs rapidly after precipitation enters soil.

Equations 2.11-2.13 or 2.14-2.16 are suitable for dealing with chemical tracers only under specific conditions. The complexity when using chemical tracers originates from three issues:

(1) A realistic value for the tracer concentration of new water is not known. Rainwater reaching the system as overland flow may, as is pointed out by NAKAMURA (1971), FRITZ et al. (1976), and PILGRIM et al. (1979), be enriched in salt by contact with soil particles. In the studies referred to in this section, the composition of the rainwater component has been assumed to be equal to that of rainfall, the composition of which is measured in a few overland flow collectors.

(2) Large amounts of solutes are flushed from the soil into subsurface outflow during the early part of storms that occur after a prolonged period without storm runoff. This happens when newly infiltrated water displaces older water that has resided in the soil for some time, which results in a high solute concentration.

(3) An increase in the concentration of dissolved solids is found with increased contact time of the water with the soil. When precipitation first reaches the watershed soil, the solution of soil minerals happens rapidly. However, the solution rate soon decreases, and days or weeks may pass before effective equilibrium conditions are reached (PILGRIM et al. 1979).

In fact, many studies have found a slight increase in the salt concentration of streamwater at the beginning of the runoff events, before the decrease takes place again. The existence of such a "salt peak" shows that the chemograph of the early stage of an event cannot be explained by dilution only (e.g. MILLER & DREVER 1977), and simple two-component separation is not possible (RODHE 1987). The salt peak may be caused by the flushing of salts by the first groundwater discharge outside the stream, where salts have accumulated due to evaporation before the event (CALLES 1982). In areas where evaporation from soil is important it could, however, also be due to flushing during overland flow on recharge areas (RODHE 1987).

2.6 Interpretation of isotope data with mathematical flow models

As indicated in the previous section, the contribution of environmental isotopes to the understanding of runoff generation includes the tracing (identification) of runoff processes and component flow. Identification of the origin of the water and determination of physical parameters relate to the flow dynamic (YURTSEVER 1995). All these uses of isotopes require a conceptual model to link the observed input data. This section will introduce the corresponding concept of flow models.

2.6.1 Conceptual model of the system

To describe the hydrologic system, simplifications are needed. One simple approach is based on the so-called black-box model, in which only measured inputs and outputs are determined. The isotopic input function is assumed to be transformed by dispersion in the subsurface reservoir. The model must be adapted to processes as they could occur in reality.

The conceptual features of the system are described using the following models (Fig. 2.5):

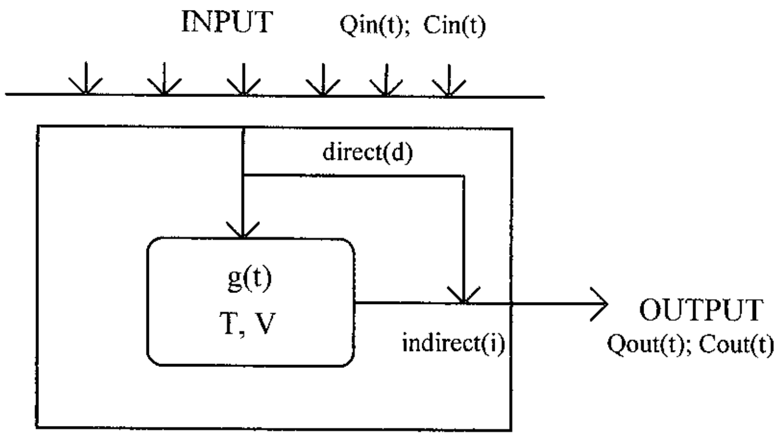


Fig. 2.5 Conceptual model of the system (SCHWARZE et al. 1993, modified)

The average age of the water leaving the system, or residence time (T), is defined as:

$$T = V_m / Q \tag{2.19}$$

where Q is the volumetric flow rate through the system, and V_m is the volume of mobile water in the system.

In this study the average age of water T is determined by mathematical models dealt with in this section, based on isotope data.

According to YURTSEVER (1995), the average age of the water (T) can be defined in theory as follows:

- *Residence time* is the time elapsed since a labeled element has entered the system, which is equivalent to the so-called "age".
- *Transit time* is the time spent by a labeled element between the entry into and the outflow from the system. In other words, the residence time taken at the "exit" is called transit time.
- *Turnover time* refers to bulk flow, and for steady-state flow through a given volume of water, it refers to the ratio of inflow (volumetric flow rate which is equal to the discharge) to the total volume.

The time equivalent for the age of water used in this study is the time of the "exit", namely the transit time. To coordinate the different times obtained by various methods, residence time is also used, meaning the same as transit time.

The volume of mobile water can be determined by

$$V_m = Q \cdot t_o \tag{2.20}$$

where t_o is turnover time.

2.6.2 Determination of Infiltration parameter α

The infiltration coefficient α is given by the following formula (GRABCZAK et al. 1984):

$$\alpha = \left[\sum_W (P_i C_{d_i}) - \overline{C_g} \sum_W (P_i)\right] \cdot \left[\overline{C_g} \sum_S (P_i) - \sum_S (P_i C_{d_i})\right]^{-1} \tag{2.21}$$

where C_{d_i} is the monthly weighted tracer content of precipitation, $\overline{C_g}$ is average annual tracer content of groundwater, P_i is monthly precipitation depth, Σ_W and Σ_S are the sums calculated for all winter and summer months of the observation period. For the winter months i = 11-4 (November-April), for the summer i = 5-10 (May-October). In this approach there is only an average value of isotope concentration accounted for in the groundwater. The above formula can be changed into the following improved form with the $\delta^{18}O$ data of groundwater:

$$\alpha = \left[\sum_W (P_i C_{d_i}) - \sum_W (C_{g_i} P_i)\right] \cdot \left[\sum_S (C_{g_i} P_i) - \sum_S (P_i C_{d_i})\right]^{-1} \tag{2.22}$$

Equation 2.22 represents a real distribution of tracer mass, including an altitude effect, in the infiltration of water, i.e., the weighted tracer mass which transfers to the groundwater every month of the year.

On the other hand, α can be approximated using hydrological data:

$$\alpha = \frac{P_w Q_s}{P_s Q_w} \tag{2.23}$$

where P_w, Q_w, P_s and Q_s are the average precipitation and discharge depths for winter and summer months.

Accordingly, α values obtained from simple isotopic infiltration models are more reliable than those obtained from simple hydrological models based on winter and summer precipitation-discharge relationships (MALOSZEWSKI et al. 1992).

2.6.3 Mathematical Models

Models in hydrology are usually applied to situations other than those described by experimental data. This means that a constructed model is tested and verified for a given set of data and is then used to predict the response of a system in extrapolated situations for which no experimental data exist (FRITZ 1986).

The most suitable mathematical models determining the mean parameters of the long-term transport of tracers in a natural system seem to be black-box models. These models assume a steady travel time distribution of a given tracer in a water-saturated environment.

In the case of steady conditions (V and Q constant) and conservative tracers, the relation between the input and output tracer concentrations may be formulated as a convolution integral:

$$C_{out}(t) = \int_0^{\infty} C_{in}(t - t')g(t')dt' \tag{2.24}$$

where $g(t)$ represents the weighting function of the mathematical flow model (age distribution function), and $C_{in}(t)$ and $C_{out}(t)$ are the input and output concentrations, respectively.

The formula for the calculation of the input function for each month is defined as follows:

$$C_{in}(t_i) = C_i \cdot \alpha_i \cdot \overline{\alpha_i}^{-1} \cdot P_i \cdot \overline{P_i}^{-1} \tag{2.25}$$

where C_{in} is the input concentration; C_i is the monthly $\delta^{18}O$ value of precipitation; δ_i and $\bar{\alpha}_i$ are the monthly and mean monthly infiltration coefficients; and P_i and $\bar{P}_i$ are monthly and mean monthly precipitation.

The following three mathematical flow models have been applied in this study:

The exponential model (EM)

The exponential model (EM) was introduced by ERIKSSON (1958) under the assumption that the exponential distribution of transit time corresponds to a probable situation of decreasing permeability with aquifer depth.

It is characterized by the system response function

$$g(t) = \frac{1}{T}e^{-\frac{t}{T}} \tag{2.26}$$

where T is the mean residence time of water in the system.

The EM is the simplest approach to evaluate the residence time, but does not take into account short-term variations in the amount of rainfall. EM can be used for situations, in which the transit time through the unsaturated zone is negligible in comparison with the total transit time (FRITZ 1986).

The exponential-piston flow model (EPM)

The exponential-piston model (EPM) represents a combination of the exponential model and the piston model. Mathematically it is equivalent to a well-mixed reservoir.

$$g(t) = \begin{cases} 0 & \text{for } t < T(s-1)/s \\ \frac{s}{T}e^{-s\frac{t}{T}+s-1} & \text{for } t \geq T(s-1)/s \end{cases} \tag{2.27}$$

where s is the ratio of the total water volume of the system to the water volume with exponential flow. If s equals 1, the EPM is the EM, and if s is ∞ then the EPM is the piston flow model.

The EPM is a more realistic model which is suitable for catchment studies with moderate amount of rainfall. It allows the model parameter (residence time) to vary from week to week or according to longer time scales.

The dispersion model (DM)

The dispersion model (DM) was first introduced by NIR (1964), but the solution of the dispersion equation used is not adequate for environmental tracers. Therefore the DM as described by MALOSZEWSKI & ZUBER (1982) is applied in the present study

(MALOSZEWSKI et al. 1983). This model is adequate in the case where isotopic output concentration is measured in the output (MALOSZEWSKI et al. 1983).

The distribution of residence times is rendered by

$$g(t) = \left(4\pi \frac{D}{vx}\frac{t}{T}\right)^{-\frac{1}{2}} \frac{1}{t} exp\left[-\left(1-\frac{t}{T}\right)^2 \left(4\pi \frac{D}{vx}\frac{t}{T}\right)^{-1}\right] \qquad (2.28)$$

D is the dispersion coefficient, v the mean transit velocity of water in the system and x the length of the passes of flow.

In homogeneous media, the dispersion constant D/v is approximately proportional to the pore length, and is also roughly equal to the length of the recharge zone measured along the streamlines (MALOSZEWSKI & ZUBER 1982). In the case of an unconfined aquifer, the average distance can be one half of the D/v value, and thus the expected maximum value of the dispersion parameter, D/vx, will be about 2 (MALOSZEWSKI & ZUBER 1982).

The DM is the most flexible model because it allows for a wide variety of residence time distributions. It seems to be the most suitable for natural systems.

A simpler procedure of EM and DM models for stable isotopes

In the case of stable isotopes, the seasonal concentration variations observed in precipitation are of a cyclic nature, mainly due to seasonal temperature variations. A simpler procedure can be applied because the isotopic input and output curves can be approximated by a sinusoidal function within a period of one year ($\omega = 2\pi$/year):

$$C_{in}(t) = A_{in} \sin(\omega t) \qquad (2.29)$$

$$C_{out}(t) = A_{out} \sin(\omega t + \varphi) \qquad (2.30)$$

where A_{in} and A_{out} are the amplitudes of the input and output sine curves, ω is the angular frequency of variation, and φ is the phase lag.

For the exponential model, the following simpler equations can be applied relating the decreasing amplitude of the variation of $\delta^{18}O$ to the mean residence time and phase lag:

$$T = \omega^{-1}[(A_{in}/A_{out})^2 - 1)^{1/2} \qquad (2.31)$$

$$\cos \varphi = (\omega^2 T^2 + 1)^{-1/2} = A_{out}/A_{in} \qquad (2.32)$$

For the dispersion model the equations are as given below:

$$T = \omega^{-1}\left(\frac{-ln(A_{out} \ / \ A_{in})}{D \ / \ vx}\right)^{1/2} \tag{2.33}$$

$$cos\ \varphi = e^{-\frac{D}{vx}\omega^2 T^2} = A_{out}/A_{in} \tag{2.34}$$

It should be pointed out that the results obtained by calculating T from the phase shift φ are rather inaccurate, because mathematically φ is limited to three months, and it already amounts to 2.5 months for a mean transit time of only one year (MALOSZEWSKI et al. 1983).

2.7 Summary

This chapter has given an overview of various definitions of flow components. For long-term isotope separation of a basin, baseflow, rather than individual groundwater data, is suggested to be appropriate for use as old water in the application of a two-component model. Additionally, the correct weighting of precipitation and baseflow, as well as sensitivity analysis are necessary.

The models that are used in this study are summarized as below:

(1) The DIFGA model: a continuous separating hydrograph using daily precipitation and discharge data

(2) The two-component model: identifying the sources of runoff using the chemical or isotopic mass balance equation

(3) Mathematical flow models, including the exponential model (EM), the exponential-piston flow model (EPM), the dispersion model (DM) and the simpler procedure of EM and DM models

3 Study area

3.1 Introduction

The research area is situated in the eastern part of the Kaiserstuhl, Southwest Germany (Fig. 3.1). The size of the drainage area is 1.2 km² for the Rippach catchment and 1.7 km² for the Loechernbach catchment (Fig. 3.2). Apart from minor changes, the Rippach catchment has not been altered in appearance and includes a dense network of irregular footpaths connected to deeply incised roads. In contrast to this, the Loechernbach catchment was altered in 1969-1971 and again in 1975-1976, when large terraces and a regular network of small roads were constructed, using heavy machinery. The consolidation resulted in the construction of very steep slopes at the edges of the new terraces and a dense drainage network.

For the past 20 years, the influence of large-scale terracing upon the hydrological process has been investigated by the comparison of the two catchments. The Institute of Hydrology at the University of Freiburg has been conducting experiments to identify, describe and quantify the hydrological consequences of this change in landscape. The previous investigations show clearly changed runoff characteristics for the large-scale terraced experimental basin Loechernbach where the peak rate of discharge was up to twelve times higher than in the natural basin Rippach, and the base runoff and low flow were reduced significantly (DEMUTH 1992). The average annual groundwater recharge amounts to approximately 15 % of the yearly mean precipitation depth (1972-1975).

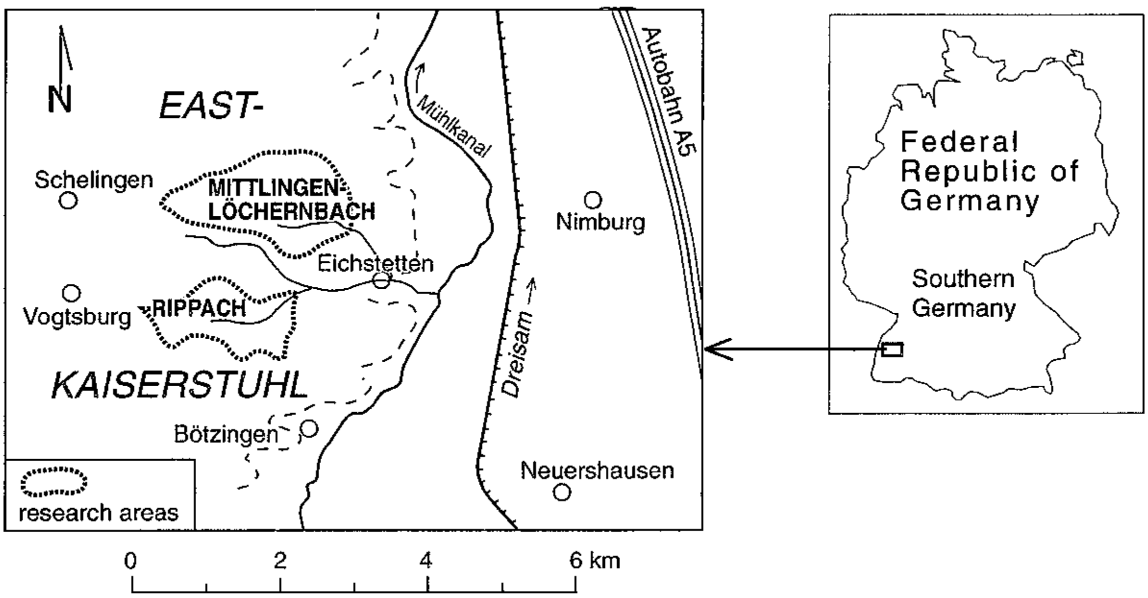
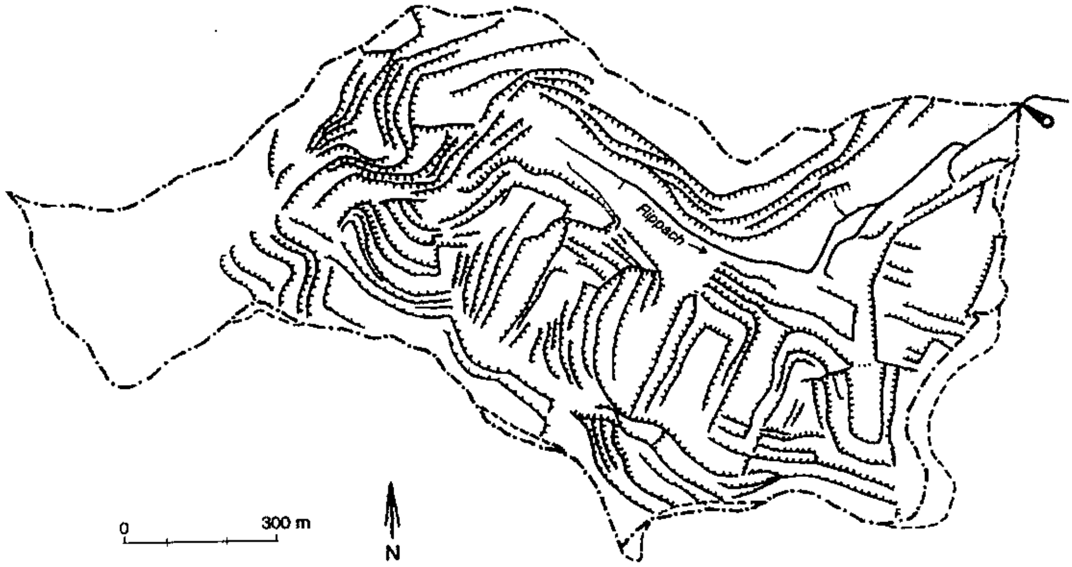


Fig. 3.1 Location of the study area

Rippach



Loechernbach

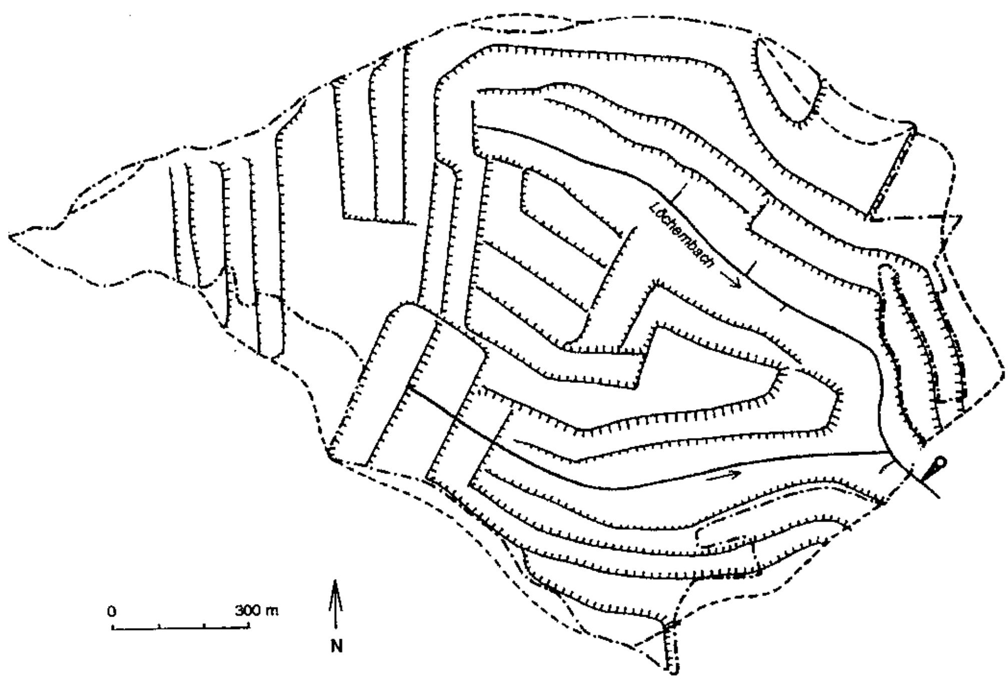


Fig. 3.2 Study sites Rippach and Loechernbach

- - - - - - - Surface catchment area
- - - - - - - Subsurface catchment area
⊥⊥⊥⊥⊥⊥⊥ Terrace

An important earlier investigation of the catchments focuses on hydrograph separation, applying such tools as the HYSTATRI model (baseflow separation using statistical parameters, LUFT 1980), and the USDAHL model (continuous simulation model, VOGELBACHER 1985). These components show that the large-terraced basin has a surface runoff up to 3.5 times higher than that in the unchanged basin; an interflow up to 2.5 times higher, and a baseflow that is about 10 % lower (VOGELBACHER 1985). At the same time, chemical investigations were conducted to examine streamwater and groundwater in the unaltered Rippach (RIEG 1985, WEGEHENKEL 1985 LANG-HAMMER 1988). In the consolidated basin, the loads in winter and the specific loads of O-PO_4^{3-} are typically higher than in the non consolidated basin. In the non consolidated basin, the loads during the summer and the specific loads of NO_3^- are typically higher than those in the consolidated catchment area (RIEG 1988). In addition, the investigations deal with the measurement of soil-water movement and show that a 2-3 m deep loess profile could store up to 700 mm, which is the average annual precipitation depth of the area (MORGENSCHWEIS 1984).

3.2 Physiogeography of the study area

3.2.1 Geology and pedology

The geological development of the Kaiserstuhl is summarized by LUFT (1980) as below:

The uplift and inclined position of the Mesozoic strata and the Tertiary sediments, accompanied with several fracture systems, faults and flexures, resulted from the evolution of the Upper Rhine Graben. After that the Oligocene strata were weathered and eroded until Pechelbronn strata. Intensive volcanism took place at the cross of the fault systems of the Upper Rhine Graben and the Bonndorfer Graben during the Miocene. The tectonic stress acted until the post volcanic time. Since the Upper Miocene the pyroclastic strata were weathered and eroded down to Pechelbronn strata which incline to East and Southeast.

The thick loess sediments were deposited between the glacial epochs over the previously deformed Tertiary Marl. The morphology of the loess traces the surface of the Tertiary Marl. The morphogenesis of the large-scale valleys took place during the Pleistocene, whereas the morphology of loess valleys is an active morphodynamic process.

Because of the shallow soil development with predominant eroded Pararendzina, and in part because of deep ploughing, the substrate area, the upper layer, and the initial material loess together constitute a loose sediment cover. This applies to the Rippach catchment. In the Loechernbach catchment there is no in situ soil development, since the overall relief is artificially altered. These two different soil developments are a main reason for causing the different hydrological processes and water balances observed between Rippach and Loechernbach.

MORGENSCHWEIS (1980) distinguishes four main pedological units in Rippach with the following characteristics:

(1) High terraced area, composed of crude loess with a depth of up to 30 m. The sediment can be described as homogeneous, unbedded and weak consolidated with high carbonate content (> 30 %).

The soil type is silt to a weak clayey silt with strong eroded Pararendizina at a relatively low development depth: 20-30 cm. The groundwater is deeper than 10 m. This area constitutes about 50 % of the basin area.

(2) Valley marginal area, compounded by fluvial reassorted loess (aqueous loess) with a higher clay content. The soil type here is a gley-pararendzina. The groundwater table reaches up to 2 m below surface. This area constitutes about 28 % of the basin area.

(3) Central valley area, composed of aqueous loess with high clay content, the soil type is clayey - weak clayey silt, with groundwater at a depth of 0-1 m. The sediment reaches to a depth of 6-10 m. The filling lies in parts over partly weathered Tertiary Calcareous Marl. The reduction running horizontally 1-3 m below surface is remarkable. The main soil types range from gley-parazendizineas to gley. This pedologic unit covers 11 % of the basin area.

(4) Steep slope to hilltop, mainly composed of volcanic rocks, such as polygonical tuffs, breccias and intrusive rocks. The soil types are shallow-founded A-C soils from sandy-silty to sandy loam. These loess-uncovered areas constitute about 10 % of the whole area.

In the Loechernbach catchment, the high terracing area covers about 62 % of the basin area; the Central valley area and valley marginal area amount to about 20 % of the basin area; the slope area is about 12 % of the basin area (MORGENSCHWEIS & LUFT 1985). The principle soil type corresponds to that of the Rippach catchment. The soils within the altered Loechernbach catchment are mainly young, poorly developed loess soils. The base capacity is very high, and the approximate grain size distribution is: 85 % silt, 6 % sand and 7 % clay. The average effective porosity is 20 % (LUFT 1980). At some sites, evidence of episodic wetness was found.

3.2.2 Hydrogeology and hydrogeography

At the research sites, two hydrogeological units can be distinguished: a pore-aquifer of unconsolidated loess, and beneath this there is a partial joint aquifer (Tertiary, mesozoic and volcanic rock) caused by tectonic stress. Towards the ridge the loess grows thinner. In the valley zone, the upper aquifer has a depth of a few centimeters and is formed by aqueous loess deposits. In some areas, relictics of a B_t-horizon divides the groundwater into several zones.

The Rippach catchment is drained by the Rippach valley-riverlet. Most of the springs of the riverlet have been tapped. The headwaters are cut off by three wells. The riverlet system consists of trenches with almost right angle profiles (depth: 30-70 cm, width: 50-100 cm). The Loechernbach basin is drained by the Loechernbach riverlet, the shape of which is trapezoidal and which was artificially excavated during the terracing. A further branched drainage system was built to drain the precipitation from the asphalt roads and from the terraces (DEMUTH 1992).

3.2.3 Hydroclimatology and hydrology

The Kaiserstuhl region, lying in the Upper Rhine Valley between the Voges and the Black Forest Mountains, has a continental variation of regional climate. Its exceptional position shows the highest air temperatures and the highest annual duration of sunshine (1700 hours, LUFT 1980) in western Germany. Only a few days of snowfall occur during winter, and there is a considerable amount of rainfall during the summer months (MORGENSCHWEIS 1984).

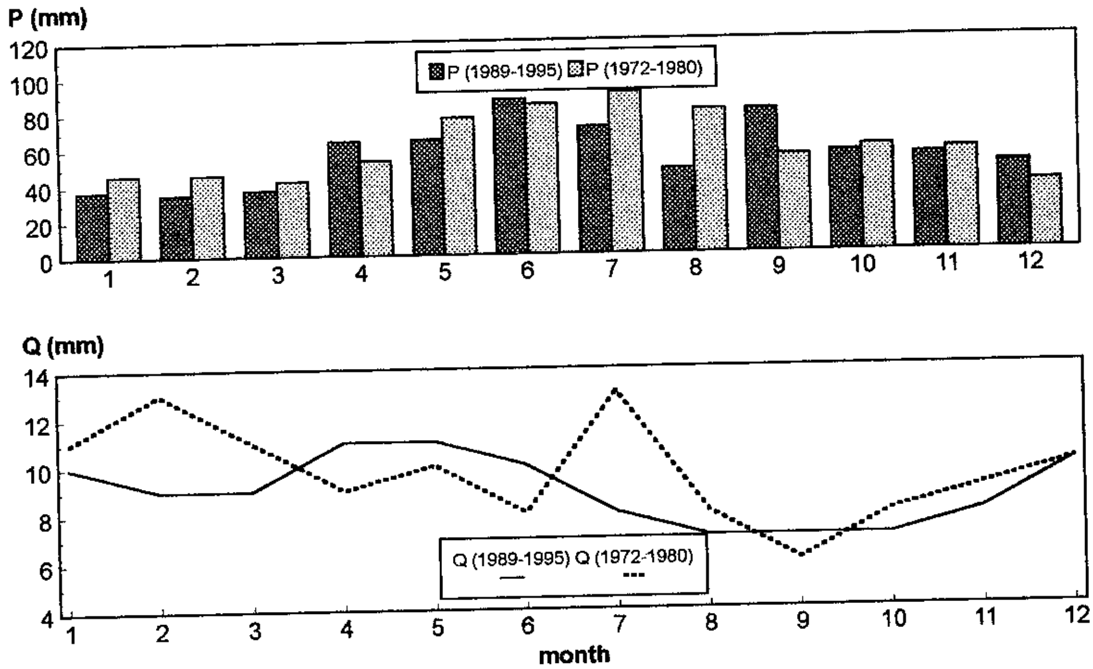
Small-scale thunderstorms often appear over Kaiserstuhl during summer. Therefore, hydrologically separated local rainfall gauging stations are installed in both catchments. In the first water-balance study (1977-1980), the mean annual precipitation was 715 mm in the Rippach catchment and 842 mm in the Loechernbach catchment (BUCHER & DEMUTH 1985), and in the second study (1989-1995), the mean annual precipitation was 682 mm and 632 mm in the Rippach and Loechernbach catchment respectively. The mean summer precipitation constitutes approximately 60 % of the annual precipitation. Precipitation peaked in July for the 1977-1980 period, and during the 1989-1995 period it was binomial with peaks in June and September (Fig. 3.3).

The values for evaporation in the basins are very high (approximately 85 % of the annual precipitation). In addition to the long sunshine duration, the loess is a dominant factor, and has the ability to store large amounts of water. It relates to the hydrogeological nature of catchments built from deep loess layers that the real evapotranspiration could equal the potential evapotranspiration if the amount of rainfall is high enough (LUFT 1980).

The infiltration rate of the Rippach catchment is shown to be 70 %, and the percolation rate is 6 % of the precipitation. According to the Haude formula, the soil evaporation of the terrace site (Mahlkuenzig is used as an example) is much lower than the computed potential atmospheric evaporation computed . The soil evaporation rate in the wet valley is high, which is plausible due to the capillary rise caused by shallow groundwater (MORGENSCHWEIS 1984).

Figure 3.3 shows a comparison of precipitation and runoff for various periods. As is already known, a result of the large terrace is an increase in direct runoff and a corresponding decrease in groundwater in the Loechernbach catchment. In contrast to an equilibrium runoff between summer and winter at Rippach (49 % in summer and 51 % in winter), the seasonal runoff at Loechernbach is more complicated. During wet years such as 1977-1980, the runoff value for summer months (167 mm) is 40 mm higher than for the

Rippach



Loechernbach

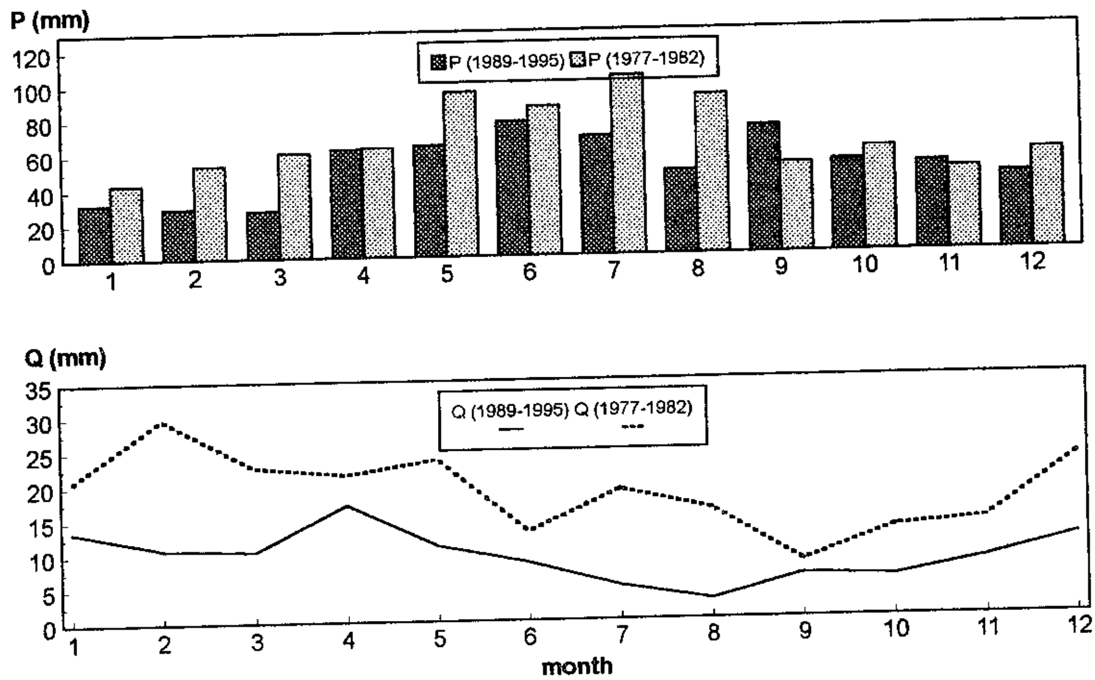


Fig. 3.3 Long-term average precipitation P and discharge Q at the study sites

winter period (127 mm). Similarly during 1994-1995 the runoff for the summer is 34 mm higher than that for the winter, caused by more direct runoff in summer. The higher direct runoff in the extremely wet years, such as 1983, 1985 and 1989, was evident (LEIBUND-GUT et al. 1992). During dry years such as 1990-1993, the winter runoff (72 mm) is almost twice that of the summer (39 mm), caused by higher groundwater level in winter. After comparing the data of last few years, a trend is observed that during normal or dry years the runoff in Loechernbach could be lower.

The average yield of a flood is 54 in winter and 515 l/s km^2 in summer at Rippach, and 443 in winter and 1367 l/s km^2 in summer at Loechernbach. The highest yield was 4694 l/s km^2 at Rippach and 4319 l/s km^2 at Loechernbach (27 July, 1980). The average yield of low water is 2.36 l/s km^2 at Rippach and 0.82 l/s km^2 and Loechernbach (LEIBUNDGUT et al. 1990). Average response times of the systems are 30 and 40 minutes in the Loechernbach and Rippach catchments respectively, based on the analysis of 13 storm events (KELLER 1985).

3.2.4 Land use and anthropogenic effects

For both catchments the main area is used for agriculture and grape growing. There are more agricultural areas in the Rippach catchment, but more grape growing on large terraces in the Loechernbach catchment. Detailed characteristics of the two catchments are summarized in Table 3.1.

After the terracing, the morphology of the Loechernbach catchment was completely changed, and these man-made changes have had an influence on the local climate (ENDLICHER 1979), the ecology of viticulture (BECKER 1977), and the soil structure (MORGENSCHWEIS 1982). Large terraces were built, where at present mainly wine produced. On the other hand, the more or less levelled terraces were partly compacted by the heavy scrapers and graders with which the grading was done. Infiltration and percolation of soil water are considerably reduced; on the other hand, the asphalt roads which are layed out mainly as a pattern of steep roads cover 6 % of the total area and prevent infiltration. They decrease the time of travel and the time a flood takes to peak. These together result in rapid water logging and flooding. To overcome these widespread problems, around 20 % of the total area is equipped with an artificial drainage system to improve the large terrace by deep ploughing. The valleys are drained by a dense tile drain system to lower the groundwater level to 1.2 m below surface (LUFT et al. 1982). However, these drainage measures have had little effect on the peak runoff characteristics, and they have reduced baseflow during the dry weather periods.

In the Rippach catchment the anthropogenic changes are rare. There are no human settlements within the catchment area and only 0.5 % is sealed by asphalt roads. No improvements to water management have been made in this small-terraced catchment area (MORGENSCHWEIS 1984).

Tab. 3.1 Catchment characteristics of Rippach and Loechernbach (KELLER 1985, modified)

Catchment Characteristics		Rippach	Loechernbach
Area		1.2 km^2	1.7 km^2
Length of asphalt roads		1.4 km	23.2 km
Length of subsoil pipes		<1.0 km	12.3 km
Length of open channels		1.3 km	2.1 km
Mole drained areas		-	19.8 %
Bank slope of 45^o		-	12.0 %
Land use	a) forest	18.0 %	5.0 %
	b) wine growing area	48.5 %	66.0 %
	c) agricultural land	33.0 %	13.0 %
	d) small asphalt	0.5 %	6.0 %

3.3 Hydrometrical measurements and analytical mehods

3.3.1 Routine hydrometrical equipment and measurements

The collection of hydrological measurements began in the Rippach catchment in 1972 while no corresponding measurements were made in the Loechernbach catchment. The measurement began in the Mittlingen-Loechernbach during the summer of 1977. The precipitation was digitally recorded and printed out on paper tape every 10 minutes (LAMBRECHT precipitation writer, R 1509-20 H42), and was controlled weekly by the HELLMANN rainfall gauge. The runoff at the output gauging station was digitally recorded every 10 minutes with a water level recorder (OTT XV) and controlled weekly by hook gauge. The groundwater level in the groundwater gauging station was recorded monthly on paper. The estimated error of precipitation is 5 %, of runoff in Rippach 5 % (LUFT 1980), and in Loechernbach 10 % (MORGENSCHWEIS 1980). A detailed description of the instrumentation and its accuracy can be found in LUFT (1980) and LUFT et al. (1981).

3.3.2 Network of sample sites

To carry out a hydrograph separation, a network of sample sites has been established in this study in the Rippach and Loechernbach catchment. The chemical composition of rainwater, streamwater, drainage water, and groundwater was monitored weekly. The samples were collected from September 1994 to September 1995.

3.3.2.1 Rainfall sampling

Rainfall samples employed for $\delta^{18}O$ and chemical analysis were collected weekly. To evaluate the spatial variability of, and the influence of evaporation on, the content of ^{18}O in precipitation in the catchments, two rainfall samples were collected from the standard rain gauges (PR1/PL1), and from two others near the gauge of the catchments (PR2/PL2), which are specially designed two liter bottles beneath a funnel with a diameter of 320 mm (Fig. 3.4). A layer of oil was used to prevent the evaporation of the collected rainfall. The bottles were placed 30 cm deep below the ground to prevent them from being exposed to direct solar radiation.

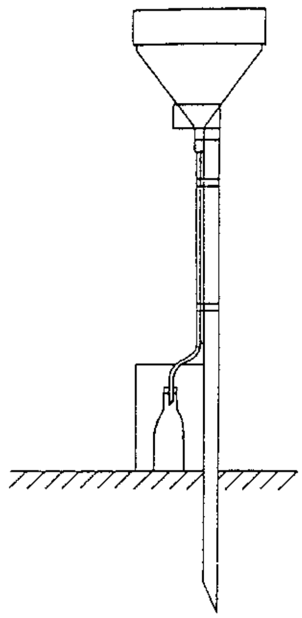


Fig. 3.4 Equipped rainfall collector (HERRMANN et al. 1984)

The precipitation collected at one station, within or close to the basins, was assumed to be representative of the precipitation input to the basins. The difference in altitude within the basin is small in the Loechernbach basin (213-360 m). Though this difference at the Rippach basin is larger (215-510 m), the forested area with the average altitude of about 400 m is only 18% of the total area. The difference between the precipitation stations is comparatively small: PR1 (rain gauge in Rippach) is 255 m above sea level, which is 30 m higher than that of PR2, while PL1 and PL2 (rain gauges in Loechernbach) are both at 215 m.

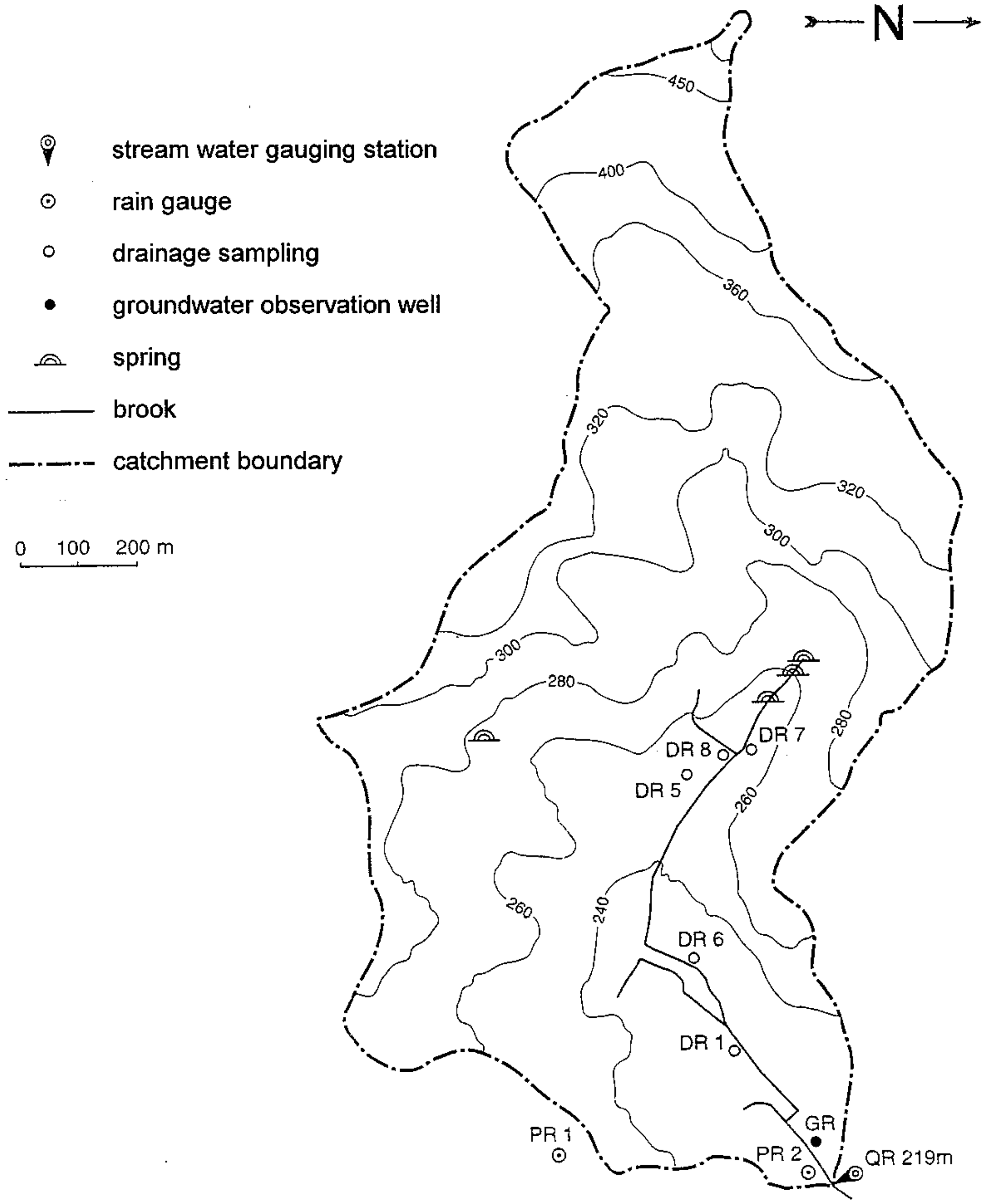


Fig. 3.5 (a) Sampling sites in the Rippach catchment

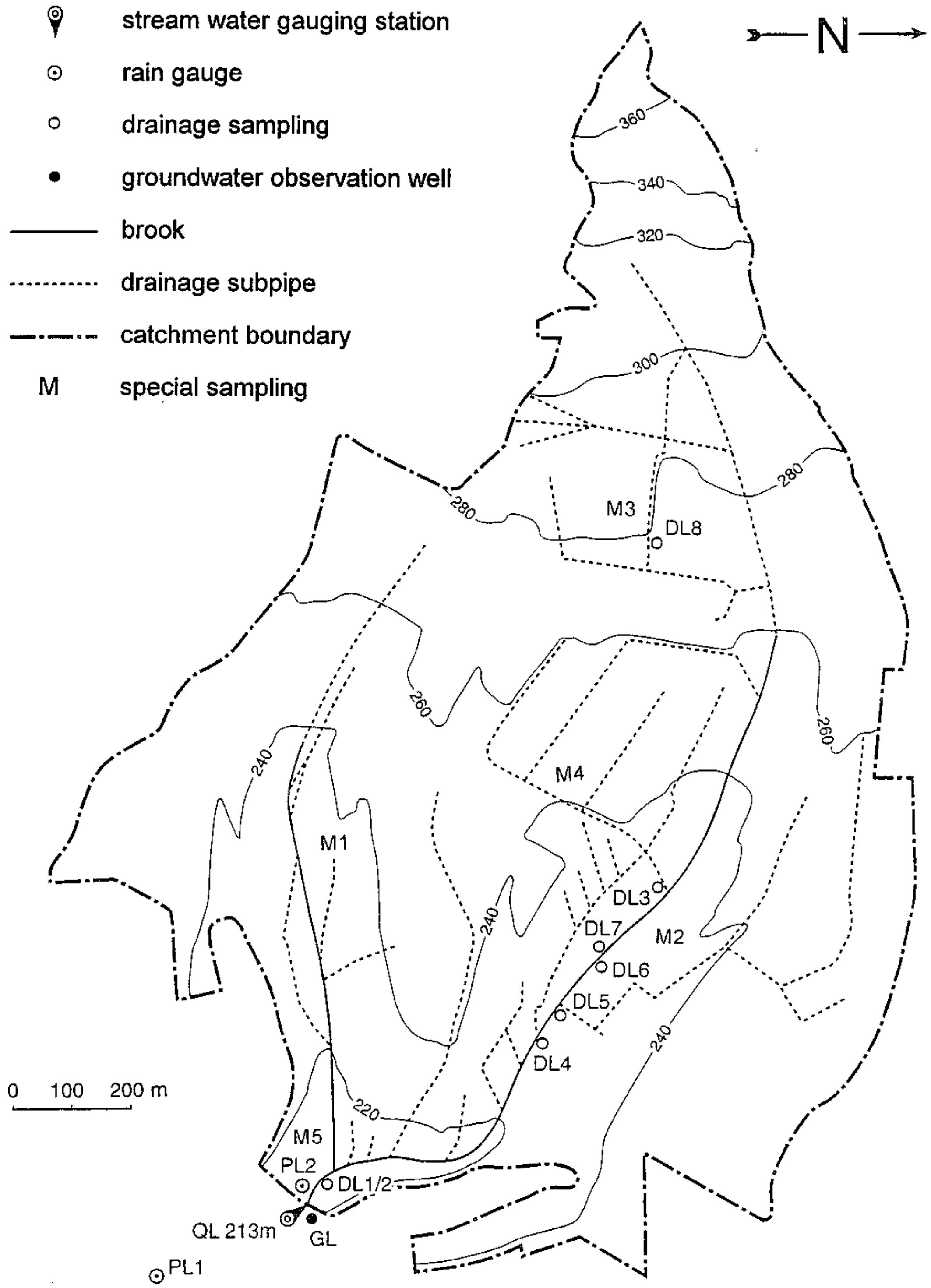


Fig. 3.5 (b) Sampling sites in the Loechernbach catchment

The $\delta^{18}O$ of precipitation is the average value of two simultaneous samples. The correlation coefficients between both samples (Fig. 3.6) within the basins are 0.991 in Rippach, and 0.986 in Loechernbach, based on a weekly sample interval for the year 1995. Moreover, the regressions are generally through the zero point of the axis. Therefore, it is

(a)

(b)

(c)

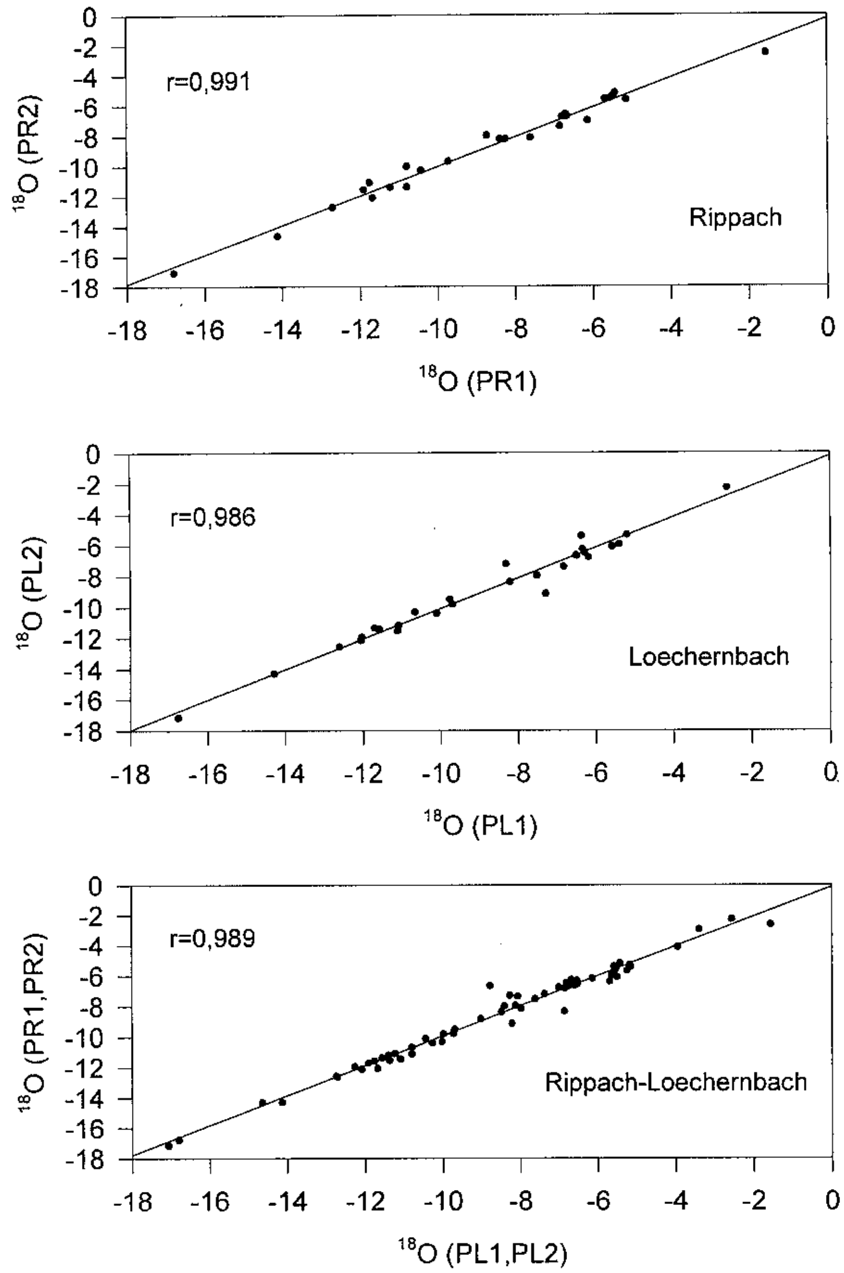


Fig. 3.6 Correlation coefficient of $\delta^{18}O$ values in precipitation between

(a) both gauges within Rippach catchment (PR1/PR2)

(b) both gauges within Loechernbach catchment (PL1/PL2)

(c) both sites taking into account all samples (PR1+PR2 / PL1+PL2) concluded that the spatial difference within these small study sites is sufficiently small that they can be neglected. Furthermore, there was little difference in $\delta^{18}O$ in precipitation between the neighboring catchments (the relationship coefficient is 0.989).

3.3.2.2 Streamwater sampling

The composition of streamwater (*Q*) was monitored by installing two automatic sampling devices at the output of the *R*ippach (Q*R*M) and the *L*oechernbach (Q*L*M) catchments (M: weekly mean values). The weekly average value of streamwater was calculated from 42 samples taken at 4 hour intervals. Streamwater was also manually sampled at weekly intervals in both QR and QL (Fig. 3.5). The automatic sampling devices were installed on October 20, 1994, and shut down on September 4, 1995. They broked off at various times due to freezing in December 1994 and January 1995.

3.3.2.3 Groundwater sampling

The samples were collected with a small hand pump from the groundwater gauging stations near the output of the Rippach (GR, Fig. 3.5 (a)) and Loechernbach catchments (GL, Fig. 3.5 (b)). The groundwater levels are found approximately 1.50 meters below the surface. To get fresh groundwater from the pipes, which are at a depth of 3.6 m, the water was not collected until EC, pH and temperature were stable after 30 to 60 minutes. The waters were regularly collected in the pipes at a depth of 3 m. In field observations, water collected at different depths could hardly be chemically distinguished, as indicated by a profile investigation of EC at 20 cm intervals.

3.3.2.4 Drainage water sampling

A number of groundwater, or mixing water samples, were collected from springs and drainages. Water was collected from these sources and springs for the following reasons:

(1) Springs are main groundwater contributors.

(2) The second order stream of the Loechernbach consists mainly of a number of drainages or springs. Accordingly, the investigation of the hydrological behavior of the drainages and the corresponding influence on the streamflows becomes necessary.

A brief description of the sample sites will be given below:

<u>Rippach</u>

DR1 Drainage close to a second order stream, located at a depth of 1.2 m under cropland. In summer, this drainage dries up.

DR5 Untapped spring, groundwater from a hill slope with small discharge.

DR6 Untapped spring from the hill slope, water is present throughout the year.

DR7 Consists of three tapped springs of the Rippach catchment. During the winter months (November-April), the water contributes to the network of the drinking water supply. During the summer months (May-October), the water remains in the catchment.

DR8 Second-order stream, consisting of groundwater from the head forest area. The origin is not clear.

Loechernbach

DL1 Drainage, mixing water site. The temperature have varied considerably compared with other drainages.

DL2 Drainage, the groundwater from the valley slope.

DL4 Main output of tracer passage of the artificial tracer experiment in April 1995.

DL5 Drainage connected to the main drainage pipe of DL4, located 50 meters from the upflow of DL4.

DL6 Water from various layers, including overland flow after rainfall.

DL7 Typical groundwater site of the Loechernbach catchment, where the water runs all year with stable delivery.

DL8 Drainage connecting to a head spring at Loechernbach.

DL13 Output of a drainage pipe with a diameter of 70 cm, in a place where the water is normally groundwater from the spring of the head catchment, overland flow can enter it after storm events.

3.3.3 Analytical methods

Samples were analyzed for their major anions and cations. Electrical conductivity (EC), pH and temperature were monitored in situ (Tab. 3.2). EC of the water samples was measured at +25°C. According to the manufacturer, the accuracy of the instrument is about 1 %. Rainfall, runoff and groundwater levels were monitored continuously. HCO_3^- was determined by the microprocessor pH/140 measuring instrument, and K^+ and Na^+ were determined by flame photometer, Ca^{2+} was determined by Titriplex-III, Mg^{2+} was calculated by the difference in hardness, and PO_4^{3-} was determined by photometer. Concentrations of major anions such as SO_4^{2-}, Cl^- and NO_3^- were analyzed by ion chromatography.

The main purpose of the analysis of major ions' concentration is to choose suitable tracers for long-term runoff separation. Accordingly, some chemical parameters which were found not to be suitable as tracers were only analyzed for 14 weeks during the period of October to December 1994.

The amount of $\delta^{18}O$ is determined by mass spectrometry. The concentration of $\delta^{18}O$ is expressed as a ratio of the heavy to light isotope ($^{18}O/^{16}O$), and reported in delta units (δ) as per mil differences relative to the international standard (SMOW).

$$\delta^{18}O = \left(\frac{R_{sample} - R_{standard}}{R_{standard}} \right) \times 1000 \qquad (3.1)$$

The uncertainty of mass spectrometric measurements is about $\pm 0.13\ ^{0}/_{00}$.

Tab. 3.2 The investigated parameters

Analysis	Parameters	Time	Measurement
in situ	EC	Sep. 1994 - Oct. 1995	EC-meter
	temperature, pH	Oct. 1994 - Dec. 1994	pH-meter
cation	K^+, Mg^{2+}, Na^+	Oct. 1994- Dec. 1994	flame photometer
	Ca^{2+}		Titriplex-III
anion	SO_4^{2-}, Cl^-,NO_3^-	Oct. 1994- Oct. 1995	ion chromatography
	HCO_3^-		microprocessor pH/140
	$O\text{-}PO_4$	Oct. 1994 - Dec. 1994	photometer
	silica	Oct. 1994 - Dec. 1994	photometer
isotope	$\delta^{18}O$	Sep. 1994 - Oct. 1995	mass spectrometry

3.4 Hydrological data and data interpolation

3.4.1 Hydrological data and missing data interpolation

For the application of the DIFGA model in the study basins, a long time series of daily data was needed, including daily precipitation and discharge for more than ten years.

As mentioned above, the continuous precipitation and runoff monitoring started in the Rippach basin since 1972. In the Loechernbach basin, however, the runoff monitoring began in 1977.

Due to the incomplete data series in both basins during the period of 1981 to 1989, the present study deals with the data of two separated periods: the first period ranges from 1972 to 1980 for the Rippach and from 1977 to 1981 for the Loechernbach; the second is from 1989 to 1995 for both basins. The data between 1981-1989 have been ignored due to vast amounts of missing data.

Precipitation data interpolation

The missing precipitation data during the study year 1972(1977)-1980 and 1989-1995 are listed in detail in Table 3.3. The largest error was a technical one for December 1992 and for April - July 1993 at Breitenweg.

As discussed above, the rainfall gauge stations Mahlkuenzig and Breitenweg are situated next to the basins. They are assumed to be representative of precipitation input to the basins. The precipitation station Breitenweg is located at the same altitude as the lower parts of the Loechernbach basin. The precipitation station Mahlkuenzig is located at an average altitude for the Rippach basin. The same weather systems can be assumed to affect both of them. Their linear correlation coefficient is 0.89, based on the daily data from 1989 to 1992. The correlation coefficient of individual months is satisfactorily high, but October is an exception; for this month the correlation coefficient is only 0.63, because of an extreme single day storm event that only affected the Loechernbach. Excluding this storm event, the correlation coefficient between the two stations is higher than 0.80.

Tab. 3.3 Missing data of precipitation and discharge in both catchments

	year	Rippach	r	Loechernbach	r
discharge	1989			14-17. Feb.	0.80
				21-22. Jun.	0.78
	1990	27-31. Jan.	0.82	24-26. Aug.	0.58
		11-15. Oct.	0.72	16-24. Oct.	0.72
	1992			7-9. Apr.	0.94
				2-5. / 9-10. Aug.	0.76
				2. Oct.-16. Dec.	0.90
	1995	15. Feb.-9. Mar.	0.85		
	year	**Mahlkuenzig**		**Breitenweg**	
precipitation	1990	2. Aug., 1. Sep.		1-2. Nov., 2. Aug., 1. Sep.	
	1991	8-9. Jul.		20-27. Dec., 9-21. Feb.	
				26-30. Sep.	
	1992	17. Jan.-1. Feb.		11. Nov.	
		1-9. Apr.		8-10. Sep.	

Discharge data interpolation

Discharge is calculated with the stage-discharge relationship (the rating equation) presented in detail in the literature (LEIBUNDGUT et al. 1990). The daily discharge is the average of the day's 10 minute discharge values.

Two large periods of missing discharge (Tab. 3.3) occurred at the Loechernbach gauge station from October to December of 1992, and at the Rippach gauge station from February 15 to March 9 of 1995 due to gauge malfunction. Because the various factors affecting runoff vary considerably in a drainage basin during different rainfall events, a direct plot of rainfall for individual storms usually does not produce a satisfactory correlation. Hence, the missing discharge data in the study period have been estimated by an analysis of monthly regressions correlated with the other gauge data in the neighboring catchment. Table 3.3 presents the corresponding correlation coefficients for the separated periods of the missing data. The procedure of interpolation of missing data is as follows:

(1) If no rainfall exists, the missing daily discharge of one or two days can be directly interpolated with the neighbouring gauge.

(2) The missing discharges of days, not in the recession stage, are estimated using the data from the neighboring gauge considering the linear regression correlation of 10 minute values, in the corresponding month. For example, the missing discharge data from June 14-16 at Loechernbach have the correlation coefficient of discharge (10 minute values) between both gauges from the corresponding month as 0.80. The corresponding daily discharge is computed from the average discharge of the 10 minute values.

(3) If data are missing for a period of more than 15 days, the estimation is based on correlation analysis of the annually averaged daily discharge data, such as the missing data from October to December 1992 in Loechernbach, in which r is equal to 0.90.

3.4.2 Hydrological behavior in 1995

During the hydrological year 1995, wet precipitation was recorded. The Rippach and Loechernbach separately received about 818 mm and 740 mm of annual precipitation, of which about 106 mm and 110 mm was discharged as streamwater, respectively. Throughout the year, discharge in the Rippach catchment varies less than that in the Loechernbach catchment (Fig. 3.7). It is surprising that compared to previous annual averages, discharge at the Loechernbach gauge is lower, though precipitation is larger. The base flow or groundwater in the Loechernbach catchment is evidently decreasing annually.

The hydrologic extremes were: maximum daily mean discharges of 25.6 l/s (138.7 l/s) at Rippach (Loechernbach), and minimums of 1.8 l/s (0.7 l/s) at Rippach (Loechernbach). The longest dry period was 24 days, from July 28th to August 20th, and the largest flood was from May 26th to June 1st, during which 83.2 mm of rain caused the flood with the year's maximum daily discharge.

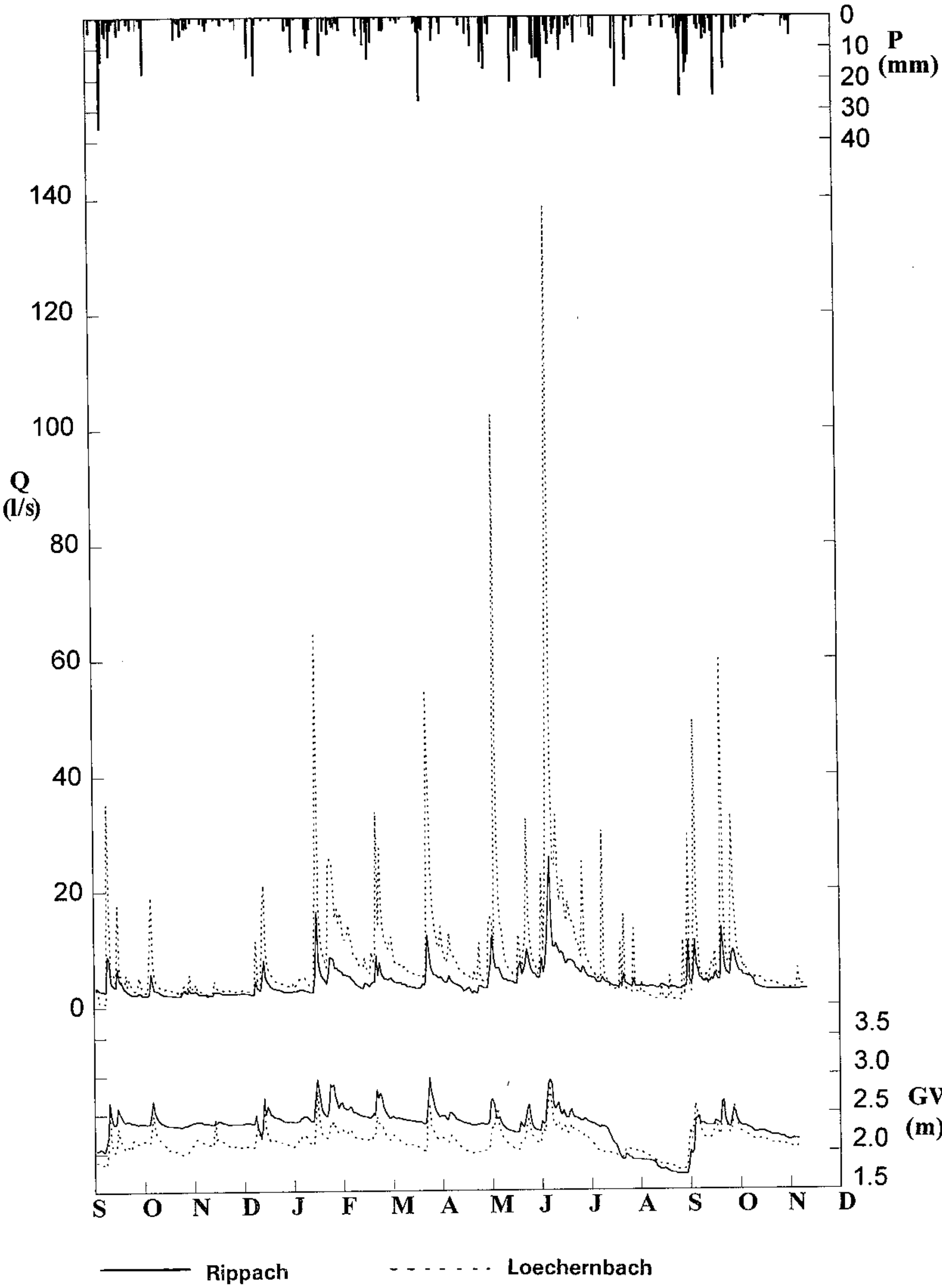


Fig. 3.7 **Variation of water budget components (September 1994 - October 1995)**
P: precipitation, Q: discharge, GW: groundwater

4 Application of the hydrograph separation model DIFGA to the study catchments

The DIFGA model is applied in this chapter to determine the runoff components, storage constants, and storage volumes of the system and to examine the water balance of the basins. Some results are given in LEIBUNDGUT & CUI (1994).

4.1 Selected details of the modeling procedure

The procedure of the program computation has been presented in detail in Chapter 2.3. The data used are daily discharge values for the period from 1972 (1977) to 1980 (1981) and from 1989 to 1995 at Rippach (Loechernbach), and the daily precipitation values from the gauge stations of Mahlkuenzig, Breitenweg and earlier Adler. According to Equation 2.9, a monthly water balance evaluation is used in the program. The exact evaluation of each parameter in Equation 2.9 is needed. The estimation of discharge-stage relationship is relatively precise, and the evaporation and storage values ($ETR+W$) are only residuals of the equation. Only precipitation can cause a visible error. To take into account the systematic error of precipitation measurement, a correction factor of 1.05 for standard gauge precipitation data in both catchments is used. This correction factor is based on the experience gathered with in the previous investigations on the study sites by LUFT (1980), combining the correction of precipitation in DIFGA for different basins. This factor is considered to be suitable for the entire year due to the gauges featuring heating devices in winter. In addition, the snow correction program of the DIFGA model is not considered, because winter at the research basins is generally too mild to permit accumulation of a large snowpack.

Three components for the Rippach (direct runoff, short-term and long-term baseflow) and four components for the Loechernbach (rapid and slow direct runoff, short-term and long-term baseflow) are considered in this study. The separation of the second fast component was not performed in the Rippach catchment because the contribution of this component to this region is insignificant due to lack of artificial subpipes. The most important phase in the DIFGA model is the determination of $CG2$ (the long-term baseflow constant). The actual $CG2$ varies annually. To estimate a representative $CG2$, the annual $CG2$ is determined by means of a long-term hydrograph printed in semi-logarithm diagrams. The final $CG2$ is the average of many annual $CG2$ values. The determination of $CG2$ at Rippach is comparatively easy. This is because the recession curve of that area usually is regular, availed by the rich recharge of groundwater throughout the year, even during dry periods. A problem arises, however, in the Loechernbach catchment where the irregular recession, especially in summer months, results in uncertainty in the under-envelope of the recession. Thus, the estimate of $CG2$ for this region is relatively imprecise.

As an example, Fig. 4.1 shows the separation procedure of the DIFGA program for the summer months of 1979. During May-August the peak runoff which occurred repeatedly reduced the storage reservoir in the large-terraced Loechernbach catchment, and baseflow of this basin is also conspicuously low.

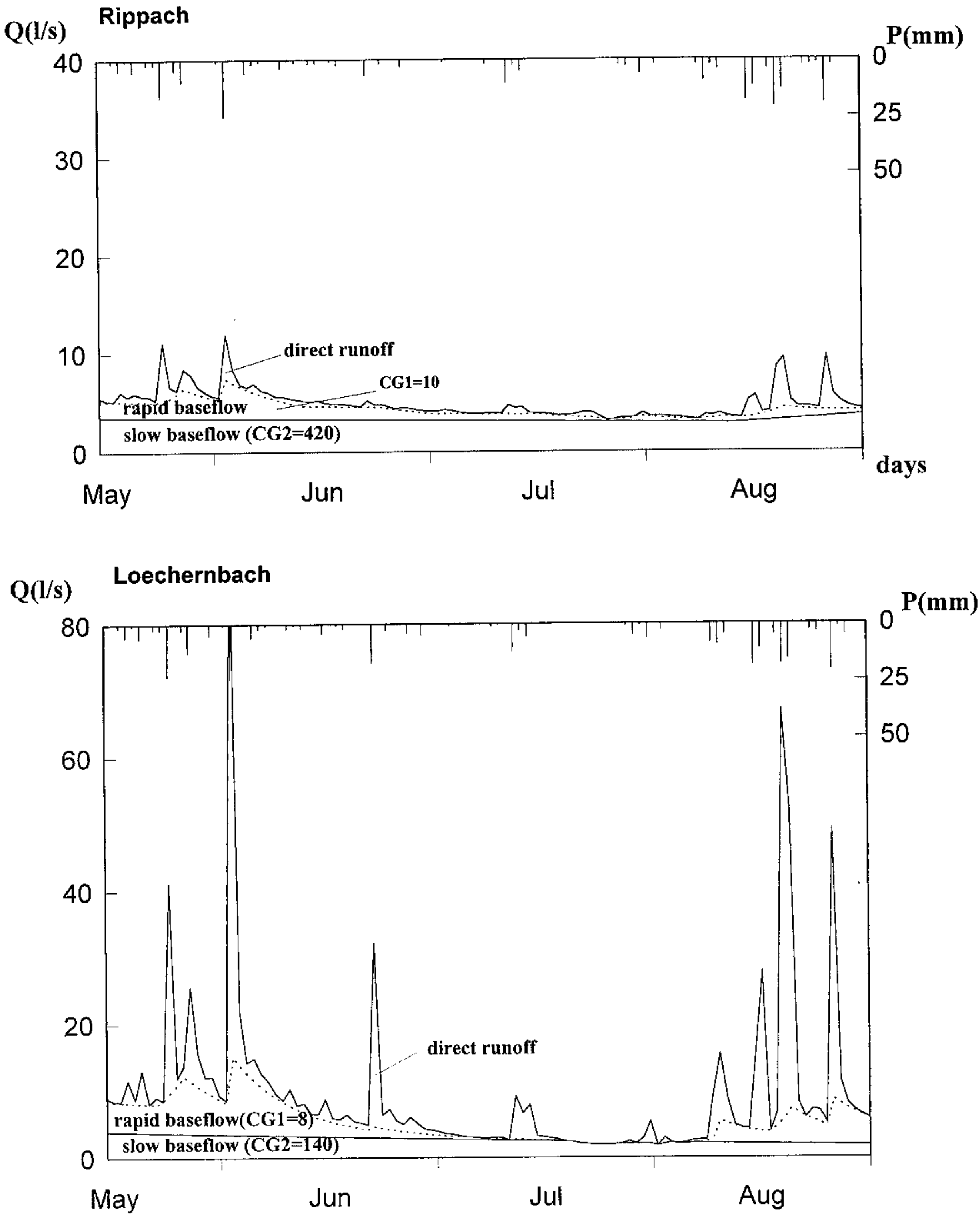


Fig. 4.1 Hydrograph separation in the study basins (1979)

4.2 Results

4.2.1 Runoff formation

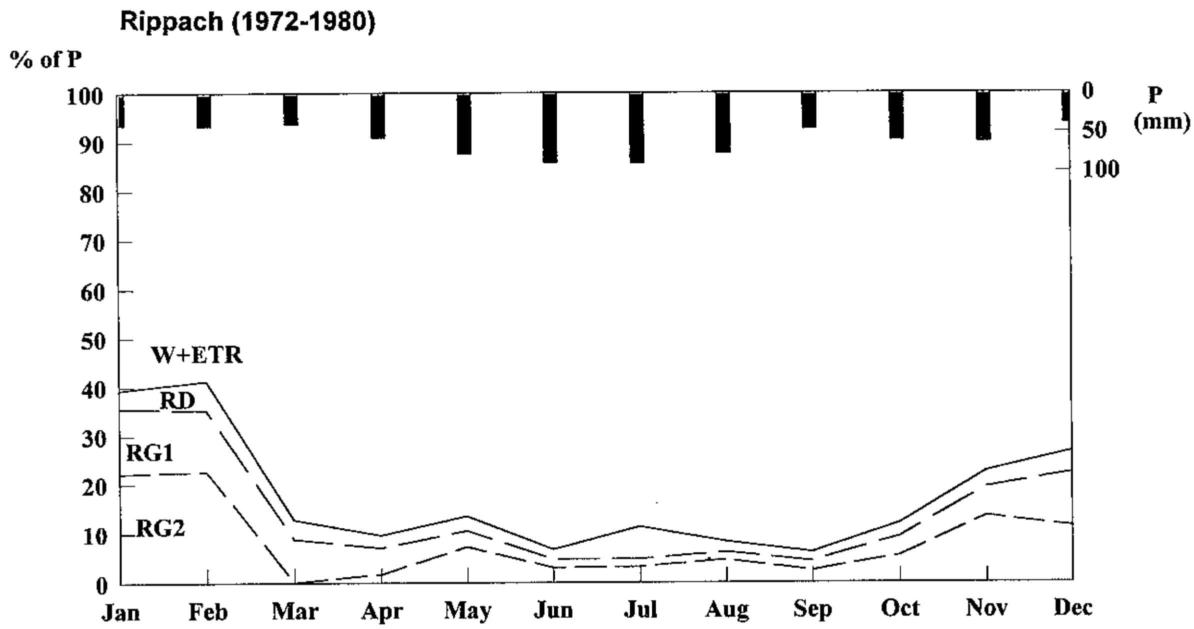
According to the results of the DIFGA model, Figure 4.2, based on the monthly precipitation and the corresponding runoff coefficients, shows only a small proportion of the effective precipitation versus the total precipitation at both study sites. The reason for the small effective precipitation is the extremely porous structure of loess, which allows a huge amount of precipitation to be stored and then to be released into the atmosphere through the ground surface or vegetation (DEMUTH 1992). Also, the vegetation cover results in an increase of the infiltration capacity. One reason for the lower runoff-effective precipitation in the Rippach catchment (16 % for the first period), compared to that of the Loechernbach catchment (26 %), is the increase in cropcover across the basin.

Based on the data of 1977 to 1980, the highest monthly runoff coefficient occurs during January in the Loechernbach catchment, and one month later in the Rippach catchment, along with the highest monthly discharge. The largest part of the runoff formation in this season is from base flow (the sum of the short-term and long-term baseflow). The lowest monthly effective precipitation occurs during September.

The annual course of the different components demonstrates that the behavior of the direct runoff coefficient *RD* is relatively constant throughout the year. The delayed subsurface runoff coefficient *RD2* (or hypodermic flow) in the Loechernbach catchment is small for the period 1977 to 1980 (Fig. 4.2b), so the short-term baseflow coefficient *RG1* plays a more important role. *RG1* and long-term baseflow coefficient *RG2* have distinct variations from summer to winter, and their remarkable recession in summer months is clearly caused by the influence of evaporation on the storage volume. It is mentioned that the *RG2* in the study basins has distinct annual variation. This means that the groundwater is easily influenced by evapotranspiration under the present conditions of climate and soil. Intensive agricultural use is responsible for this sensitivity.

The fluctuation of total runoff coefficients is large in winter. During the period of December to January the catchments' runoff amounts differed by over 20 %. Storage capacity played an important role, at this time allowing more winter precipitation to be stored at Rippach than at Loechernbach. Thus more water is available to be brought back into the atmosphere by evaporation in summer at Rippach than at Loechernbach.

Unlike the period of 1972 to 1980, the runoff formation in the small-terraced Rippach catchment for the period of 1989 to 1995 appears with a peak in April instead of February (Fig. 4.3a). This period is relatively dry, so the contribution of direct runoff *RD* was reduced, and the recharge of groundwater increased. During this period, the total runoff in the Loechernbach catchment is maintaining the earlier monthly distribution, with only *RG2* reduced, suggesting an annual decrease of long-term baseflow.



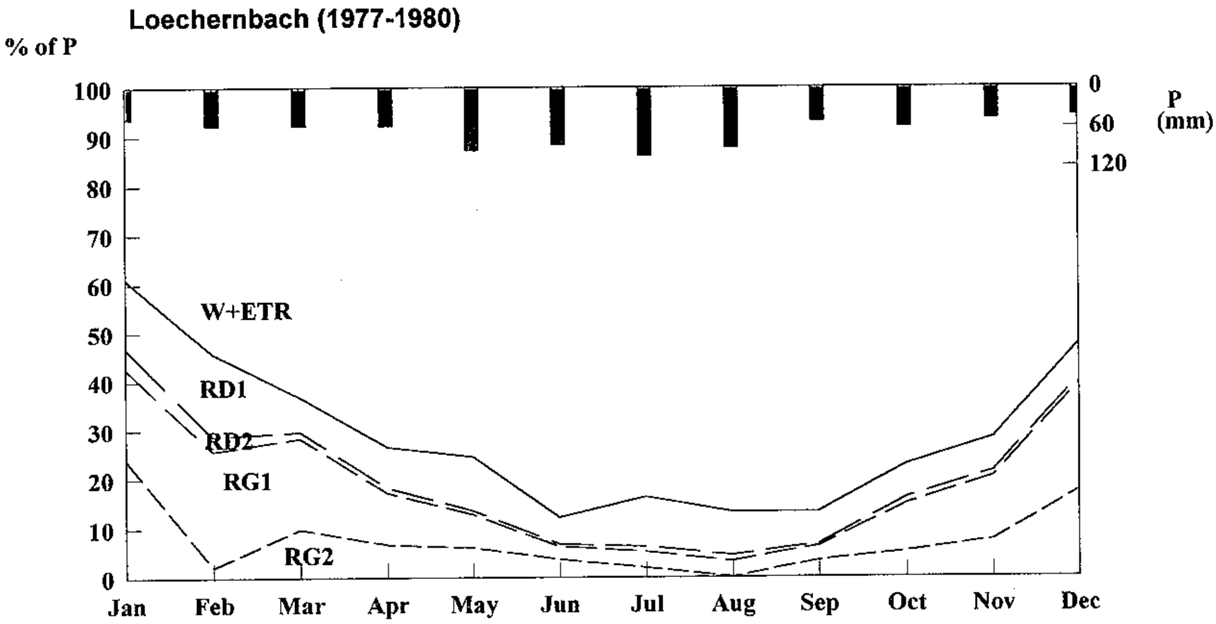


Fig. 4.2 **The monthly precipitation and its percentage distribution to ETR (evapotranspiration) as well as to the various runoff components in the research catchments Rippach and Loechernbach 1972/1977-1980**

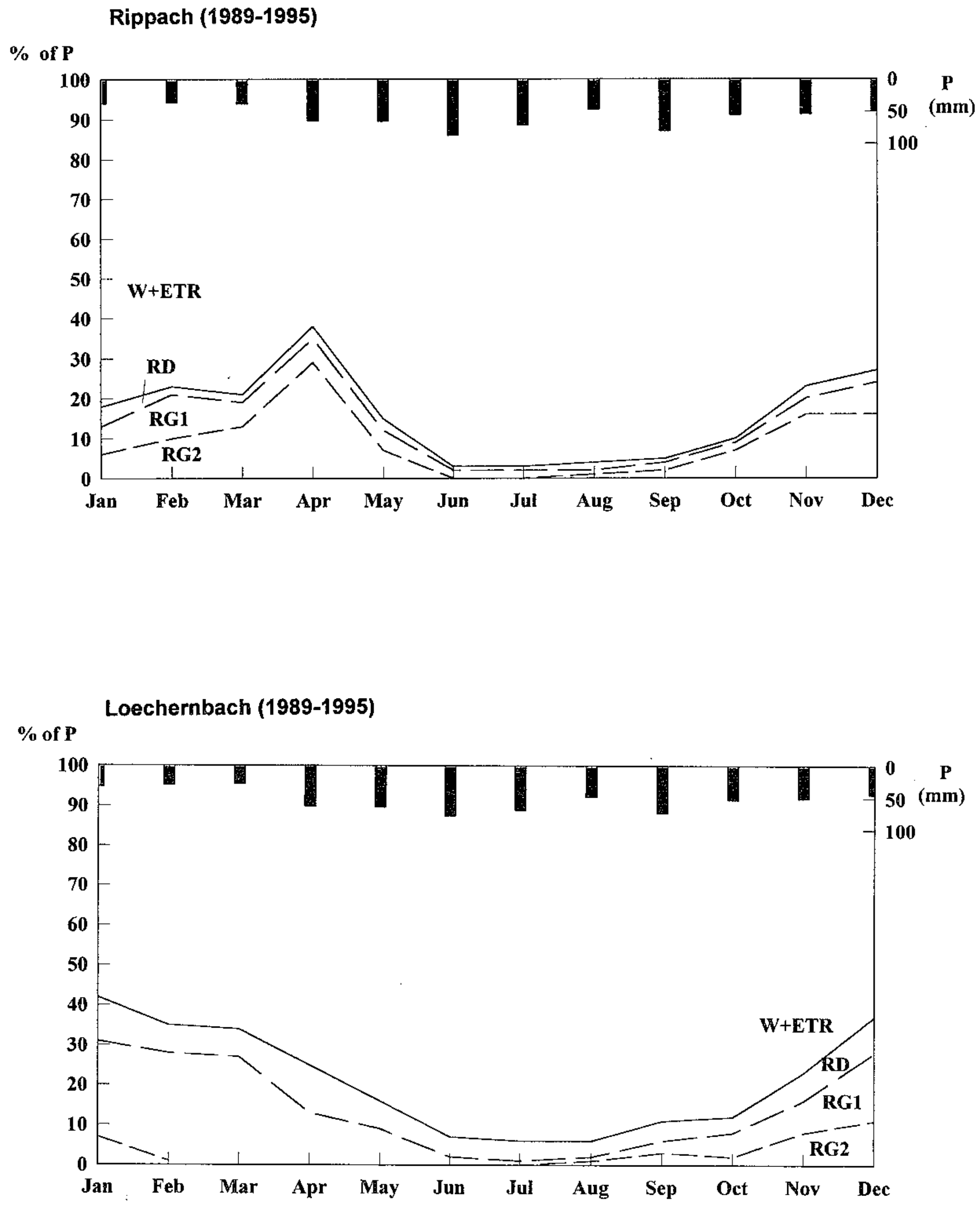
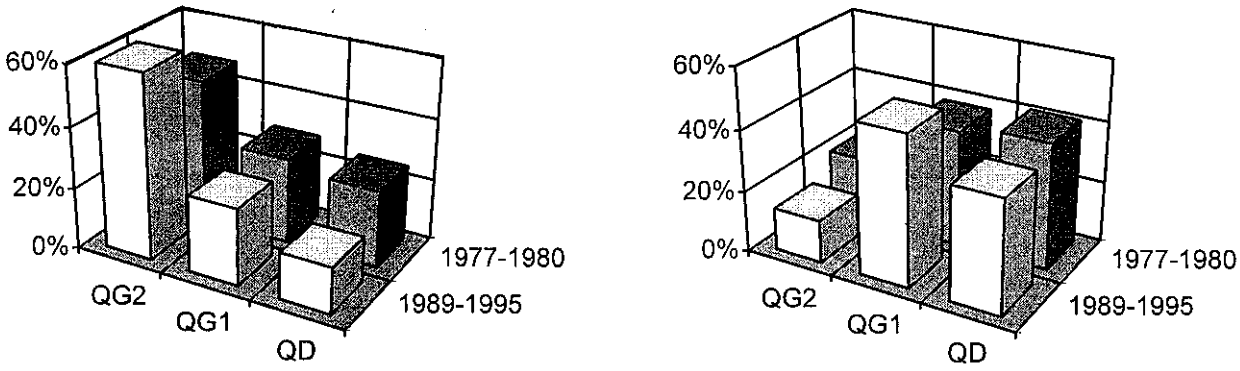


Fig. 4.3 The monthly precipitation and its percentage distribution to ETR (evapotranspiration) as well as to the various runoff components in the research catchments Rippach and Loechernbach 1989-1995

4.2.2 Runoff concentration

When applying the DIFGA model, runoff is separated into the following components: direct runoff (*QD*), short-term baseflow (*QG1*) and long-term baseflow (*QG2*). Based on the data from 1977 to 1980, long-term base flow made a more important contribution to total runoff in Rippach (47 %) than in Loechernbach (21 %); direct runoff amounted to only 25 % of the total runoff. In other words, 75 % of the total runoff originated from underground water in this basin. On the other hand, direct runoff at Loechernbach was as high as 41 %, and the short-term baseflow was 38 % of the total runoff, 10 % higher than in Rippach (Fig. 4.4).



(a) Rippach **(b) Loechernbach**

Fig. 4.4 Average proportion of various runoff components to total runoff (%)
QD: direct runoff, QG1: short-term baseflow, QG2: long-term baseflow

Long-term baseflow from 1989 to 1995 decreased from 21 % to 14 % at Loechernbach, and the corresponding short-term baseflow increased from 38 % to 49 %. The direct runoff fraction rarely changed there. In contrast to this, *QG2* at Rippach increased and the corresponding *QD* decreased.

The subdivision of the annual contribution of the runoff components into their winter and summer subcomponents (Tab. 4.1) shows that the amount of direct runoff in both catchments is much higher in summer than in winter, in particular at Loechernbach where direct runoff is the dominant factor accounting for 50 % of total runoff. In winter the situation reversed; short-term base flow is important. While the proportion of long-term base flow to total runoff at Loechernbach decreases annually, the *QG2* fraction at Rippach seems to increases annually. Such a change is particularly clear during the summer months, indicating the influence of the large-terracing in the Loechernbach.

Tab. 4.1 Comparison of the seasonal runoff components in the study basins for different periods in mm
QD: direct runoff, *QG1*: short-term baseflow, *QG2*: long-term baseflow

		Summer		Winter	
period	component	Rippach	Loechernbach	Rippach	Loechernbach
1977	*QD*	22.6 (32.8%*)	48.7 (50.0%)	12.7 (17.9%)	43.3 (33.9%)
–	*QG1*	12.0 (17.4%)	23.3 (23.8%)	27.0 (38.0%)	61.6 (48.2%)
1980	*QG2*	34.3 (49.8%)	25.6 (26.2%)	31.3 (44.1%)	22.9 (17.9%)
1989	*QD*	7.0 (14.2%)	19.2 (48.9%)	8.6 (15.1%)	22.3 (31.7%)
–	*QG1*	10.3 (20.8%)	14.7 (37.5%)	16.7 (29.4%)	39.3 (55.0%)
1995	*QG2*	32.0 (65.0%)	5.3 (13.6%)	31.5 (55.5%)	9.9 (13.8%)

*( ) = the percentage of the total runoff

4.2.3 Water balance, storage constants and storage volumes

In Table 4.2a-b the water balance, storage constants, and storage volumes in both catchments are listed. The higher runoff of the first period at Loechernbach is due to the fact that annual precipitation is higher than Rippach by 130 mm. Similar runoff of both basins during the second period implies a reduction of the runoff in the large-terraced basin, though the smaller amount of annual precipitation could also be a reason. The runoff behavior of each catchment is quite different. In July and August the direct runoff amounted to more than 65 % of total runoff in the Loechernbach catchment, compared to only 13 % in the Rippach catchment.

As mentioned above, direct runoff is an important component in the Loechernbach. For this reason the delayed direct runoff component - interflow - is separated using DIFGA during the first investigation period. The corresponding storage constant amounts to three days, but only accounts for 1 % of the total annual precipitation.

Based on the data of the first period, the storage constant of the long-term baseflow at Rippach is 3 times higher than that of Loechernbach (420 days vs. 140 days). The storage constants for the rapid baseflow are 10 days at Rippach and 8 days at Loechernbach. The storage constants in DIFGA are initially expected to estimate residence time. The estimate of residence time with the DIFGA model is quite inaccurate, which Chapters 7 and 8 will discuss in detail. As for surface runoff, *RD*, its residence time is impossible to estimate by DIFGA, because DIFGA uses daily data, as should be evident, so the computation of a smaller time scale than one day is impossible. The storage constants of slow direct runoff

Tab. 4.2a Mean values of the runoff components related to the water balance of the Rippach using DIFGA (C: storage constant, Vg: storage volumes)

		mm/year	% of P	% of Q*	C(days)	V_g(mm) V_{max}	V_{min}	V_{mean}
1972-1980								
P	(precipitation)	714.9						
E	(evaporation)	604.6	84					
Q	(runoff)	109.7	16					
RD	(direct runoff)	24.7	4	23	≅0			
$RG1$	(rapid base flow)	34.7	5	31	10	6.3	0.0	0.9
$RG2$	(slow base flow)	50.9	7	46	420	133.7	25.3	58.1
1989-1995								
P	(precipitation)	682.1						
E	(evaporation)	575.9	84					
Q	(runoff)	106.2	16					
RD	(direct runoff)	15.6	2	15	≅0			
$RG1$	(rapid base flow)	28.4	4	25	12	7.5	0.1	0.9
$RG2$	(slow base flow)	55.8	8	60	550	216.8	42.3	95.8

* corresponding: to *QD. QG1 and QG2*

Tab. 4.2b Mean values of the runoff components related to the water balance of the Loechernbach using DIFGA (C: storage constant, Vg: storage volumes)

		mm/year	% of P	% of Q*	C(days)	V_g(mm) V_{max}	V_{min}	V_{mean}
1977-1980								
P	(precipitation)	842.7						
E	(evaporation)	622.2	74					
Q	(runoff)	225.5	26					
$RD1$	(rapid dir. runoff)	82.3	10	37	≅0			
$RD2$	(slow dir. runoff)	10.6	1	4	3	2.9	0.0	0.1
$RG1$	(rapid base flow)	86.2	10	38	8	6.6	0.0	1.7
$RG2$	(slow base flow)	41.4	5	21	140	53.9	8.0	19.8
1989-1995								
P	(precipitation)	632.4						
E	(evaporation)	518.8	82					
Q	(runoff)	110.7	18					
RD	(direct runoff)	41.5	7	37	≅0			
$RG1$	(rapid base flow)	56.4	9	49	9	13.9	0.0	1.3
$RG2$	(slow base flow)	15.8	2	14	160	21.6	1.7	6.7

* corresponding: to *QD. QG1 and QG2*

obtained from the DIFGA model are also imprecise. Average values obtained from 13 rainfall events by LUFT (1980), of 30 minutes in the Loechernbach and 40 minutes at Rippach, could be used for the first rapid component.

The annual storage constants of *CG2* (long-term baseflow) during the second period are somewhat longer than that during the first period due to some dry years during the early 1990s. At Rippach the storage constant increases by 130 days and at Loechernbach by 20 days. The storage constant of *CG1* (short-term baseflow) increases from 10 to 12 days at Rippach and from 8 to 9 days at Loechernbach.

Using Equation (2.2) the maximum, minimum and average storage volumes of each component are calculated. These storage volumes mark the swaying behavior of the runoff effective storage capacity and are suitable for the evaluation of the available groundwater connected with the runoff formation of both baseflow *RG2* and *RG1*. The corresponding storage volume in Rippach is three times larger than in Loechernbach from 1977 to 1980. Remarkably, the rapid base flows in both catchments hardly contribute to the storage volumes (0.9 mm at Rippach and 1.7 mm at Loechernbach, Tab. 4.2), even though this component is the most important one in the Loechernbach catchment. Compared to the earlier period, the storage volume of long-term baseflow *RG2* from 1989 to 1995 increased significantly at Rippach and decreased at Loechernbach.

4.2.4 Influence of terracing on hydrological processes

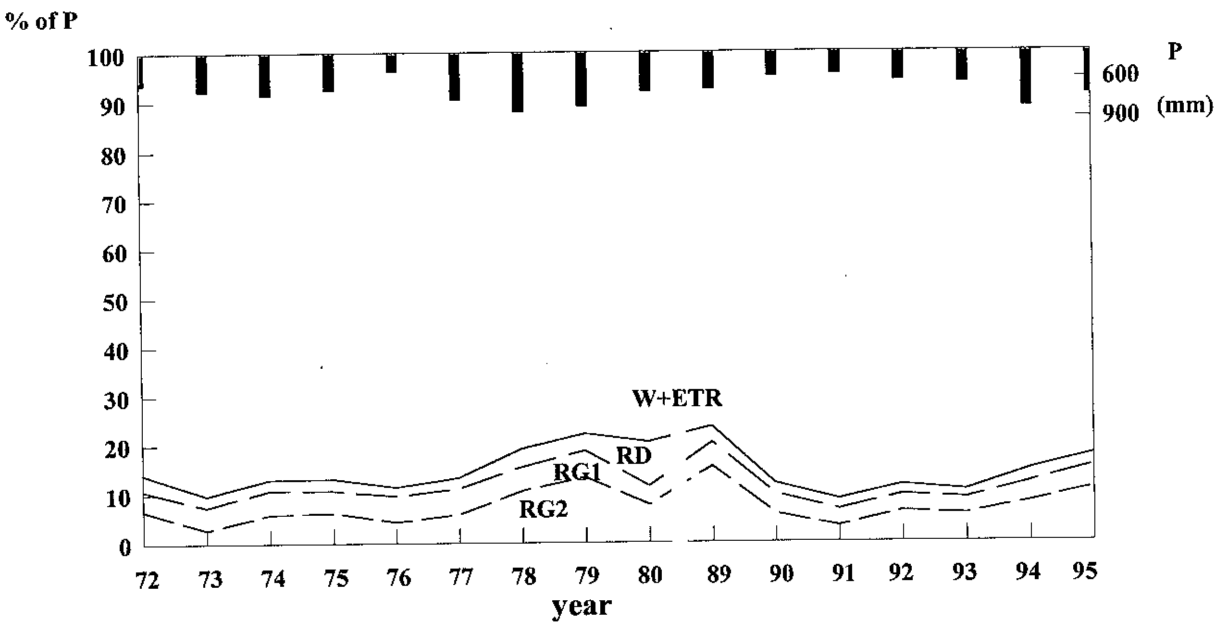
Terracing has a great influence on runoff formation at Loechernbach, causing unbalanced runoff formation throughout the year, increase direct runoff and lower groundwater levels. The difference between the catchments is more evident in the winter months, mainly due to the increase in rapid baseflow. This trend can also be seen in Figure 4.5, which shows a fluctuation of runoff coefficients over the years. At Rippach only long-term baseflow coefficient *RG2* varied considerably year after year. *RG2* increases during wet period, decreases during dry period, and is the major contributor to annual fluctuation of the total runoff coefficient. In the altered catchment this variation of runoff coefficients is quite complex. During the last period, *RG2* seems to get smaller, even during the wet years of 1994 and 1995.

4.2.5 Application of the DIFGA model for the hydrological year 1995

To compare the results obtained by DIFGA and by experimental investigations, the results of DIFGA for 1995 are calculated separately.

As presented above, an important step in the separation using DIFGA is the determination of a storage constant of long-term baseflow (*CG2*), the storage constant of slow base flow. The separation in long wet periods is often critical due to the long distance between the supporting points for the construction of long-term baseflow *QG2* (GURZ et al. 1990). Hence, the DIFGA program suggests that reference be made to groundwater level data in determining the *CG2* (Chapter 2.3).

Rippach



Loechernbach

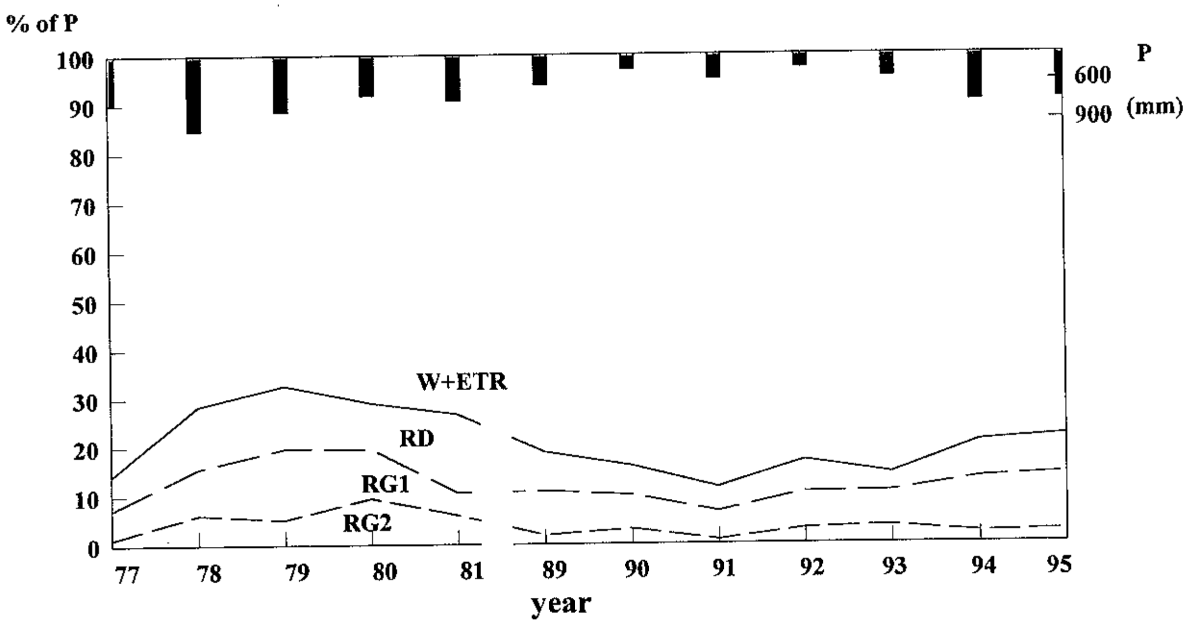


Fig. 4.5 Yearly fluctuation of runoff coefficients

For this purpose, the groundwater level data at Rippach and Loechernbach for 1995 have been analyzed and the daily values computed. Based on these data, the following relationship between *QG2* and the groundwater level *L* was calculated. The corresponding equations are as below:

$$QG2 = 2.44\, L^{0.30} \quad \text{(Rippach)} \tag{4.1}$$

$$QG2 = 0.27\, L^{1.36} \quad \text{(Loechernbach)} \tag{4.2}$$

The correlation coefficients between *QG2* and *L* during the recession are 0.94 at Rippach and 0.92 at Loechernbach. According to Equations 4.1 and 4.2, *QG2* is determined directly from groundwater data. The corresponding results are listed in Table 4.3, and compared to results obtained without the groundwater data. Evidently, Approach 2 results only in a new distribution of proportions between the short-term and long-term baseflow values; the proportion of direct runoff versus total runoff is hardly influenced. In this case, the proportion of long-term baseflow at Rippach increases, while Loechernbach behaves opposite. The storage constants of 1995 are presented in Tab. 4.4. These results will be discussed in a latter chapter compared to experimental results.

Tab. 4.3 Hydrograph separation based on the DIFGA model for the year 1995
Approach 1: *QG2* determined by *CG2,* Approach 2: *QG2* determined by Equations 4.1 and 4.2

		P	*ETR*	*Q*	*RG2*	*RG1*	*RD*	*QG2*	*QG1*	*QD*
Rippach		799.9	609.9	115.9						
Approach 1	mm				93.2	33.4	18.9	66.0	31.1	18.9
	% of P/Q				12%	4%	2%	57%	27%	16%
Approach 2	mm				148.5	24.4	17.0	77.1	21.7	17.0
	% of P/Q				19%	3%	2%	67%	19%	14%
Loechernbach		812.7	615.4	169.7						
Approach 1	mm				24.7	93.8	22.8	20.9	88.9	60.0
	% of P/Q				3%	12%	3%	12%	52%	36%
Approach 2	mm				38.1	96.4	62.7	14.2	92.8	62.7
	% of P/Q				5%	12%	8%	8%	55%	37%

Tab. 4.4 Storage volumes based on the DIFGA model for the year 1995

Components	Rippach	Loechernbach
direct runoff	≅0.00 mm	≅0.00 mm
rapid baseflow	0.71 mm	2.54 mm
slow baseflow	116.20 mm	6.22 mm

4.3 Comparison of the results with other catchments

The DIFGA model has been used in more than 100 basins with areas between 0.5 and 300 km^2 (SCHWARZE et al. 1991). As examples the results for the Lange Bramke catchment (SCHWARZE et al. 1991), for the Wernersbach (SCHWARZE 1996) and for the Rietholzbach catchment (KOENIG et al. 1993) are presented in Table 4.5. The runoff coefficients in these catchments are larger than in the Kaiserstuhl region, and precipitation there turns mainly into direct runoff, while the dominant runoff coefficient is long-term baseflow *RG2* at Rippach, and short-term baseflow *RG1* at Loechernbach. Various explanations of these differences between the catchments can be given. One main reason might be a difference of soil types. The soil of Rietholzbach consists mainly of gley (60%), whereas Lange Bramke and Wernersbach are, for the most part, Braunerde (KOENIG et al. 1993), and the soil of the Kaiserstuhl consists mostly of silt. On the other hand, Kaiserstuhl region receives less precipitation; the mean slopes in the Kaiserstuhl are less than 3° (Lange Bramke / Rietholzbach around 12.5°, Wernersbach 3°).

Tab. 4.5 Comparison of the results of various catchments using DIFGA*

		P	*ETR*	*Q*	*RG2*	*RG1*	*RD*	*RD1*	*RD2*
Lange Bramke	mm	1230	530	700	123	247	321	237	84
	% of *P*		43	57	11	20	26	19	7
Rietholzbach	mm	1600	554	1046	95	414	537	91	446
	% of *P*		35	65	6	26	34	6	28
Wernersbach	mm	907	651	256	29	91	135	101	34
	% of *P*		72	28	3	10	15	11	4
Rippach**	mm	715	605	110	51	35	25	-	-
	% of *P*		84	16	7	5	4	-	-
Loechernbach**	mm	843	622	226	41	86	93	82	11
	% of *P*		74	26	5	10	11	10	1

* other catchment data are from KONNIG et al. (1993)

** data from 1972/1977 to 1980/1981

Direct runoff is 12 % of the total runoff at Lange Bramke and 43 % at Rietholzbach. 17 % of this fraction at Rippach is close to that at forested area Lange Bramke. Direct runoff accounts for 41 % of the total runoff at Loechernbach, which is similar to the figure at Rietholzbach, which has a small forested area and large agricultural lands. The comparison of both fast component fractions demonstrates that interflow is the dominant flow at Lange Bramke, where it constitutes 19 % of the total runoff, while this fraction at Rietholzbach is only 6 %.

Seasonal runoff coefficients of base flow in the Kaiserstuhl sites differ significantly from those in the other catchments. For example, the recession of runoff coefficients in summer at Lange Bramke and Rietholzbach is mainly caused by the large decrease of short-term baseflow *RG1*, both *RD* and *RG2* (delayed direct runoff and long-term baseflow) at these sites varied little with the seasons. In the Kaiserstuhl, not only *RG1* but also *RG2* is highly influenced by evaporation, suggesting that the dominant groundwater contributing to runoff is due to shallow groundwater with a distinct seasonal variation.

There is no doubt that the basin characteristics – including soil type and depth, geology, forest cover, size of agricultural area, slope, and regional climate and weather affect the results of different catchments. Soil depth and type seem to play an important role. the agricultural area influences the amount of direct runoff contributed to the stream.

4.4 Discussion of the results

The different runoff distribution characteristics of both catchments reflect the influence of terracing and agricultural use on the hydrologic processes. After the large terracing direct runoff increased by 15 % at Loechernbach, direct runoff in summer at 50 % of the total runoff, and rapid base flow in winter at 48 %, are the main contributors to streamflow. In contrast to this, the long-term baseflow in the small-terraced Rippach at 46 % of the total runoff plays a major role throughout the year, the storage constant is therefore 3 times larger than that of Loechernbach.

The short-term baseflow is the component with the highest variability. In addition to evapotranspiration, the spatial distribution of soil types and their depths might be the reason for this. The high variability of the absolute values of the long-term baseflow in the Kaiserstuhl implies that the saturated area – the shallow groundwater zone – is easily influenced by evapotranspiration under the present soil conditions. In fact, a 2-3 m deep loess profile could store up to 700 mm which is basically the mean annual precipitation of the area, as has been shown in previous studies by MORGENSCHWEIS (1984).

Strong recession in dry periods, particularly from August to October, seems highly probable in the large-terraced Loechernbach catchment. Terracing and the corresponding enlarged drainage network reduce the groundwater level, but strong evaporation and the process of ascending capillary water result in a reduction of the water inflow from underground in the high evaporation or low soil moisture phase. Moreover, the water intake in later summer months and the extraction of water by vegetation increases the recession. Clearly, this strong recession phase and the subsequent rebound phase cannot be explained by balanced precipitation. This largely increases the difficulty of determining *CG2* at Loechernbach, and the explanation of the vegetation theory in DIFGA does not shed much light on the matter.

The storage constants obtained by DIFGA are inaccurate when considering them as residence times. Particularly, using DIFGA is impossible to estimate the response times of fast components. An accurate evaluation of these parameters should be provided by the tracer method.

4.5 Summary

The results of applying the DIFGA model to the small-terraced Rippach and the large-terraced Loechernbach catchments are summarized as below:

The results of runoff formation calculations verify the fact that effective precipitation is small compared to total precipitation at both study sites, and the reason for the small effective precipitation is the extremely well-established pore structure of loess. The runoff concentration calculations show the dominant contributions of long-term baseflow (*QG2*) in the Rippach catchment, and of short-term baseflow (*QG1*) in the Loechernbach catchment. These results are consistent with previous studies conducted at these sites. The storage constant for the long-term baseflow at the Rippach is 3 times higher than that for the Loechernbach (420 days vs. 140 days). The storage constant for the rapid baseflow is 10 days at Rippach and 8 days at the Loechernbach. Compared to other basins (e.g. Lange Bramke, Wernersbach, and Rietholzbach), the present study basins have smaller runoff coefficients and large seasonal variations of both baseflow components.

5 Long-term investigations

In this chapter the long-term results, namely the seasonal investigations with the help of isotope and hydrochemical tracers, are presented. These long-term investigations provide the basis for comparing the results to those obtained by the DIFGA model. In the meantime, some mathematical flow models are applied to the isotopic data to determine the mean residence time of groundwater.

5.1 Isotope investigations

5.1.1 Isotope characteristics

Isotopic concentration in precipitation

The measured values of $\delta^{18}O$ for the precipitation and streamwater are presented in Figure 5.1. The weekly average $\delta^{18}O$ values of precipitation, which have been weighted, are generally lighter in winter and heavier in summer. The $\delta^{18}O$ contents in precipitation are characterized by seasonal variations on which short time fluctuations due to weather systems are superimposed. Distinct tracer input signals occur, in which deviations of the $\delta^{18}O$ concentrations from the seasonal average coincide with high precipitation amounts.

The total range of variation of $\delta^{18}O$ contents in precipitation from September 1994 to September 1995 is about 18.70 $^0/_{00}$; the lightest value was -16.96 $^0/_{00}$ during the first week of December 1994 and the heaviest was 1.77 $^0/_{00}$ during May 5th through 11th, 1995, mainly due to temperature variations. The most divergent values, usually enriched in the heavy isotopes, originated from small showers. They carry little weight in the monthly mean composition, because the amount of precipitation during these rainfall events is usually small - being lost together by plant interception or complete evaporation from the soil (GAT 1971).

There are about six small heavy-light periods of isotope content in precipitation within the investigated year. The mean monthly compositions of precipitation show little difference between the two research sites. Also, their average annual values are nearly identical (-8.076 $^0/_{00}$ / -8.085 $^0/_{00}$).

Isotopic concentration in streamwater

The isotopic content peaks in precipitation can be identified in the total runoff but are distinctly clipped. The range of fluctuation of precipitation is restricted to about 1.2 $^0/_{00}$ for the Rippach and 4.0 $^0/_{00}$ for the Loechernbach, demonstrating that the seasonal $\delta^{18}O$ variation of groundwater dominates that of streamwater (Fig. 5.1 a-b).

Evaporation in summer months plays an important role in these catchments with evapotranspiration sending 2/3 ~ 3/4 of the precipitation back to the atmosphere. In winter

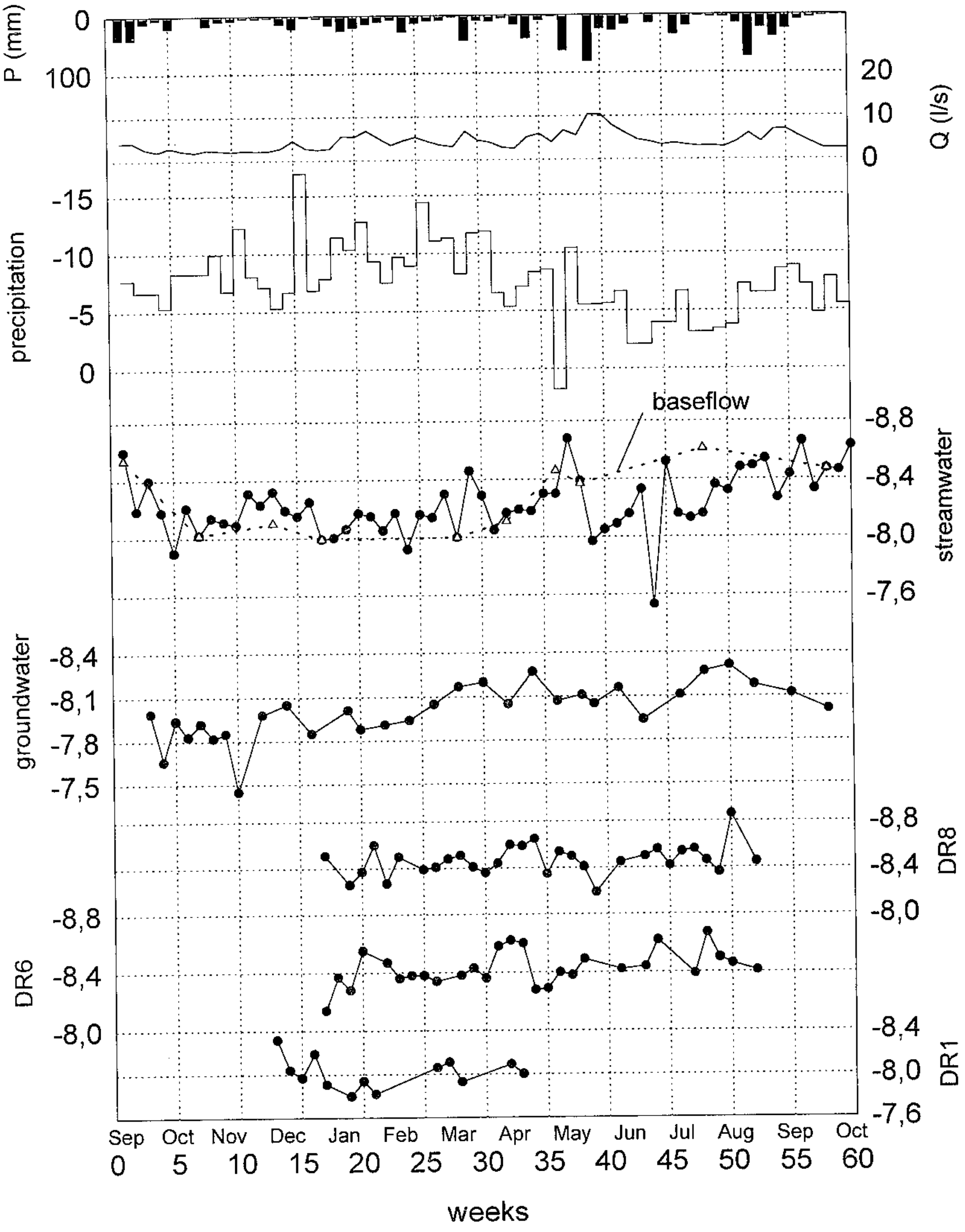


Fig. 5.1a The variations of oxygen-18 values of precipitation, streamwater and groundwater (DR1, DR6, DR8: drainage water in the Rippach basin)

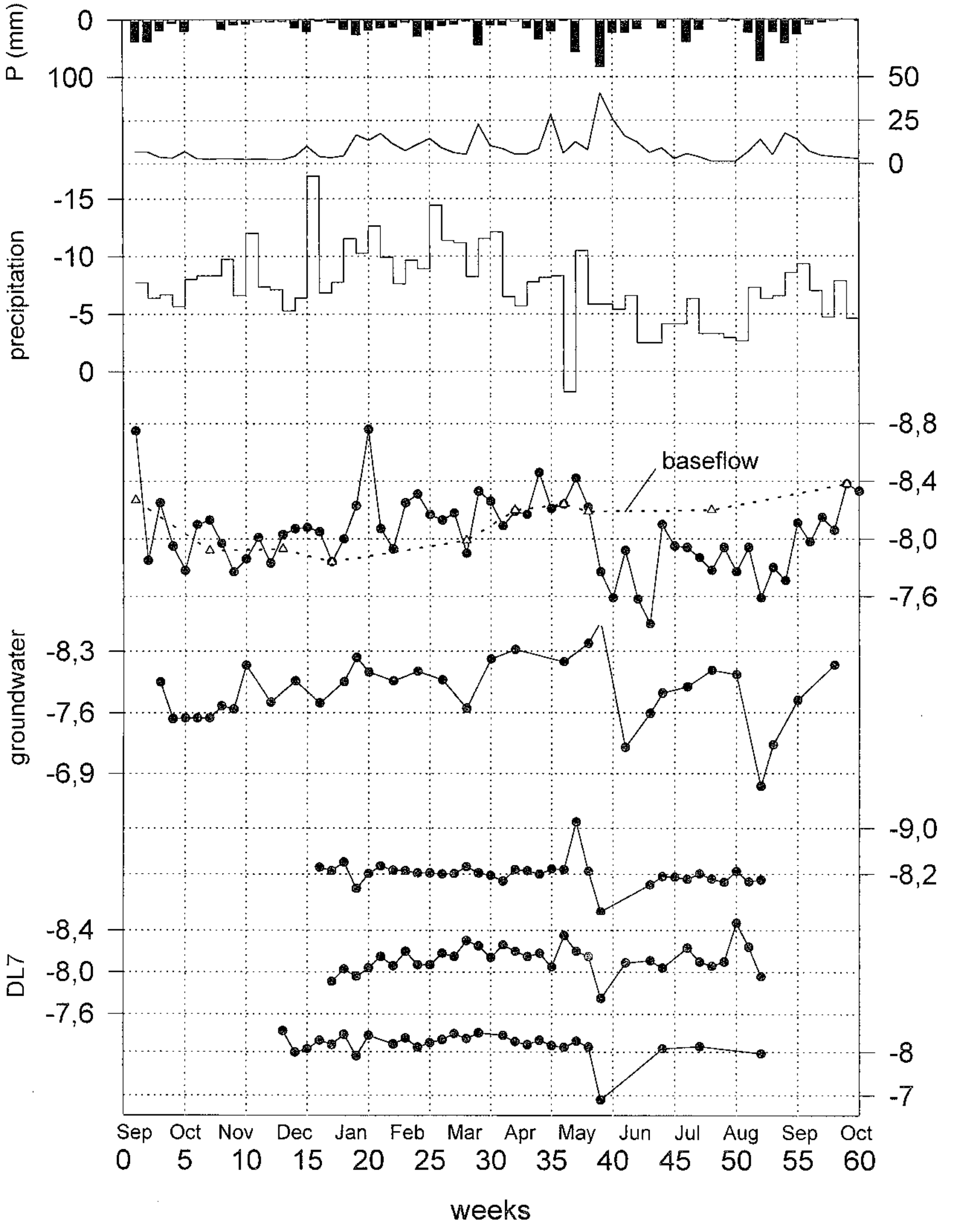


Fig. 5.1b **The variations of oxygen-18 values of precipitation, streamwater and groundwater** (DL2, DL4, DL7: drainage water in the Loechernbach basin)

a nearly homogeneous $\delta^{18}O$ profile can be assumed, because during winter evapo-transpiration is low. In the spring, with increasing biological activity and temperature, water stored in the soil is increasingly exhausted. In the surface layers evapo-transpiration fractionates the oxygen isotopes, resulting in an enrichment of $\delta^{18}O$. This effect may be enhanced by isotopically heavy rain in summer. The enrichment of $\delta^{18}O$ in summer is evident in the altered Loechernbach catchment because small amounts of groundwater from lower levels are supplied to this region. On the other hand, the abundance of deep groundwater in summer results in lighter $\delta^{18}O$ levels in the Rippach catchment.

Isotopic concentration in groundwater

Isotopic contents of groundwater at Rippach (GR) and at Loechernbach (GL) vary with the season. GR and GL are situated in the valley bottoms, with an mean annual isotopic content of -8.05 $^0/_{00}$ and -7.90 $^0/_{00}$, respectively. Their annual mean values are about 0.20$^0/_{00}$ heavier than those of stream water, due to an increased enrichment in $\delta^{18}O$ in near-stream areas during the summer months.

Isotopic concentration in baseflow

Baseflow integrates the isotopic signatures of near-stream groundwater. Hence, the isotopic contents of baseflow at Rippach (BR) and at Loechernbach (BL) are different from those of groundwater GR and GL. This difference is reflected in the relatively stable isotopic content of BR and BL in the winter months and the regular winter and summer variation within the year. In contrast to the baseflow values, the isotopic content of GR and GL vary considerably, depending on the temperature and the influence of infiltrated precipitation water, because GR and GL are only 0.8-1.6 m below the surface. In addition, the trend of the heavier isotope levels in the baseflow at Rippach during the summer months could be caused by the recharge of spring water upstream of the basin.

In general, the isotopic content of baseflow under-envelops the stream waters mean weekly values - with the heaviest enveloping occurring in winter and the lightest in summer. During the alternating period (the middle of April until the middle of May) the relationship between streamwater and baseflow is often irregular. The mean annual values of baseflow, BR and BL, are similar to those of precipitation, suggesting the objective choice of baseflow as old water in the study of hydrograph separation as noted in the Chapter 2.2.

Isotopic concentration in drainage water

Isotopic analysis was also performed on a series of drainage samples collected at Rippach (DR1, DR6, DR8) and Loechernbach (DL2, DL4, DL7). It is probable that the drainage water data presented in Fig. 5.1 are indicative of both the average oxygen-18 composition and its general variability over time.

The water samples taken near midstream or upstream (DR8, R6, DL4 and DL7) were generally isotopically more negative and less variable than those of the groundwater gauging stations near the outlets. They fluctuated within a narrow range during the observation period, indicating that they come from the deeper zone where less fractionation happens, and that their dominant source is the winter precipitation.

These samples responded to the change of the weekly isotopic composition of rainfall for about two weeks. An exemplary sample is in the 39th week, June 1st, when water from drainage gauges responded to the heavy oxygen-18 of rainfall with about a two week delay.

The long term variability of isotopic composition in the altered Loechernbach is larger than in the Rippach catchment. This fact is demonstrated by a rapid response caused by the dominant shallow groundwater or deep soil water component in this basin, as well as by contributions from a number of installed drainage channels.

5.1.2 Results of isotope hydrograph separation

As presented in Chapter 2.4, the separations in this study are performed using $\delta^{18}O$ in the baseflow as the isotopic signature of old water. For that the isotopic content of baseflow was determined using seven values from low-flow periods. Clearly, during the alternating season (mid-April to mid-May), the calculated old water fractions are often inexact. This is because the requirement that precipitation and groundwater have distinctly different isotopic content could not often be satisfied in this period.

The average new water proportions in summer and winter of the hydrological year, calculated from weighted weekly $\delta^{18}O$ concentration of precipitation and streamwater, are listed in Fig. 5.2. The seasonal average of new water contribution to the stream is 10 % in winter and 23 % and summer at Loechernbach, up to 12 % greater than that at Rippach. The isotopic composition of old water should represent an integrated value for local rainfall, which over the observed twelve month period has an overall weighted mean isotopic composition of -8.24/-8.03 $^{0}/_{00}$ at Rippach and Loechernbach, about 0.12/0.03 $^{0}/_{00}$ heavier than that in precipitation. The different values between average precipitation and old water at Rippach should relate to the spring water transported out of the basin during the winter months.

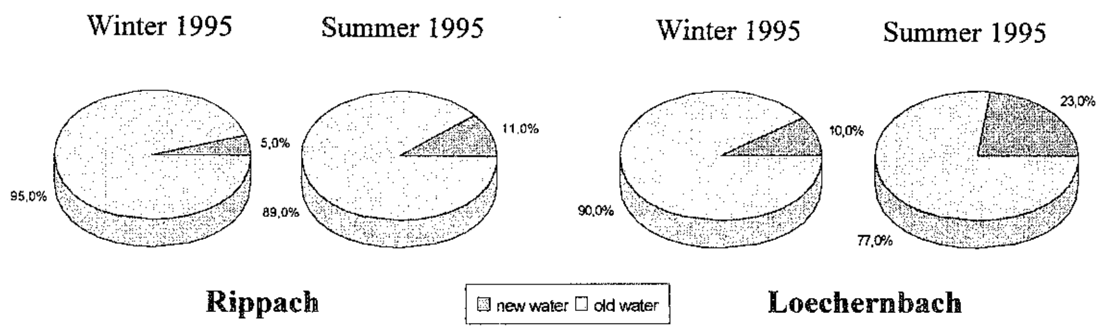


Fig. 5.2 Hydrograph separations using $\delta^{18}O$
Winter: November - April / Summer: May to October

5.2 Chemical investigations

5.2.1 The chemistry of different components

The temporal variations of runoff chemistry reflect the dynamics of runoff and solute sources. Based on data collected from September 1994 to October 1995, the mean concentrations of major ions in precipitation, streamwater, and groundwater are summarized in Table 5.1. Some results are given in CUI et al. (1995).

Precipitation

The ionic concentrations at rain collected during the total observation period are usually under 5 mg/l, and the range of spatial and temporal variation was relatively small (up to 3 mg/l). Compared to the high ionic concentrations in the river, those found in rain water are sufficiently small that they can be neglected in the study sites.

Streamwater

In the study basins streamwater chemistry varies strongly with the season, depending on antecedent hydrological conditions and the nature of individual rainfall events. The streamwaters are alkaline, with pH values ranging from 8.1 during low flows to 7.5 during periods of high discharge accompanying spring rains. Based on the weekly data collected from September to December 1994, runoff chemistry is dominated by calcium and bicarbonate, which account for 63 % / 80 % of total anion/cation concentrations at Rippach and 68 % / 78 % at Loechernbach. Variations of their concentrations during a runoff event are similar to those of EC. Because of grape growing and other agricultural activity, high concentrations of NO_3^- and PO_4^{3-} were observed. In comparison K^+ and Na^+ concentrations are very low and show little variation.

The general concentrations of the major cations rank as follows: calcium > magnesium > sodium > potassium; the concentrations of the major anions ranked as follows: bicarbonate > sulfate > chloride > nitrate.

In general, the weekly concentration of each solute derived from Loechernbach is higher than that from the nearly natural Rippach catchment, due to the increased impermeable area associated with the installation of a dense network of pipes and mole drains. In addition, larger variations in concentrations occur in the Loechernbach catchment than in the Rippach catchment. Two possible sources of the ions needed to produce such hydrological behavior are a steady input into the drainage system from the catchment caused by diffuse infiltration of precipitation, and activation and wave-like input into the drainage system caused by surface runoff (RIEG et al. 1991).

Groundwater

The chemistry of the groundwater reflects the atmospheric inputs and particularly the subsequent reactions. In both catchments groundwater is less alkaline than streamwater and has higher HCO_3^-, Ca^{2+}, PO_4^{3-} and K^+ concentrations than the corresponding streamwater. The pH in groundwater is similar in both catchments (average 7.05). The groundwater of the Loechernbach catchment, unlike that of the Rippach, has low NO_3^- but high PO_4^{3-} concentrations.

The chemistry of the groundwater remained quite stable through the winter of 1995, despite comparatively large changes in stream chemistry. It was somewhat surprising that there was a different chemical composition between the two groundwater measuring sites GR (at Rippach) and GL (at Loechernbach); chemical transport was 38 % lower at Loechernbach than at Rippach. Therefore, it is concluded that the water at GL, which is situated near the outlet of the catchment, is uncontaminated groundwater from another origin, i.e., it comes from another side of the slope out of the catchment and does not contribute to runoff in the output.

Tab. 5.1 Mean electrical conductivity and ionic concentrations of different types of water (September 1994 - October 1995)

Catchment	Sampling site	EC (µs)	Cl^- (mg/l)	SO_4^{2-} (mg/l)	NO_3^- (mg/l)	HCO_3^- (mg/l)	PO_4^{3-} (mg/l)
Rippach	PR	21	2.3	5.8	4.3	2	
	QR	855	34.1	90.2	42.8	368	0.10
	GR	1149	42.5	146.7	34.4	480	0.17
	BR	853	36.6	93.6			
	DR1	1066	44.7	121.5	79.8	365	
	DR6	992	36.7	121.1	63.6	366	
	DR7	909	30.1	96.3	31.8	444	
	DR8	883	33.0	85.4	36.8	412	0.19
Loechernbach	PL	21	2.0	5.7	2.4	5	
	QL	892	36.9	99.7	33.9	396	0.10
	GL	825	23.2	38.5	5.9	475	0.42
	BL	910	39.4	96.0			
	DL2	851	23.5	34.3	39.0	426	
	DL4	1020	35.3	73.1	34.5	498	
	DL7	1047	39.2	103.4	32.2	425	0.15
	DL13	1013	39.9	145.2	20.7	457	0.51

Drainage water

Most of the drainage water sampled is groundwater, particularly during low flow periods, but during periods of or high soil moisture, drainage water can be a mixture of soilwater, groundwater, and even precipitation. An example is DL13 (Loechernbach) where the water usually is groundwater from upstream flows. After rainfall events, direct runoff from asphalt roads also enters the drainage system. The corresponding processes include atmospheric deposition, biological uptake of nutrients, decay of organic matter, cation exchange, sorption-desorption, mineral weathering, and secondary mineral formation.

Drainage water samples DR6, DL4, and DL7 showed few temporal variations of solutes during the observation period, indicating that they consist largely of shallow groundwater with long residence times. DL2 and DR8 show lower concentrations of solutes than stream water does; DR8 connects with spring water from the hill slope, and the origin of DL2 is unknown. In contrast, DL13 and DR1 have higher concentrations of solutes than does streamwater. High concentrations of solutes at DR1 are most likely the result of agricultural activity. The high concentrations of solutes at DL13 may be due to a different origin of upstream groundwater. DR7 underwent a short period of rapid decrease in the concentrations of solutes (such as sulfate and EC) in May. This is because the discharge from tapped springs increased significantly, as noted in the Chapter 3.3.

The dominant anion in drainage water is HCO_3^-, which comes from the soil matrix through dissolution of rock. HCO_3^- proportion in anion is higher than in streamwater, especially in the Loechernbach. The higher NO_3^- concentration at Rippach (such as that in DR1) is due to that region's relatively heavy agricultural activity, particularly vegetable farming.

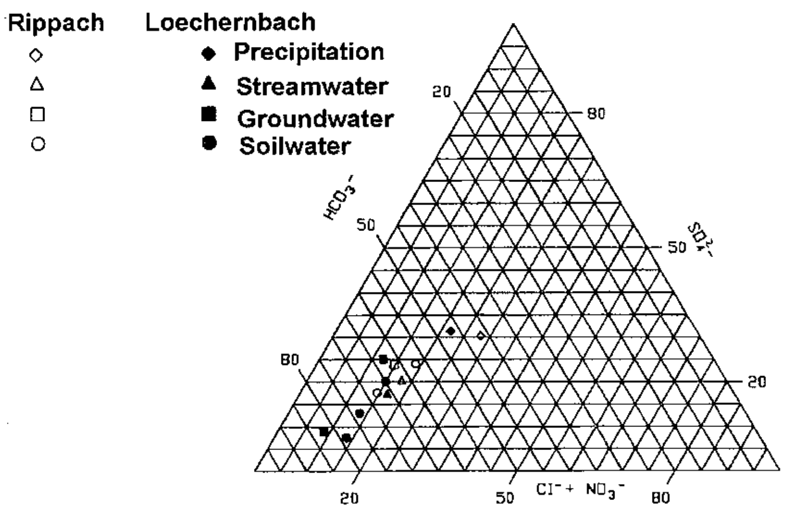


Fig. 5.3 The triangular diagram of anion during the winter of 1995

Chemical comparison of rainwater, streamwater, and groundwater

The various anion components are shown in a triangular diagram (Fig. 5.3). For comparison, concentrations for precipitation are also included. In the triangle the following characteristics are shown: the Rippach has higher levels of SO_4^{2-}, Cl^-, and NO_3^- anions than the Loechernbach; in other words, ratio of HCO_3^- concentration to the total anion concentration is smaller in the Rippach than in the Loechernbach. The variation of HCO_3^- levels shows a high correlation with the variation of the corresponding discharge in streamwater and groundwater.

The deficits between anion and cation concentrations in the mean weekly streamwater and groundwater are less than 3 %, but the deficits in the current streamwater are roughly 8 %. These deficits cannot be explained by experimental error alone. A probable origin of the error is the relatively high concentration of organical carbon. In other words, the error is caused by a high anion concentration (FOERSTER 1992).

5.2.2 The chemical characteristics of some solutes

Electrical conductivity values and concentrations of chloride, nitrate, and sulfate are analyzed in runoff and groundwater during the observation period, from September 1994 to October 1995. The aim is to discover whether these chemical parameters can be used as alternative tracers to support isotope results and whether they can give insight into routing aspects of runoff sources.

Chloride chemistry

There are some distinct features of Cl^- streamwater chemistry which need explanation in the context of separation. These features include:

1. The observed variation of Cl^- concentration in rainfall is small (1.2-3.1 mg/l), while in streamwater the range is comparatively large (27-39 mg/l at Rippach and 29-42 mg/l at Loechernbach, Fig. 5.4). This is because the Cl^- concentration in groundwater is ten times larger than that in precipitation. The observed variation of Cl^- levels in groundwater was 34-39 mg/l at BR (baseflow at Rippach) and 36-42 mg/l at BL (baseflow at Loechernbach). The mixing processes in the catchments are reflected in the streamwater variations. The lowest Cl^- concentrations in the stream occur from May 24th to June 1st, 1995, when the largest rainstorm occurred delivering 82 mm of rain, with a Cl^- concentration of 2 mg/l.

2. Systematic behavior in the seasonal patterns of the Cl^- concentrations in the stream is observed. Cl^- concentrations are higher in winter than during summer baseflow periods. Furthermore, higher Cl^- concentrations are observed during vegetation's growth period in the Loechernbach catchment, indicating that Cl^- during this time is not always conservative. This is probably due to the use of fertilizer in spring.

3. Groundwater concentrations of Cl^- vary little in winter. These levels decrease in summer, particularly in the altered Loechernbach catchment. These decreases can be attributed

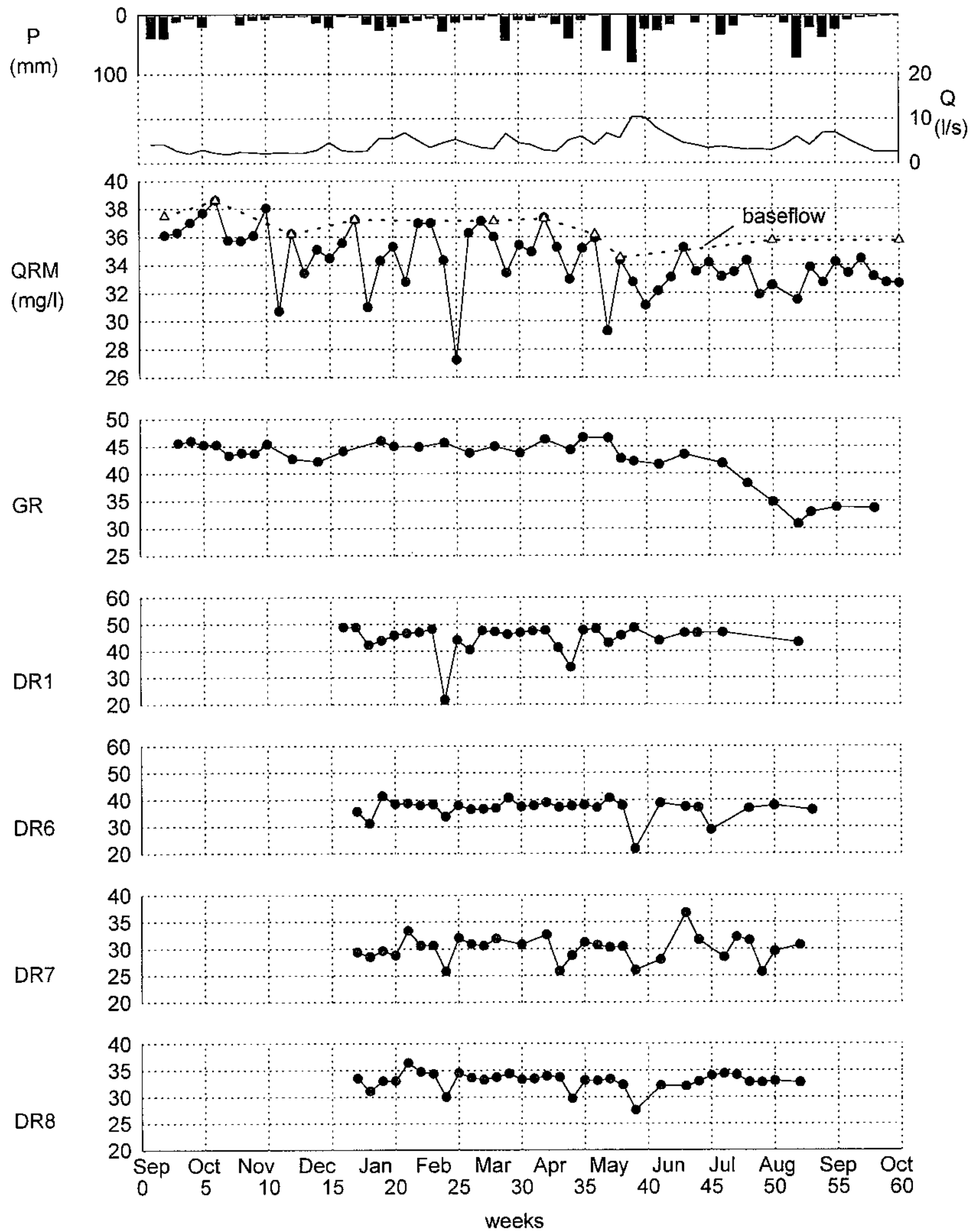


Fig. 5.4a Chloride concentrations for the hydrograph separation (Rippach) (QRM: average weekly values of streamwater; DR1, DR6, DR7, DR8: drainage water)

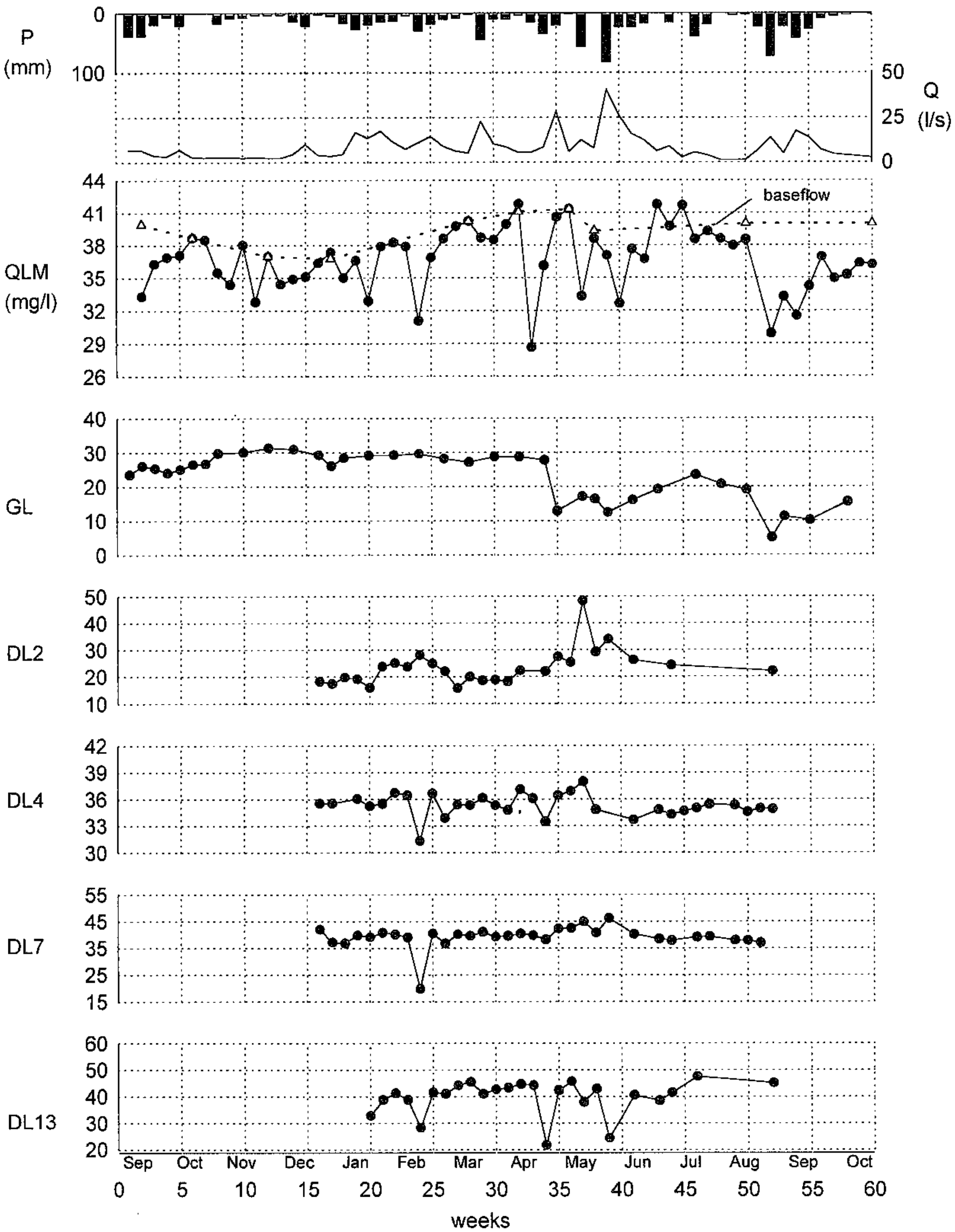


Fig. 5.4b Chloride concentrations for the hydrograph separation (Loechernbach)
(QLM: average weekly values of streamwater; DL2, DL4, DL7, DL13: drainage water)

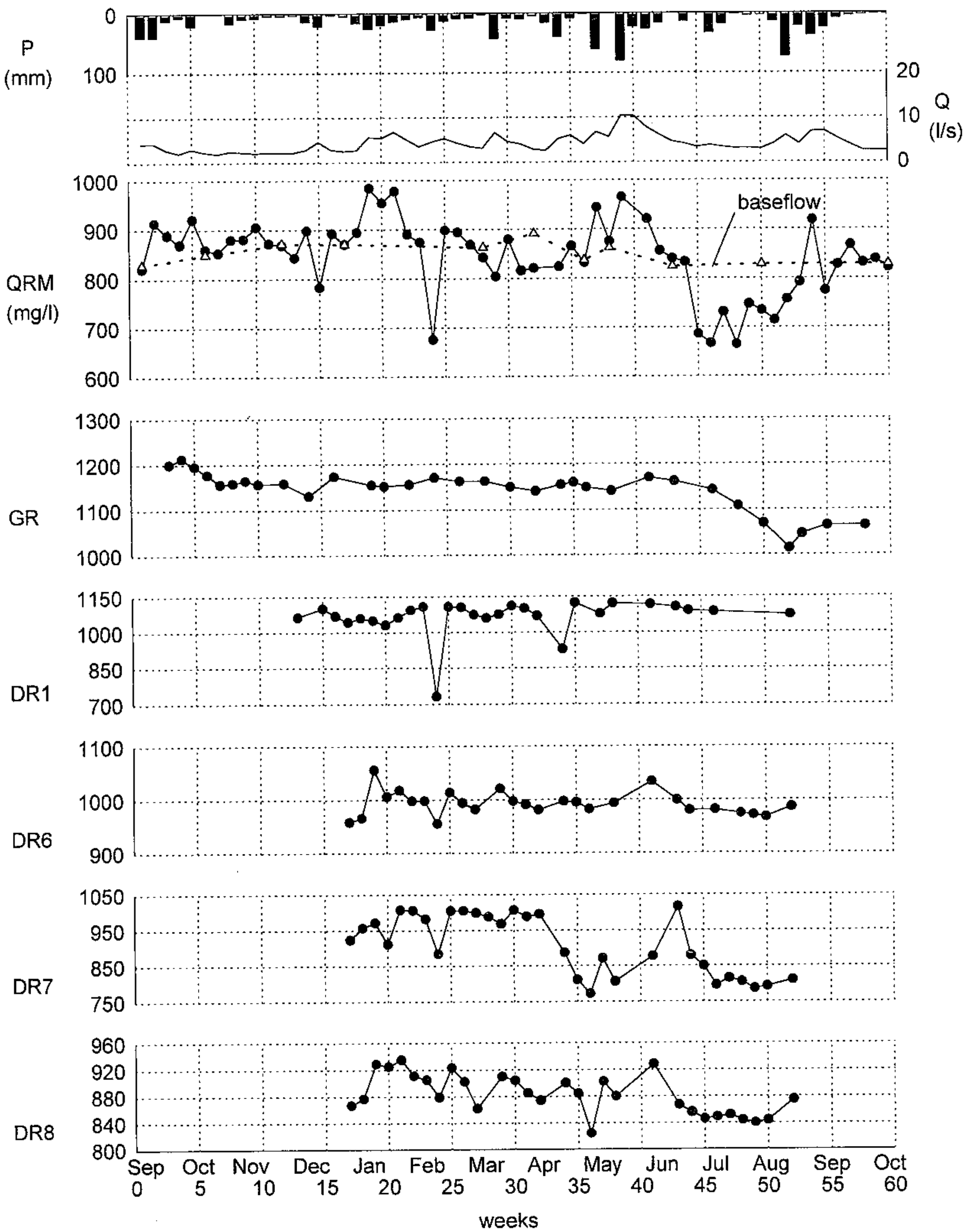


Fig. 5.5a Electrical conductivity values for the hydrograph separation (Rippach)
(QRM: average weekly values of streamwater; DR1, DR6, DR7, DR8: drainage water)

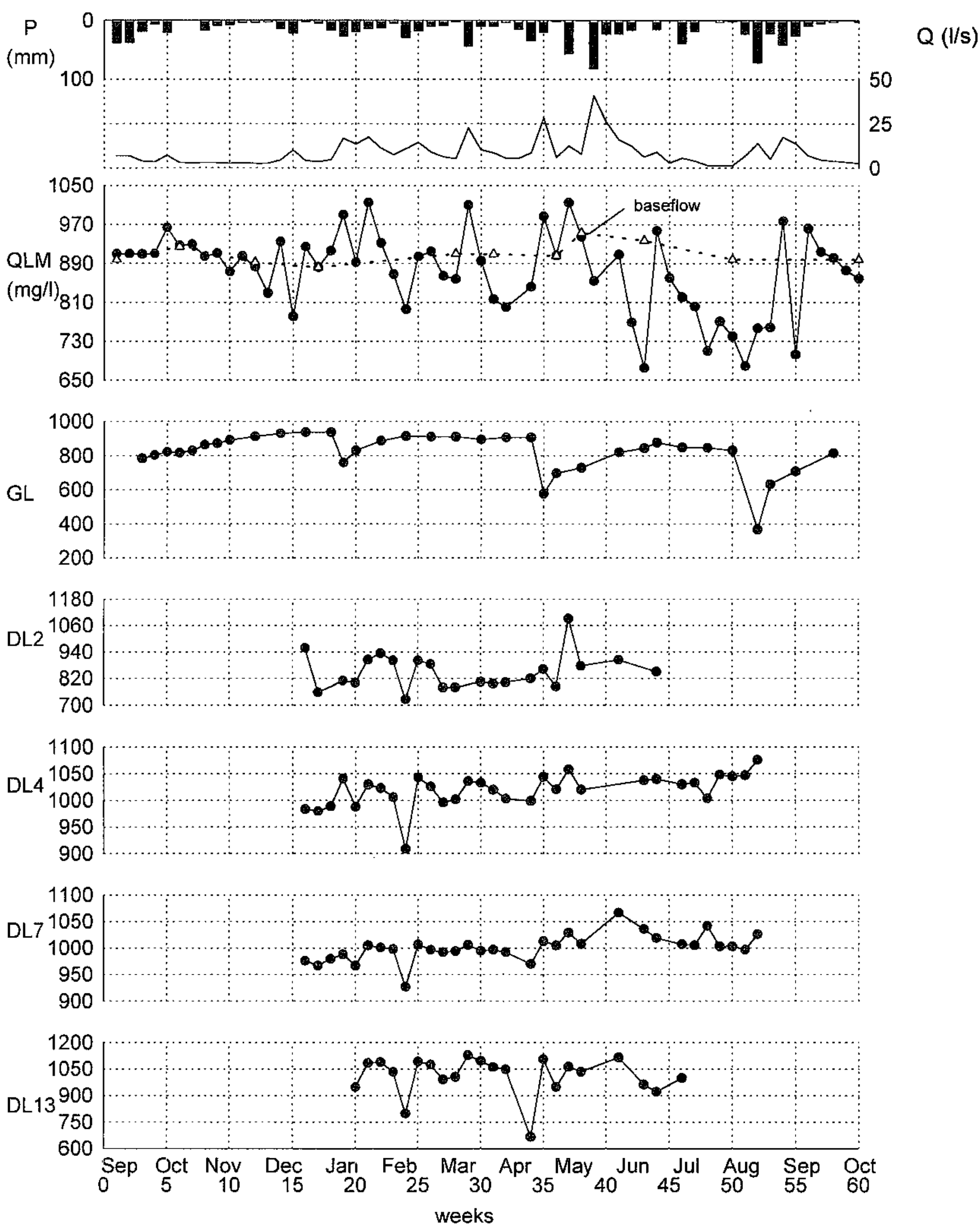


Fig. 5.5b Electrical conductivity values for the hydrograph separation (Loechernbach) (QLM: average weekly values of streamwater; DL2, DL4, DL7, DL13: drainage water)

to the entrance of fresh rainfall into groundwater, particularly during a series of rainstorms after April.

4. The Cl^- concentration for the Loechernbach streamwater is systematically higher than those for the Rippach, attributed mainly to the higher amounts of subsurface water, denoting the significant influence of drainage subpipes beneath the surface.

5. Small temporal variation of Cl^- concentration in most drainage waters suggests that these waters are normally groundwater. Their non-identical weekly concentrations reflect the different origins of the water, where drainage water DR1 at Rippach has the highest Cl^- concentration (44.7 mg/l), and DL2 the lowest (23.5 mg/l).

In general, the increase of Cl^- concentrations is partly due to agricultural activity. However, determining the quantities is difficult. The ratio between Na^+ and Cl^- concentrations in groundwater of near-shore regions has been used to prove whether Cl^- concentrations increase due to anthropogenic addition of Cl^- (SCHULZ 1973). Due to the change of solution content of precipitation with increased distances from shore, Na^+ and Cl^- levels decrease and their ratio will be changed in off-shore regions, so that in the Kaiserstuhl region this anthropologic influence cannot be proved. However, the present results show smaller seasonal variation of Cl^- concentration, compared with those of sulfate and nitrate.

EC chemistry

Both the Rippach and Loechernbach catchments show the same general pattern of EC variation with time and discharge. Weekly EC values ranged from 714 to 965 μs in the Rippach catchment, and from 675 to 1015 μs in the Loechernbach catchment. Generally, EC values were higher in the Loechernbach than in the Rippach catchment.

Several studies indicate that the relationships affecting the EC of natural water are complicated (PILGRIM 1979). Those complicated characteristics are reflected in the present investigation. For example, an inverse relation between EC and discharge is normally considered. However, in field observations EC values in stream water frequently increase with discharge during the first phase of a rainfall event. With the further increase of Q, the EC begins to decrease. This is in contrast to the inverse relationship of Q to Cl^- concentration, indicating that EC is not conservative.

An important finding in the field observations is a significantly higher EC value in surface flow than in precipitation. The EC values of overland flow has been investigated at five sampling points during a storm event in the Loechernbach catchment (Fig. 3.2). Unlike the EC value of 30 μs in precipitation, their average EC values on measured field roads (Tab. 5.2) are 94 μs. This means that the EC of new water should be approximately 100 μs instead of 30 μs, if new water is only from the asphalt roads. This is because water may also be enriched in salts during overland flow. Here overland flow is not inclusively that from the saturated natural areas. In those areas a higher EC is expected. This will be dealt with in Chapter 6.1.3.

Tab. 5.2 Observed EC values for overland flow in μs (M: measured point)

	M-1	M-2	M-3	M-4	M-5	M-mean
EC (μs)	115	74	37	94	148	94

5.2.3 Choice of chemical parameters as tracers

Numerous naturally occurring chemical tracers are used in order to separate hydrographs. Commonly used chemical tracers include electrical conductivity, chloride, silica, potassium, and sulfate. The issue of which chemical parameters are conservative, depends upon the soil characteristics of each catchment and upon the biological activity of the stream.

By comparing the temporal variations of major ions and PO_4^{3-}, EC, and silica of the different components in the first period of the investigation, from September to December 1994, it is found that major cations and HCO_3^- are not suitable as tracers for long-term runoff separation, because of their erratic variation or participation in oxidation reactions. The variation of PO_4^{3-} depends on the season and is difficult to determine.

Although silica has the advantage of being consistently absent from rainwater, the results of silica analysis were not satisfactory, after analyzing 12 weekly samples from October to December 1994. The average concentrations of silica were 12.5 vs. 12.8 mg/l for streamwater (QR) vs. groundwater (GR) at Rippach, and 9.4 vs. 9.3 mg/l for streamwater (QL) vs. groundwater (GL) at Loechernbach. Clearly, the difference between streamwater and groundwater is not distinct (about 2 %). Moreover, once December begins, the silica concentrations of groundwater at both measurement points were lower than those of streamwater. One probable reason for this could be the small sizes of the forest areas relative to the sizes of the basins.

Similarly, EC and SO_4^{2-} are not concluded to be conservative tracers after comparing their graphs over eight baseflow measurements, in which significant seasonal variations of baseflow are observed, especially in the altered Loechernbach catchment. The strong increase of SO_4^{2-} content from winter to early summer could arise from grape growing and the introduction of fertilizer with high levels of sulphurous nitrogen such as sulphuric ammonia.

In contrast to these parameters, Cl^- showed relatively conservative characteristics, in particular during the winter months in the Rippach catchment. Its slight fluctuations in the summer months are due to evaporation effects and increased supplies of spring water with relative low Cl^- concentrations (Fig. 5.4).

Finally, Cl^- is suggested as the hydrochemical tracer to be used for the long-term separation. In fact, Cl^- has the simplest chemical behavior compared with other anionic candidates. As for the possible influence of fertilizer due to agricultural use of chloride, there has been no significant evidence of this in groundwater, as indicated in previous investigations by WEGEHENKEL (1985) and LANGHAMMER (1988). Moreover, this influence does not bring significant seasonal variation in runoff, as proved by the stable Cl^- in baseflow throughout the observation year (Fig. 5.4). These quasi-conservative characteristics of Cl^- distinguish it from SO_4^{2-} and NO_3^-. The choice of Cl^- as the tracer is supported by the comparison of the Cl^- results with isotope results, and the conservative characteristics of Cl^- will be dealt with in Chapter 7.2.1.

5.2.4 Results of chemical hydrograph separation

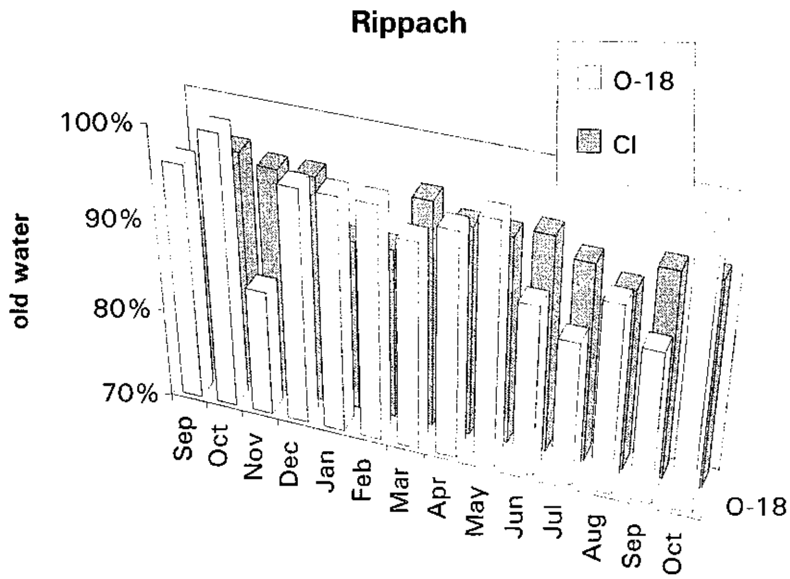
The average old water proportions obtained by Cl^- are presented in Table 5.3. The annual averages between the two sites are identical. High old water estimated at Loechernbach associates with higher interflow contribution, which is discussed in Chapter 7.

Tab. 5.3 Comparison of the contribution of old water to stream by different methods from September 1994 to September 1995

	Winter		Summer		Year	
	Rippach	Loechernbach	Rippach	Loechernbach	Rippach	Loechernbach
Chloride	91%	92%	92%	89%	92%	91%
Oxygen-18	95%	89%	89%	77%	92%	83%

The prerequisites for Equations 2.14-2.16 (Chapter 2) are less fulfilled than they are for $\delta^{18}O$. EC is an example. Water will get continuously enriched in salts on its way through the unsaturated- and saturated zones, which causes considerable variations in the salt concentration of groundwater within the basin.

The monthly average old water fractions obtained by ^{18}O and Cl^- are summarized in Figure 5.6. Old water is the dominant factor throughout the year, in particular during the low flow period. Significant contribution of new water appeared only for the main peak period such as the beginning of June 1995 and mid-September 1995. It is noted that hydrograph separation for long-term observations is very different from that for short-term storm events, old water and new water of the former is built in an arithmetic average, while the latter relates to pre-event and event water with clear physical explanation.



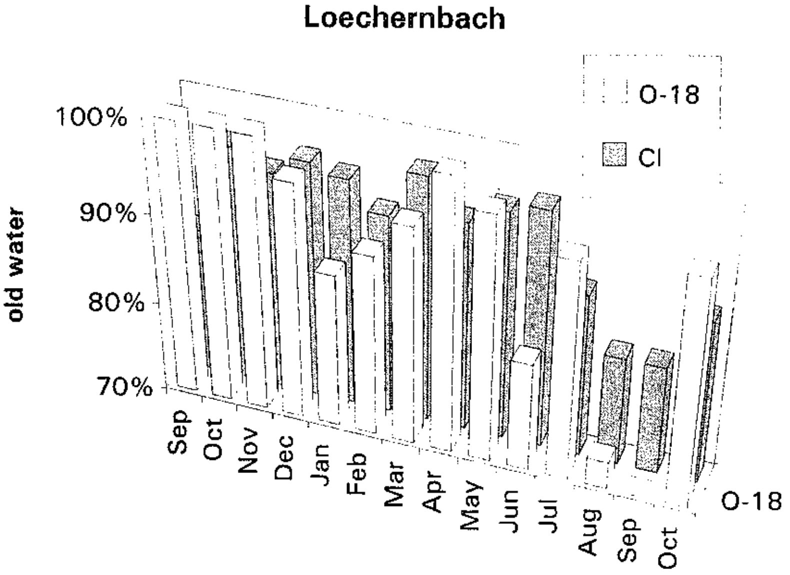


Fig. 5.6 Monthly variation of old water in the stream calculated by oxygen-18 and chloride during the observation period from September 1994 to October 1995

5.3 Sensitivity analysis

The accuracy of Equations 2.14-2.16 (Chapter 2) is often discussed in the previous studies. ROHDE and some other authors point out that the uncertainty in the calculations using Equation 2.14 mainly depends upon the accuracy in the $\delta^{18}O$ of rainwater and groundwater (old water). It has, however, been regarded as impossible to completely quantify this error (CHRISTOPHERSON et al. 1984, RODHE 1987).

Indeed, comparison of old water proportions from Equation 2.14 using the variances of C_g and C_d must be made in the light of the sensitivity to errors in the $\delta^{18}O$ values used. Sensitivity in the Equations 2.14-2.16 arises from systematic and random causes such as sampling and analytical error. Uncertainty in Q_g/Q_t attributable to analytical error (ΔX) was estimated following RODHE (1987):

$$|\Delta X| = \frac{1}{C_g - C_d}\sqrt{\Delta C_t^2 + \Delta C_d^2 + X(\Delta C_g^2 + \Delta C_d^2)} \tag{5.1}$$

where ΔC_t, ΔC_g and ΔC_d are errors in the $\delta^{18}O$ of streamflow, new and old waters, which included two errors: analytical error and representative (sampling) error. Thus $|\Delta X|$ is inversely proportional to C_g-C_d for a certain value of X. But the relative uncertainty $\Delta X/X$ decreases as X increases.

A comparison of average ΔX for summer and winter is performed in the present study, in a different way from the previous studies by RODHE. An average uncertainty of ΔC_t, ΔC_g and ΔC_d for a half-year scale is assumed to be 0.05 $^0/_{00}$, ignoring systematical error. This assumption is acceptable, because the sampling and analytical errors are independent of each other and thus, long-term average error is relatively small. This is supported by the fact that the annual average isotope content of precipitation at Rippach (-8.085 $^0/_{00}$) differs only 0.009 $^0/_{00}$ from that at Loechernbach (-8.076 $^0/_{00}$).

Tab. 5.4 shows a relative uncertainty for different seasons. The relative uncertainty of both catchments for winter half-year is identical (5 %), but this uncertainty is somewhat larger (6-8 %) for summer months in the Loechernbach catchment. This is because old water fraction X to this basin is small, so that the relative uncertainty is larger. Generally, the uncertainty at Loechernbach is larger than that at Rippach, and the summer uncertainty is larger than winter one. Moreover, new water (precipitation) results in larger sensitivity compared with old water. This is the result of a comparison for a series of probable ΔC_g and ΔC_d values. The computation suggests that uncertainty of old water proportion evaluation can arise up to 45 % for summer, when the analytical and representative errors of $\delta^{18}O$ in precipitation are 0.5 $^0/_{00}$.

Tab. 5.4 Estimate of relative uncertainty of old water using oxygen-18
Ct/Cg/Cd: weighted average isotopic content of streamwater / old water / new water, $\Delta Ct/\Delta Cg/\Delta Cd$: uncertainty of streamwater / old water / new water, $\Delta X/X$: relative estimate uncertainty of old water fraction

		C_t	C_g	C_d	ΔC_t	ΔC_g	ΔC_d	$\Delta X/X$
Rippach	winter	-8.16	-8.05	-10.19	0.05	0.05	0.05	5%
	summer	-8.29	-8.49	-6.72	0.05	0.05	0.05	6%
Loechernbach	winter	-8.19	-7.97	-10.12	0.05	0.05	0.05	5%
	summer	-7.90	-8.25	-6.79	0.05	0.05	0.05	8%

5.4 Infiltration parameter α

The Infiltration parameter α is calculated based on isotopic data of precipitation and groundwater from September 1994 to September 1995 (Equation 2.21-22, Chapter 2). The infiltration coefficients are calculated to be 0.98 at Rippach and 0.68 at Loechernbach. The winter infiltration exceeds summer one by about one-third in the Loechernbach catchment. Such a difference is not visible in the small-terraced Rippach catchment.

Based on discharge and precipitation data (Equation, 2.23 Chapter 2), the infiltration coefficients are approximated to be 0.86 at Rippach and 0.69 at Loechernbach for the year of 1995. The average infiltration coefficients considering the hydrological data from 1989 to 1995 amount to 0.59 and 0.36 in the Rippach and Loechernbach catchment, respectively.

In spite of similar infiltration coefficients for isotope and hydrological data of 1995, the values based on the hydrological data from 1989 to 1995 may underestimate the infiltration of the summer. Commonly, α determined by isotope data represents the real distribution of the tracer mass. This probably includes an effect of altitude, in the infiltration water, i.e. the weighted tracer mass transfers to the groundwater during every month of the year (MALOSZEWSKI et al. 1992).

5.5 Determination of residence times by using mathematical flow models

5.5.1 Applications of the simpler procedure of the exponential and dispersion models

According to Equation 2.29-34 (Chapter 2), a rough estimation of the residence times with the simpler exponential model and dispersion model are given as below.

The average monthly peak-to-peak amplitude of precipitation (A_{in}) during the investigation time is about 3.4 $^0/_{00}$. The amplitudes of isotopic concentrations in streamwater are only about 5-10 % of that in rainfall, this amplitudes in groundwater are about 5-7 %. Table 5.5 shows the mean residence times for various outputs. Here streamwaters QR at Rippach and QL at Loechernbach are the water at the output of the basins, base flow BR at Rippach and BL at Loechernbach are considered to be old water, groundwaters deal with only GR at Rippach and GL at Loechernbach.

Tab. 5.5 Results of EM and DM models with Sinusoidal Input (T: years)

Type	Basin	Amplitude (A_{out})	T (EM)	T (DM) D/vx=0.01	T (DM) D/vx=0.02	T (DM) D/vx=0.05
Streamwater	Rippach	0.20 $^0/_{00}$	2.8	2.7	1.9	1.2
	Loechernbach	0.32 $^0/_{00}$	1.7	2.5	1.7	1.1
Base flow	Rippach	0.18 $^0/_{00}$	3.2	2.8	1.9	1.2
	Loechernbach	0.25 $^0/_{00}$	2.2	2.6	1.8	1.2
Groundwater	Rippach	0.25 $^0/_{00}$	2.2	2.6	1.8	1.2
	Loechernbach	0.50 $^0/_{00}$	1.1	2.2	1.6	1.0
Drainage waters	DR6	0.19 $^0/_{00}$	2.9	2.7	1.9	1.2
	DR8	0.10 $^0/_{00}$	5.8	3.0	2.1	1.4
	DL2	0.21 $^0/_{00}$	2.6	2.7	1.9	1.2
	DL4	0.14 $^0/_{00}$	4.0	2.9	2.0	1.3
	DL7	0.15 $^0/_{00}$	3.7	2.8	2.0	1.3

* Amplitude of rainfall (A_{in}) is 3.4$^0/_{00}$

Applying the simpler procedure of EM (Equation 2.31-2.32, Chapter 2), streamwaters display residence time (T) of 2.8 years in the Rippach and 1.7 years in the Loechernbach, about 0.4 year longer than those in base flow (3.2 and 2.2 years respectively). The residence times of near-stream valley groundwater were shorter than those in stream water (2.2 at Rippach and 1.1 years at Loechernbach).

Applying the simpler procedure of DM (Equation 2.33-2.34, Chapter 2), various residence times are computed, depending upon varied dispersive parameter. The dispersion model with $D/vx = 0.01$ and $D/vx = 0.02$ gives similar T-values as EM for the stream water at Rippach and Loechernbach, respectively. In the case of $D/vx = 0.05$, the residence times show little difference between various outputs.

The longest residence times appear in drainage water, especially in the case of using EM. This again implies that the drainage water originate from deep groundwater or spring water.

5.5.2 Applications of the exponential model, the exponential-piston model and the dispersion model

In order to use EM, EPM and DM models to simulate the varying output data, a long-term series of input data is necessary. Unfortunately, this condition is difficult to fulfill in the present study because the sampling of precipitation lasted only fourteen months. Also, isotope input data from other stations such as Stuttgart or Konstanz for the same period is not yet available at this time, so that an interpolation of data with the help of regression analysis between the present site and the other stations has been impossible. Furthermore, the fluctuations of the isotope values in streamwater and baseflow are usually smaller than 1.0 $^0/_{00}$ in the study sites, so direct interpolation of the other station data may be very sensitive.

For these reasons, the present simulation is based on the understanding that a regular annual variation of $\delta^{18}O$ in precipitation exists, and only seasonal variation of isotope concentrations is considered. In other words, hydrological year of 1995 represents the average case of recent years. This assumption may cause some errors in the estimated residence times.

Application of the EM model

Based on Equation 2.26 (Chapter 2), the EM is applied to simulate the output of stream water, baseflow and groundwater (Fig. 5.7). The EM has only one parameter T and thus, it has difficulty of simulating complex curves of output. Apart from the base flow fittings, the simulations for stream water and groundwater cannot give satisfying results, especially for the case of the reverse amplitude in stream water. The reason is most likely to small variations of the output and less flexibility of the EM in use.

Applications of the EPM Model

The EPM is the combined exponential-piston flow model and is more flexible than the EM model because of its additional parameter *s* which provides additional information about the aquifer.

The results derived from Equation 2.27 (Chapter 2) for streamwater and groundwater are given in Tab. 5.6. In the case of baseflow, *s* equaling 1 means the EPM model is the same as the EM model. There is a discrepancy between the calculated and the real value for June and July in both catchments. The calculated value for groundwater GR at Rippach is close to the empirically derived one. More variation can be discerned for groundwater GL at Loechernbach; In the case of streamwater, the calculated value for streamwater at Loechernbach (QL) is close to the empirically derived one, but such a fitting in streamwater QR at Rippach is imprecise because of insignificant seasonal variations in the $\delta^{18}O$ content.

Tab. 5.6 Residence times determined by EM, EPM and DM models in years

Type	Basin	*T* (EM)	*T* (EPM)	*s* (EPM)	*T* (DM)	*D/vx* (DM)
Streamwater	Rippach		2.33	1.20	1.60	0.25
	Loechernbach		2.29	1.35	1.46	0.09
Base flow	Rippach	1.75	1.75	1.00	1.71	0.04
	Loechernbach	1.75	1.75	1.00	1.75	0.05
Groundwater	Rippach		3.17	1.40	1.63	0.05
	Loechernbach		1.92	1.65	1.42	0.07

Applications of the DM Model

Application of the DM (Eq. 2.28, Chapter 2) causes good fittings for baseflow BR at Rippach and BL at Loechernbach. The calculated residence time of BR (1.7 years) can be considered as the basic scale in the research basins. The scale of other residence times can be approximated from this value.

The dispersion parameter *D/vx* is between 0.04 - 0.09, except for 0.25 at QR (streamwater at Rippach). The low value of the *D/vx* for most measuring points indicates that the hydrological system is hydraulically highly homogeneous and isotropic. It is most likely that diffusion in stagnant water plays an important role.

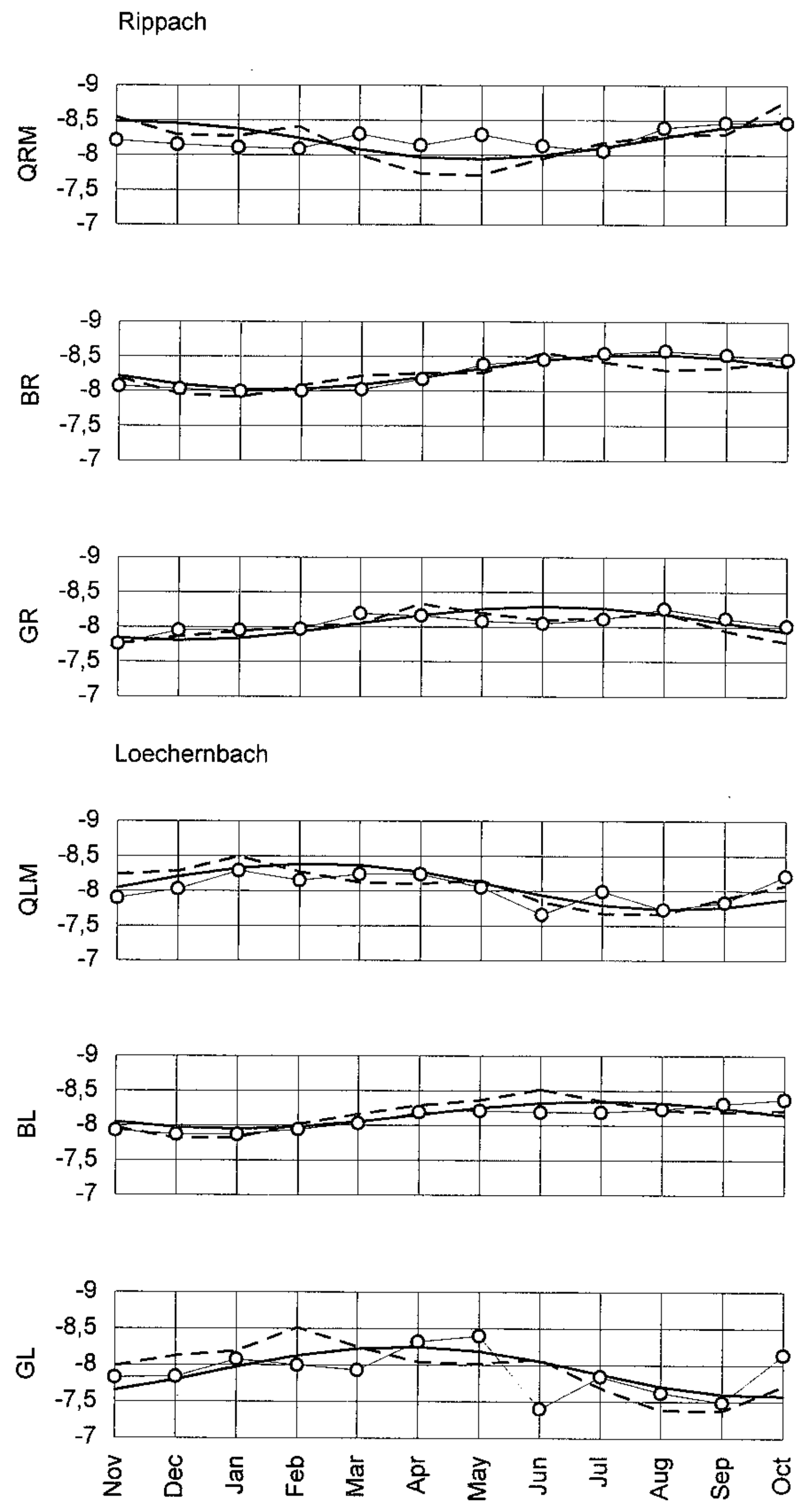


Fig. 5.7 **Theoretical output function found as a best fit to the $\delta^{18}O$ concentrations measured in stream water (QRM/QLM), baseflow (BR/BL) and groundwater (GR/GL)**

——————— simulated values using DM model
– – – – – – simulated values using EPM model
o—————o measured values

5.5.3 Comparison of different models

A variety of models have been used to interpret $\delta^{18}O$ variations in streamwater and groundwater. The best one is the DM model which produced results that correlated very well to the data. In fact, The DM is the most flexible model because this model allows for changing the parameter *D/vx* to suit the various cases. A comparison of the results shows that the simple procedure of the EM model can overestimate the residence times. The disadvantage of the simpler procedure of the EM and DM is that they are too rough and imprecise because a regular sinusoidal function does not exist.

The residence times obtained from the simpler procedure of the EM and the DM are longer than those obtained from the EM and the DM. The better fits of total runoff at the gauging station of QR/QL were found when using the dispersion model with T = 19.2/17.5 months (1.60/1.46 years) and D/vx = 0.25/0.09, which reflects average residence times of the study sites. There is a relative discrepancy between the experimental and the theoretical curves for streamwater at Rippach (QR), caused by small seasonal variation of the outputs, and the change of the water volume from winter to summer depended upon the drinking water supply. The fits of groundwater GR/GL were found using the dispersion model with T = 19.5/17.0 months (1.63/1.42 years) and D/vx = 0.05/0.065.

The best fitting results were obtained for the $\delta^{18}O$ contents in baseflows by using the DM model. The residence times are 20.5 and 21.0 months at Rippach and Loechernbach, respectively. The results from the theoretical curves show a very close correspondence to those from the experimental curves.

The annual average $\delta^{18}O$ contents of outputs are slightly different from that of inputs. This is because a part of water of the outputs was the precipitation stored before the year of 1995. In addition, fractionation may play a role. Evaporation of throughfall or interception loss can cause enrichment in oxygen. Interception loss is expected to be at its greatest in summer. In fact, all measuring points at Loechernbach reflect enrichment in oxygen. At Rippach, only groundwater GR point appeared somewhat enrichment in oxygen. However, this influence is expected to be limited in the present areas, because more humid environment in the research basin should lead to less enrichment (STEWART & MCDONNELL 1991). For the same reason, the enrichment at Rippach is expected to be smaller than at Loechernbach.

The mean residence times of old water 20.5 months at Rippach and 21 months Loechernbach, determined by the dispersion model (DM), are in very good agreement with the measured outputs. Considering 92 % / 83 % of old water proportion and negligible residence time for new water, the mean residence times of the whole catchments would be 18.9 months and 17.5 months, which is very close to the results obtained by the simulation using the dispersion model (DM) for streamwater QR (19.2 months) at Rippach and QL (17.5 months) at Loechernbach. The similar results obtained by different methods support each other.

5.6 Summary

The successive sampling for stream waters and the weekly sampling for groundwater and drainage water were performed during the hydrological year 1995 to investigate the annual variations of isotopic and hydrochemical tracers and to determine the old water proportion and corresponding flow parameters. These samples are believed to cover most flows that occurred in the hydrological year 1995.

The results of the investigation using $\delta^{18}O$ suggest that the stream water in the present study catchments be dominated by old water, i.e., 92 % old water in the Rippach catchment and 83 % old water in the Loechernbach catchment. The relative uncertainty of old water fractions are 5-8 %.

Higher concentrations of most solutes in the weekly sampling of the Loechernbach catchment have been observed compared with the Rippach catchment. Cl^- is chosen as hydrochemical tracer for long-term separation. The separation using Cl^- supports the results of dominant old water contributing to stream obtained from $\delta^{18}O$.

The sensitivity analysis showed that the uncertainty in the Loechernbach catchment is larger than that in the Rippach catchment, and that the uncertainty in summer is larger than that in winter. In addition, field observations indicates that the EC value of event water should be between 100-200 μs, rather than 15-35 μs of precipitation.

The mean transit times, determined by the dispersion model, were 1.60 years in the Rippach and 1.46 years in the Loechernbach. The best fittings were found to be by both baseflow. The infiltration coefficients between summer and winter are 0.98 at Rippach and 0.68 at Loechernbach respectively.

6 Short-term investigations

Short-term investigations include runoff investigations of two storm events and an experiment of using artificial tracers in the unsaturated zone, the latter is performed by UHLENBROOK (1995).

As opposed to long-term investigations, the assumptions of the two-component model for short-term investigations can easily be met. This is because tracer concentrations in stream water vary quickly and those of baseflow change minimally during individual rainfall events.

6.1 Experimental investigation of a short storm in the Loechernbach catchment, July 3rd, 1995

6.1.1 Hydrograph separation by oxygen-18 and hydrochemical tracers

A storm event on July 3, 1995 was investigated by using artificial tracer on a road along the Loechernbach. The isotopic and chemical concentrations ($\delta^{18}O$, Cl^-, EC, NO_3^-, SO_4^{2-}) of the samples collected within intervals of 1-10 minutes were simultaneously analyzed in the laboratory to assess the relative contribution of event water and pre-event water. The storms proved suitable for this purpose as the proportion of rainfall in $\delta^{18}O$ (average -4.2 6 $^0/_{00}$) was markedly heavier than that of baseflow (-8.08 $^0/_{00}$) immediately before the storm. The amount effect of rainfall during this event is not considered.

Fig. 6.1 shows stream discharge, $\delta^{18}O$, EC, Cl^-, NO_3^- and SO_4^{2-} in the Loechernbach catchment in response to this storm. With respect to the first event, there is a synchronous variation of $\delta^{18}O$ and EC. EC responses with double peaks of minimal values, such a phenomena is often observed during the storm events in the study basin. The values of the first peak show an initial rapid decrease of EC, caused by rainfall in the channel with very low value. The following increase of EC values is the result of available soluble salts which were exhausted while discharge increased (dilution effect). The subsequent minimal peak of EC probably relates to the delay of overland flow derived from asphalt roads or natural areas to the collecting stream. Another explanation of double peaks of the EC values could be the higher values of quick-return interflow. During the decline the EC values rose gradually because the increased contact time between water and suspended sediment allows the dissolution of the sediment to progress.

The Cl^- concentration initially decreased during rising stages in response to overland flow input and then increased as long-residence soil water enters the channel. At the beginning of the rising stage, runoff derived from roads treated with deicing salts causes a rise of Cl^- concentration; near peak discharge this effect ceases. SO_4^{2-} has a similar variation as Cl^-, indicating that SO_4^{2-} is also suitable as a tracer for the separation of this short-term event.

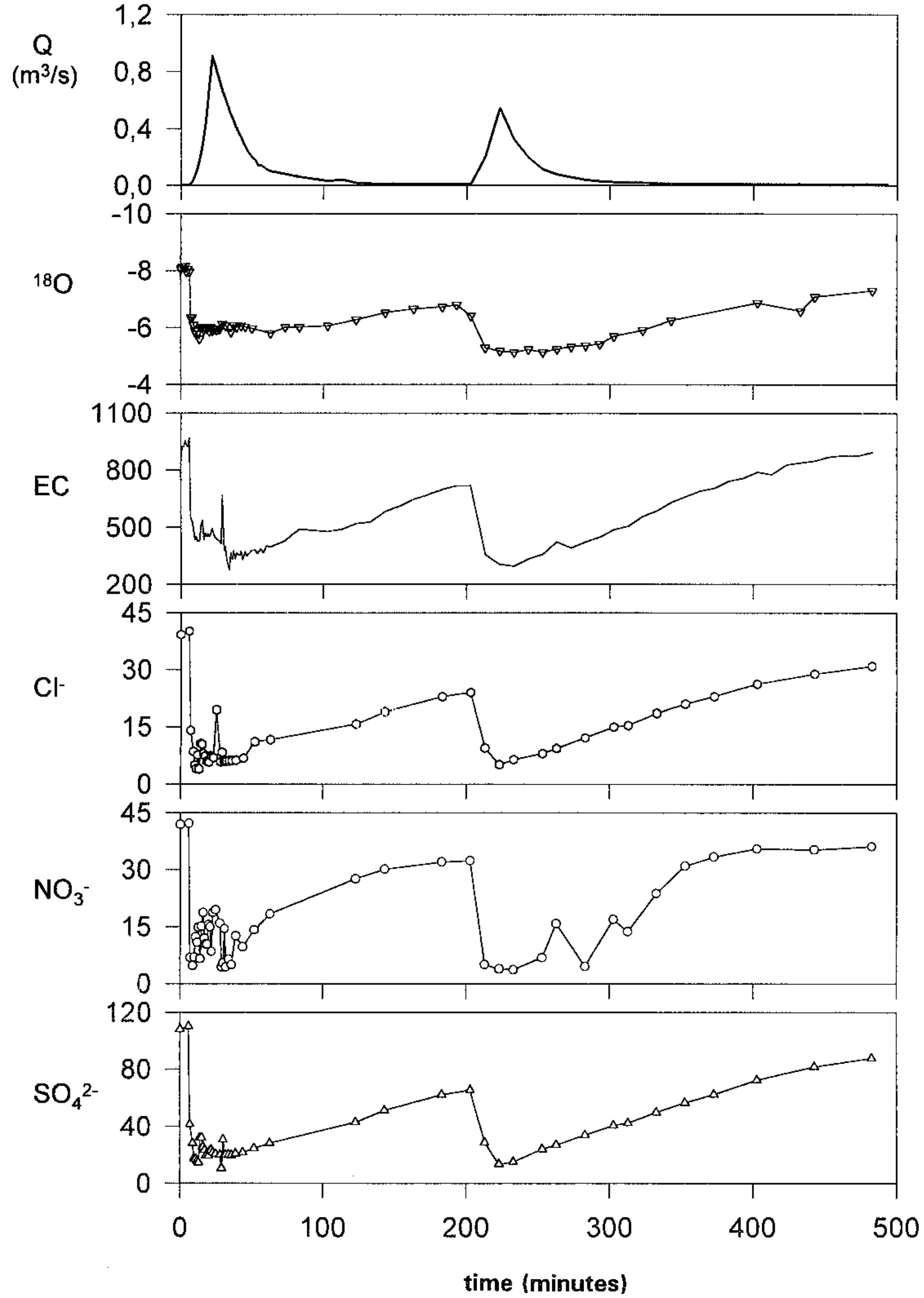


Fig. 6.1 The stream discharge, $\delta^{18}O$, EC, Cl^-, NO_3^- and SO_4^{2-} in response to the storm event of July 3rd, 1995 (Loechernbach)

The minimal concentrations of all tracers occurred before peak discharge, with the exception of NO_3^-, indicating significant contribution of groundwater to peak discharge, though this contribution is not dominant. The earlier minimal peak of NO_3^- most likely relates to the influence of fertilizer from agricultural areas.

With the exception of EC, higher event water fractions are observed than detected from $\delta^{18}O$ using chemical tracers,. The latter section of this chapter discusses these findings.

During the second storm event, the rainfall contribution to the flow is appreciable. The event water contribution rose up to 73 % using $\delta^{18}O$. Hydrochemical tracers reveal similar values (Tab. 6.1). The initial large fluctuation of solute concentrations disappeared, and believed to result from the depletion of dry and mineralized deposits during the first precipitation event.

Tab. 6.1 Event water contributions of two short-term storm events
(July 3rd, 1995, Loechernbach)

	$\delta^{18}O$	Cl^-	NO_3^-	SO_4^{2-}	EC
first event	55%	78%	68%	80%	57%
second event	73%	77%	82%	78%	61%
total	61%	78%	74%	79%	59%

In addition to discharge from asphalt roads, rapid interflow and overland flow from natural saturated areas are also an important contribution of water during storm events. This can be verified by the following calculation: the total discharge of 1660 m^3 during the storm event is equal to 0.98 mm on the basin area of 1.7 km^2 500 minutes after the start of the event. Assuming that the total 6 mm of rainfall is fairly well-distributed across the whole catchment, and precipitation (event water) averaged at 78 %, then about 13 % of event water or 0.76 mm rain water in depth reached the exit of the basin 500 minutes after the start. In total, asphalt roads cover 6 % of the total area. This means that the rest 7 % of event water should be derived from rapid subsurface flow in natural areas. In fact, rapid subsurface flow from natural areas in this event may be largely underestimated, when considering the following two factors: 1) event water is still dominant in stream water 500 minutes after the start of the event (Fig. 6.2); 2) the storm is accumulated to be from upstream areas, as observed in the field, so the average rainfall volume of 6 mm for the whole basin should be overestimated in reality. There is no doubt that macropore flow played an important role in this event, particularly in the latter phase of the events during and after the rainfall, macropore flow rapidly transported the water into the subpipes, these subpipes facilitate the waters rapidly entering the stream.

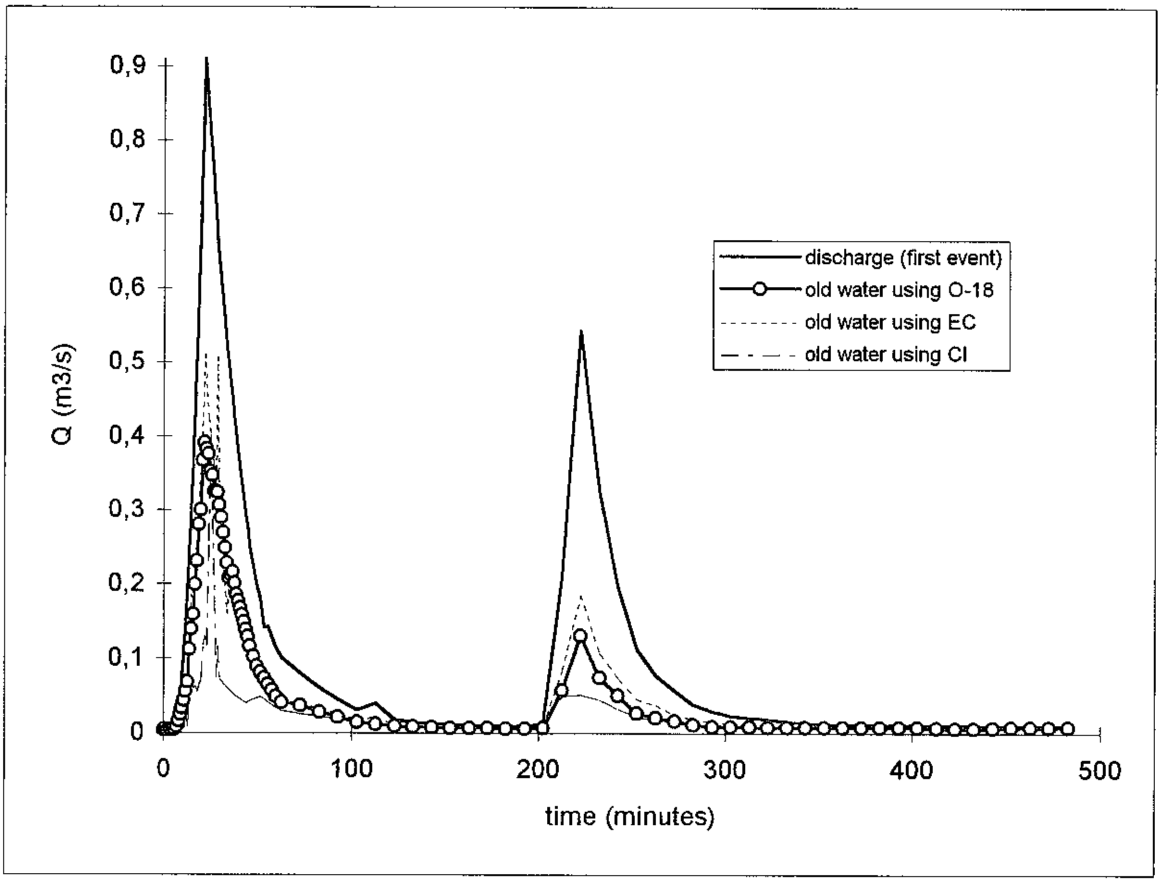


Fig. 6.2 The hydrograph separation using $\delta^{18}O$, EC, Cl^-, NO_3^- and SO_4^{2-} during the storm event of July 3rd, 1995, in the Loechernbach basin

According to the calculation, event water forms approximately 61 % of the peak flow rate, with an average proportion of about 61 % determined from $\delta^{18}O$. With the exception of EC, using chemical parameters resulted in higher event water proportions (Tab. 6.1). Here solute compositions of event water are assumed to be the same as measured in rain water (2.2 mg/l for Cl^-, 2.0 mg/l for NO_3^-, and 5.5 mg/l for SO_4^{2-}). EC is, however, assumed to be 100 μs (Chapter 5.2.2).

6.1.2 Relationship between oxygen-18 and hydrochemical tracers and discharge

Stream water chemistry responds closely to changing discharge throughout the event. This presumably demonstrates both changes in the chemistry of water following different flowpaths and the changing relative contributions from different flow components.

A comparison of $\delta^{18}O$ and discharge is presented in a mixing diagram (Fig. 6.3). The discharge rose from 10 l/s to 902 l/s during the first peak. The maximum discharge of the second peak is 542 l/s. $\delta^{18}O$ first went up sharply at the rising phase of the first event, contributed to by a dominant rainfall discharge from channel and asphalt roads, then contribution of pre-event water to streams is on the increase. After discharge exceeded 20 l/s, the ratio of event and pre-event water became relatively stable. As the result, $\delta^{18}O$ content during the following rising limb remained static with the increase of the discharge, and responded similarly during the recession limb. This phenomena is very different from the theoretical procedure from SKLASH & FARVOLDEN (1979) where ^{18}O content in streamwater continuously varied with the increases of event water proportion. The rapid and continuous increase of the groundwater contribution in the study basin, even before the peak, can be a contribution of the amount of subsoil pipes. After the peak, the curve returned the same way. Rainfall and groundwater decreased simultaneously. In the last phase, the absence of rainfall results in a sharp decline of the $\delta^{18}O$ values.

The instantaneous increase of pre-event water is remarkable, this can be explained by the piston-flow effect (BUTTLE 1994). In the soil containing macropores, pre-event water can be pushed out quite rapidly due to lower friction. A groundwater ridge, which causes enhanced discharge of pre-event water, can also be generated faster in an area with soils containing macropores. In the present loess sites, the exchange of water in upper- and lower layer is believed to be very large, so the recharge of groundwater is remarkable, in particular in dry period.

The ^{18}O-Q relationship during the second event demonstrated a slight contribution of the third component, or vadose water, demonstrating hysteresis (counter-clocking). The contribution of the third component cannot be accurately determined, due to the missing data during the rising limb, but the path of the rising limb may be very similar to that of the recession limb. Generally speaking, this contribution is limited, because the values in the falling limb were only slightly heavier.

To check the flow path in the basin, the analysis of EC and Cl^- values against discharge was also used. The Cl^--Q (Fig. 6.4) and EC-Q (Fig. 6.5) relationships for the second event are very similar to that of ^{18}O-Q (Fig. 6.3). The irregular rising limb of Cl^- and EC for the first event is related to runoff from roads treated with deicing salts which caused the increase of Cl^- concentrations for a short period with discharge between 23 and 45 l/s. The marked counter-clockwise curve of EC-Q relationship is clearly related to the "dilution effect", indicating that the EC values for event water are significantly higher than those in precipitation during the initial phase of the storm. This thereby confirms the hypothesis dealt with in Chapter 2.5.2 and the measured values of EC in the field dealt with in Chapter 5.2.3.

Comparing Figure 6.4 and 6.5 (Cl^--Q, EC-Q) with Figure 6.3 ($\delta^{18}O$-Q), the Cl^- concentrations during main peak phase are found to be close to the Cl^- concentration of rainfall. A most possible explanation is the inaccuracy of the estimated average $\delta^{18}O$ content in precipitation. This rainfall event seems to be a typical summer thunderstorm. The rain sampling gauge PL at Loechernbach is far from the center of the storm, so that the representativeness of the rain sampling collected at PL is very questionable. In addition, this inaccuracy may also be due to the change of rain intensity and fractionation on $\delta^{18}O$ content. Hence, event water estimated using Cl^- and SO_4^{2-} are more reliable than those of $\delta^{18}O$ and EC for summer rainfall events.

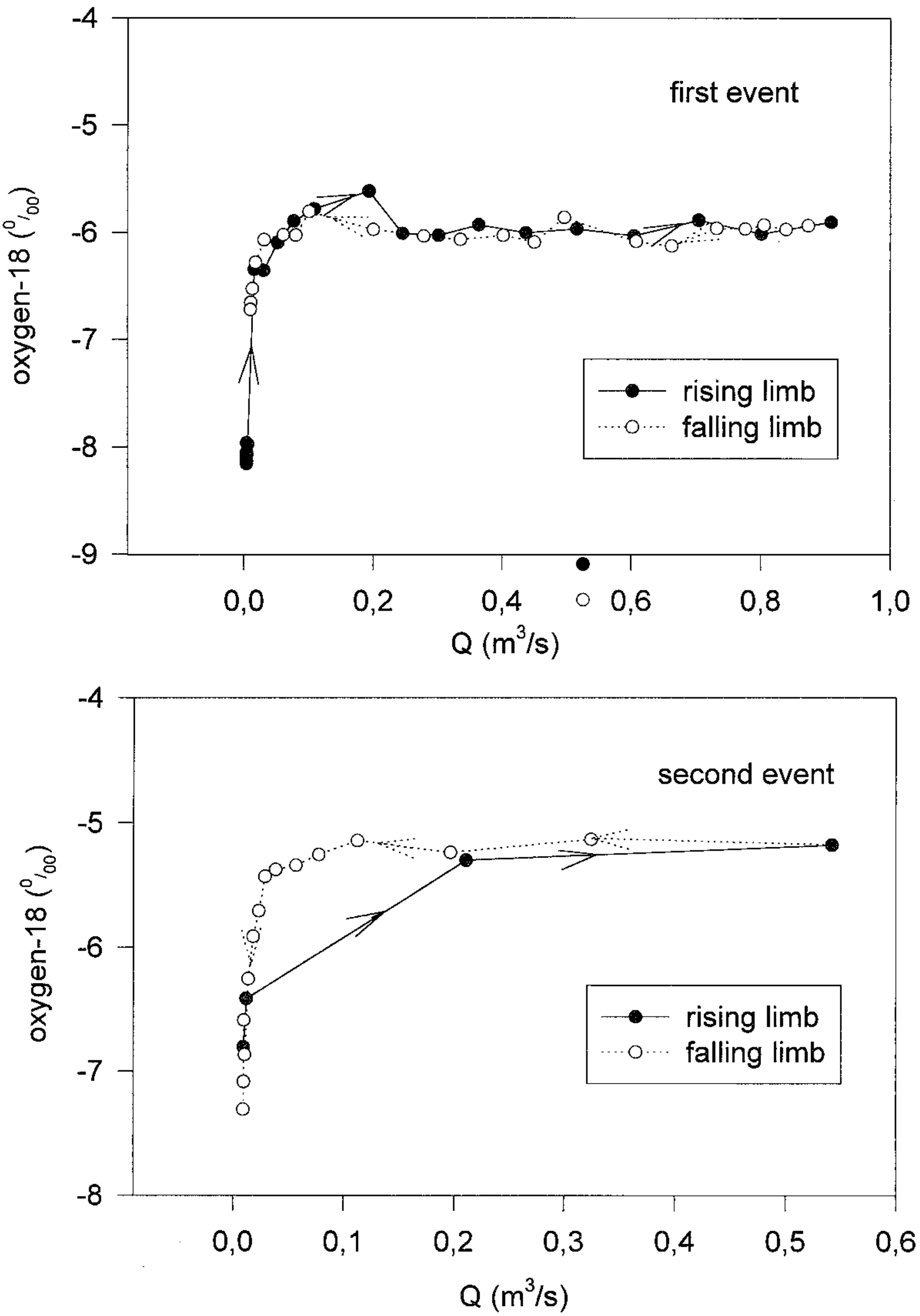
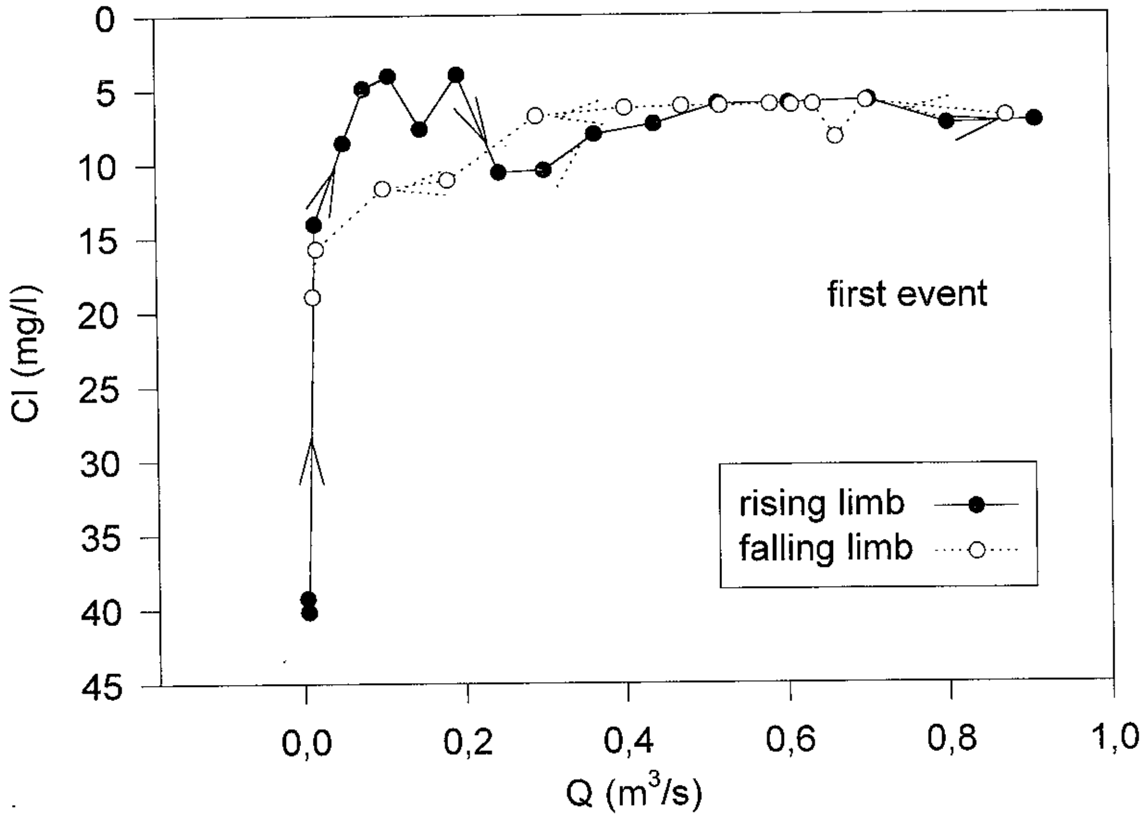


Fig. 6.3 Generalized version of the discharge - $\delta^{18}O$ relationship
July 3rd, 1995, Loechernbach



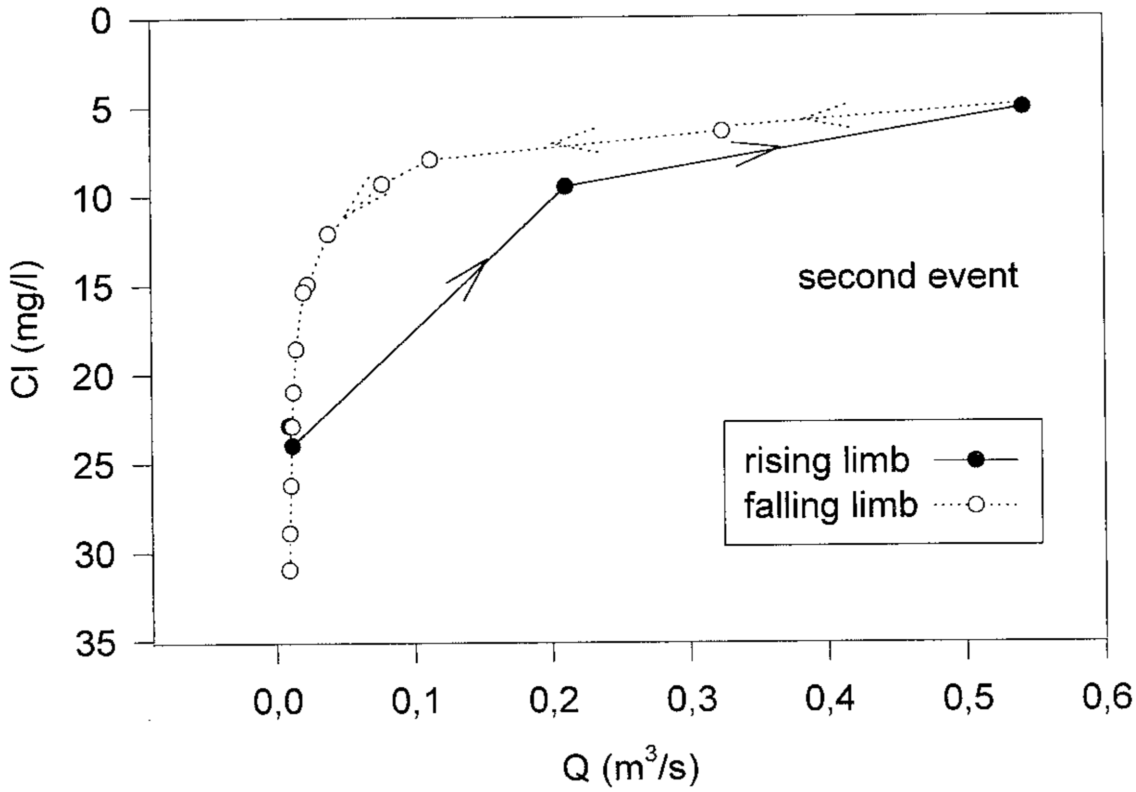
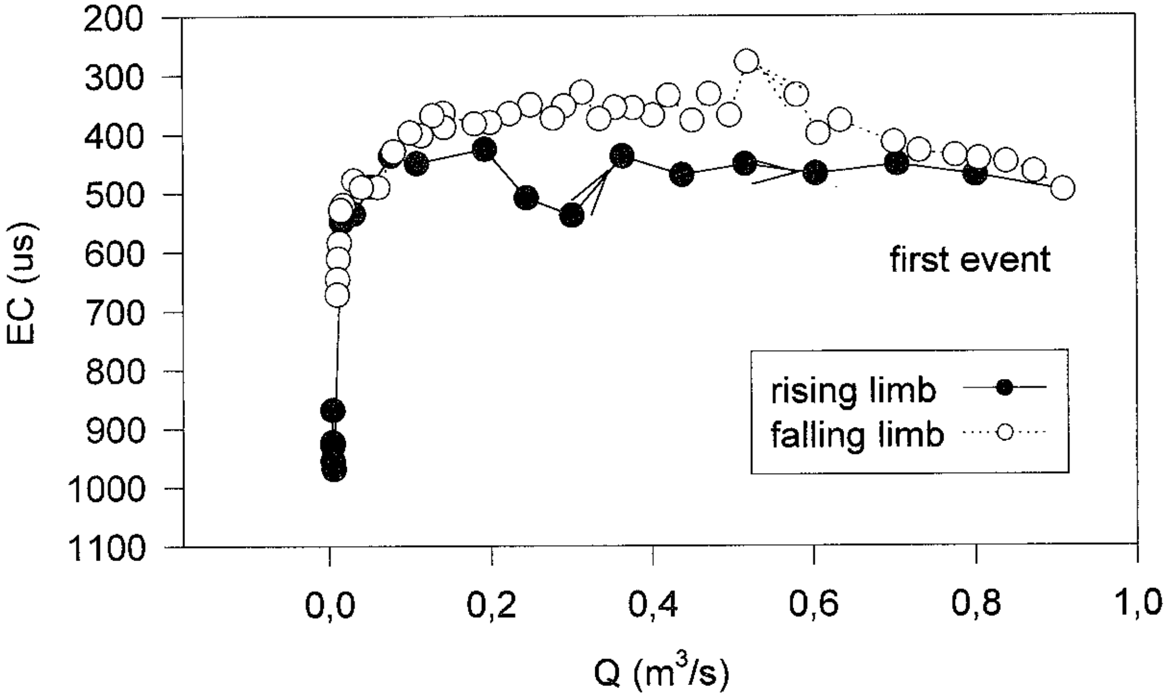


Fig. 6.4 **Generalized version of the discharge - Cl^- relationship**
July 3rd, 1995, Loechernbach



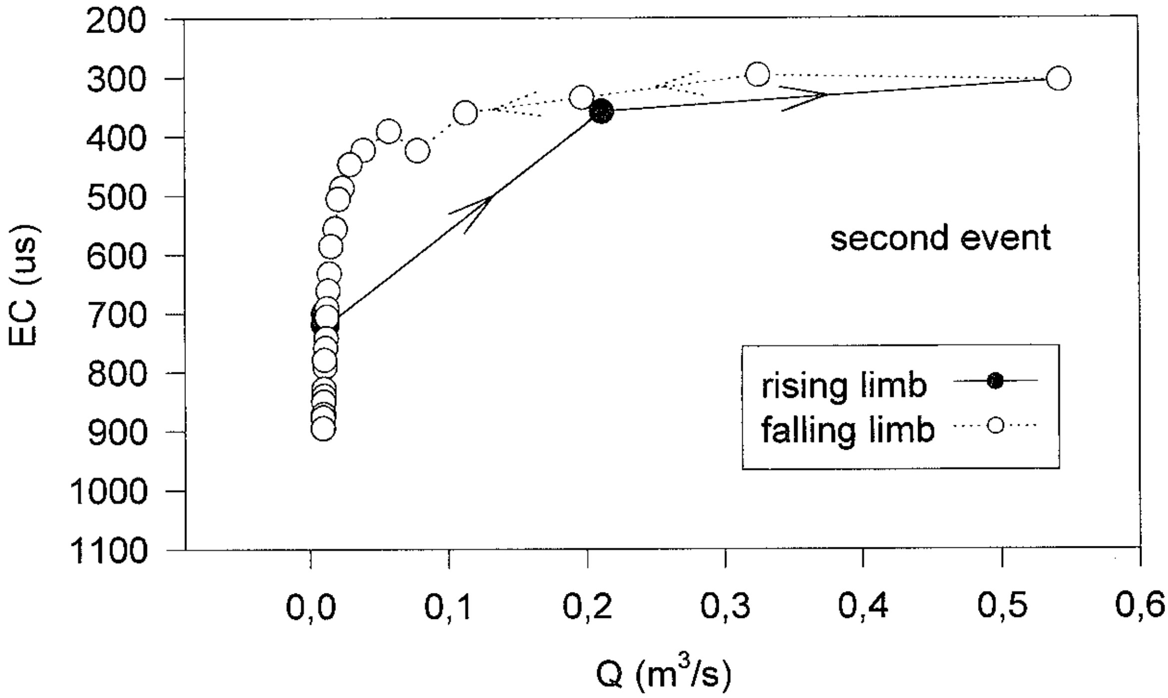


Fig. 6.5 Generalized version of the discharge - EC relationship
July 3rd, 1995, Loechernbach

EC-Q relationship (Fig. 6.5) is more complicated than that of Cl^--Q. The EC values during the rising limb toward the peak are about 450 μs, which is higher than the observed 100 μs of event water derived from asphalt roads (including overland flow from natural area). This reflects the marked contribution of fast subsurface flow through drainage which had been computed from a fast uptake of ions in the soil. It indicates that the dissolution from silty soils can occur very rapidly.

6.1.3 Sensitivity analysis

Runoff separation of short-term events is more sensitive to any change of solute compositions in event water, compared with long-term separation. Therefore, sensitivity analysis for individual events is necessary. The aim in this section is to check the probable change of event water fractions against assumed series of solute concentrations of event water, and to compare the diversity of the sensitivity between solutes.

Sensitivity analysis is based on Equation 5.1 (Chapter 5). The test results of the sensitivity are listed in Tab. 6.2. As an example, the increase of Cl^- value in event water from 2 mg/l to 10 mg/l results in the increase of event water proportion by about 22 % (from 78 % to 100 %), while the corresponding change by long-term separation would be only 3 % (from 86 % to 89 %).

Comparing the results in Table 6.2 with 61% event water fraction obtained from $\delta^{18}O$, it becomes clear that a higher event water fraction (for example, >78 % for Cl^-) would occur in the case of higher solute composition occurring in event water. This applied to Cl^-, NO_3^- and SO_4^{2-}.

On the other hand, the direct use of EC value in precipitation as event water (measured 35 μs for this event) can greatly underestimate new water fraction for the present study basins. By inference, this underestimate of new water proportion may be over 20 % (Tab. 6.2). According to the field observations for overland flow, the average value of EC for surface flow from asphalt roads should be at least about 100 μs (Chapter 5.2.2), with a corresponding value of 59 % event water. The latter in discussion will demonstrate that this estimate is most likely too conservative. If considering the higher concentrations of overland flow from fields and rapid interflow from drainage, the EC values can be higher than 200 μs.

6.1.4 Residence time of surface runoff obtained by artificial tracers

During the sampling for the storm event, 1 g of the artificial tracer uranine was injected at the asphalt road near the weighted point of the basin - approximately 1000 m away from the gauge station. It is assumed that the corresponding residence time of the tracer represents the mean value of surface runoff. The concentration of the traced water was measured at the output of the basin.

Tab. 6.2 Sensitivity analysis of different chemical compositions by assumed series of the solute concentrations of event water (July 3rd, 1995, Loechernbach)

Number	*1*	*2*	*3*	*4*	*5*	*6*	*7*	*8*	*9*	*10*	*11*
Cl^- values (mg/l)	0	1	**2.2***	3	4	5	6	7	8	9	10
event water Cl^-	73%	75%	**78%**	80%	83%	85%	88%	92%	95%	99%	100%
NO_3^- values (mg/l)	0	1	**2***	3	4	5	6	8	10	11	12
event water NO_3^-	70%	72%	**74%**	76%	78%	80%	83%	88%	94%	98%	100%
SO_4^{2-} values (mg/l)	0	2	4	**5.5***	7	10	12	14	16	20	24
event water SO_4^{2-}	75%	77%	78%	**79%**	81%	84%	86%	88%	90%	94%	100%
EC values (mg/l)	0	**35**	50	75	100*	130	150	200	250	300	400
event water EC	52%	54%	55%	57%	**59%**	61%	63%	67%	73%	79%	95%

bold: measured values and corresponding new water fractions
* assumed values for new water

The convection-dispersion model (CDM) from MALOSZEWSKI (1991) was used for the interpretation of the breakthrough curves of the tracer experiment (UHLENBROOK 1995). The rate of recovery for the tracer was 86 %. Calculated by the CDM model, a dispersivity of 1.9 m and a mean residence time of 20 minutes of the rainwater falling on sealed area during the event were obtained.

This residence time of 20 minutes agrees with the time order of the system response to rainfall events calculated by BUCHER (1984). In his study, 31 individual rain events are observed. An average residence time of 31 minutes reaching the peak (crest) at Loechernbach was obtained, also 10 minutes shorter than at Rippach, despite the larger area.

6.2 Hydrograph separation for Rippach, September 19th, 1995

A storm event at Rippach was sampled by GÄSSLER (1995) on September 19th. This allows for a comparison of runoff behavior in this small-terraced catchment with the large-terraced Loechernbach catchment (Fig. 6.6). An important fact in this investigation is the smaller event water contribution for the whole event as compared to the event at Loechernbach (Fig. 6.7). Using different parameters the similar results are obtained. Particularly, the estimate of event water using EC for the peak phase agrees with the Cl^- method.

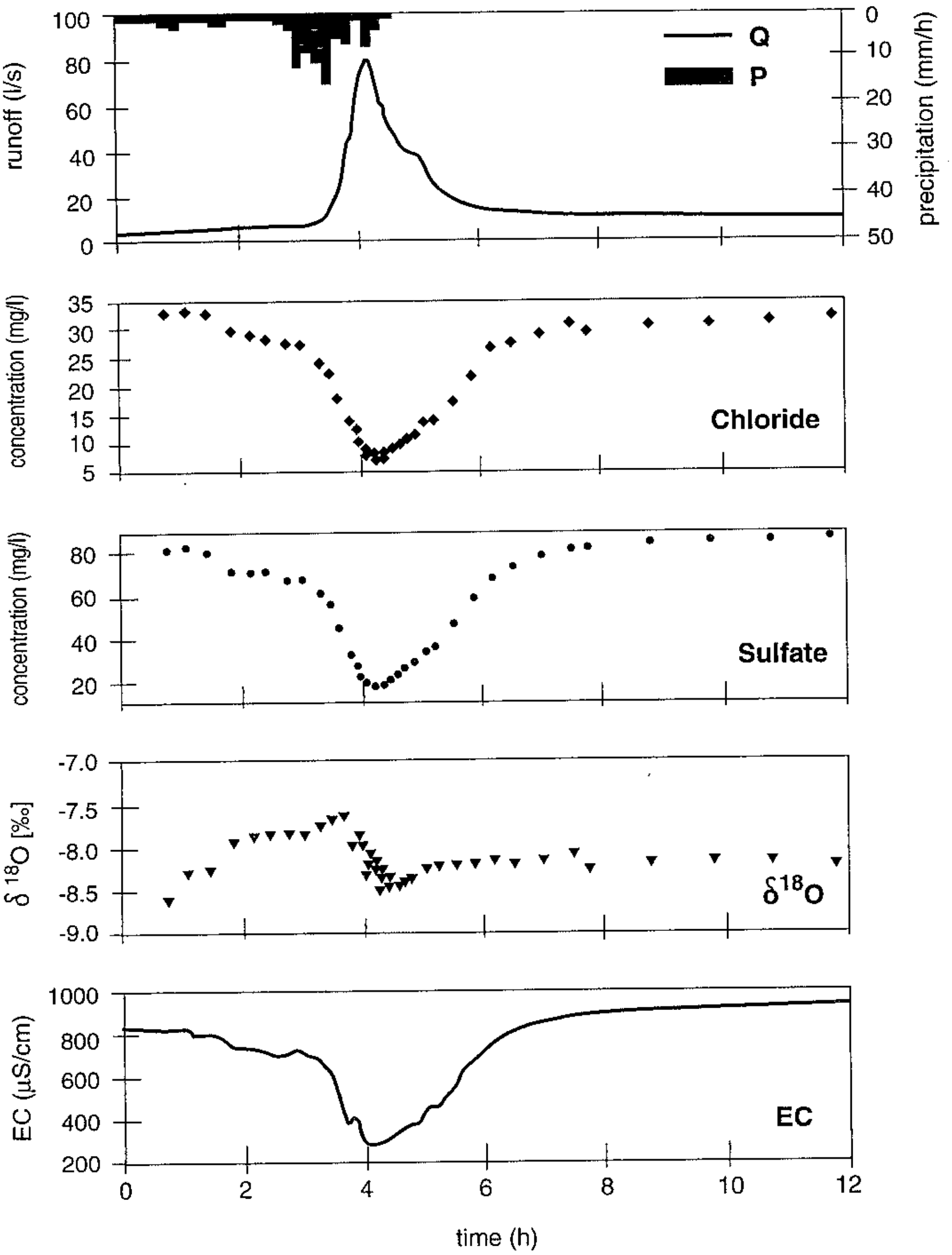


Fig. 6.6 The stream discharge, $\delta^{18}O$, EC, Cl^- and SO_4^{2-} in response to the storm event on September 19th, 1995, in the Rippach catchment (GÄSSLER 1995)

Unfortunately, the $\delta^{18}O$ value of precipitation in the latter phase (-8.26 $^0/_{00}$) was too close to the baseflow value (-8.22 $^0/_{00}$), to allow separation of runoff components with the isotopic method.

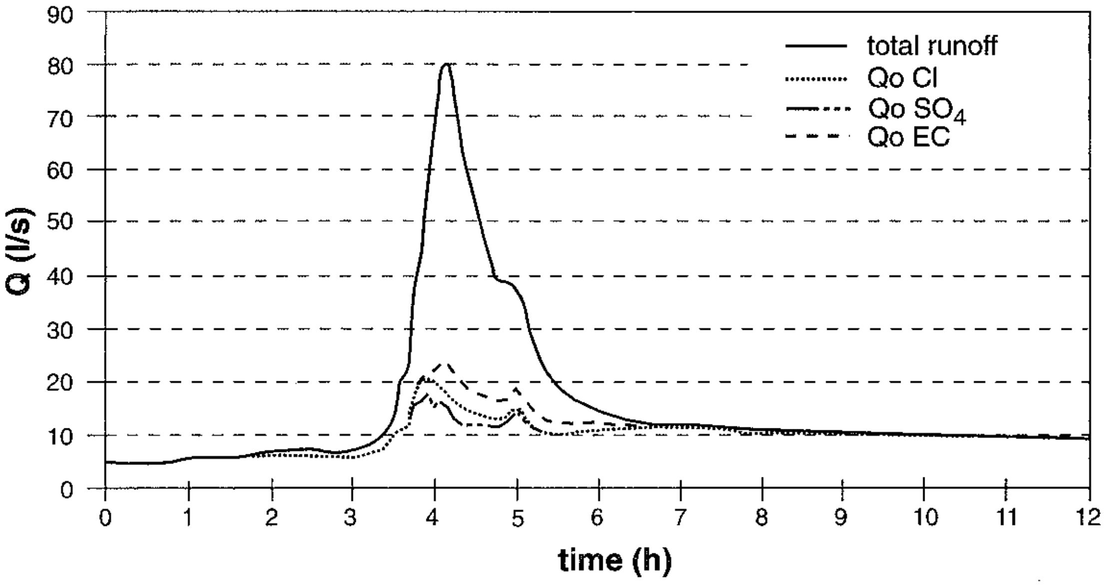


Fig. 6.7 Hydrograph separation using $\delta^{18}O$, EC, Cl^- and SO_4^{2-} during the storm event of September 19th, 1995, in the Rippach catchment (GÄSSLER 1995)

Event water fractions corresponding to various time scales of the calculation are compared in Table 6.3, to obtain indications of why event water was secondary for this whole event. Precipitation water is contributing up to 83 % to the peak flow. As an example, large event water proportion using Cl^- is observed during the main phase of the event. An objective choice of the computing time interval seems to be important. On this event, the whole event can be calculated until 8:00 a.m. of next day when 31.0 mg/l of the Cl^- concentration is close 32.9 mg/l in the beginning phase.

GÄSSLER (1995) suggested that the contribution of event water may be largely under-estimated, because the field observations detected a fast uptake of ions during flow over and in the soil. According to the calculation for assumed series of the tracer concentrations of event water, computed event water percentages are highly sensitive to the increase of total dissolved solids in event water during the peak period (Tab. 6.4, case 1), but this sensitivity is not evident for time interval of the whole event (Tab. 6.4, case 3) because groundwater is dominant in the latter phase of the event. The probable under-estimation, if any, is only up to 10 % based on the data (3) in Table 6.4. Unfortunately, this conclusion cannot be proven by $\delta^{18}O$ data for this event.

A relationships of Cl^--Q and EC-Q (GÄSSLER, 1995) are shown in a mixing diagram (Fig. 6.8). The discharge rose from 5 l/s to 80 l/s during the event. The first rising phase at Rippach is not so sharp as occurred at Loechernbach, indicating a lower contribution of direct runoff. The subsequent rise of discharge exceeding 10 l/s is less drastic and shows a dilution of solutes due to the parallel increase of rain and groundwater, but dominated by rain water. The recession with lower Cl^- concentrations differs from that at Loechernbach, indicating a small contribution of soil water.

Tab. 6.3 Event water contributions of two short-term storm events
September 19th, 1995, Rippach*

	Cl^-	NO_3^-	SO_4^{2-}	EC
1) peak phase (19:45 - 21:35)	65%	70%	69%	66%
2) main phase of event (18:00 - 21:35)	56%	61%	60%	58%
3) whole event (18:00 - 08:00)	33%	40%	30%	27%
4) observation period (17:00, 9th - 18:00, 10th)	28%	34%	23%	19%

* original sampling data from GÄSSLER (1995)

Tab. 6.4 Sensitivity analysis of different chemical compositions by assumed series of the tracer concentrations of event water
September 19th, 1995, Rippach

		1	*2*	*3*	*4*	*5*
Cl^- in event water[mg/l]		0	**0.5**	2	4	10
event water with Cl^-	1)	64%	**65%**	68%	76%	96%
	3)	39%	**40%**	42%	47%	59%
NO_3^- in event water [mg/l]		0	**1**	5	10	12
event water with NO_3^-	1)	69%	**70%**	78%	90%	96%
	3)	46%	**47%**	52%	60%	64%
SO_4^{2-} in event water [mg/l]		0	**5**	10	20	24
event water with SO_4^{2-}	1)	67%	**71%**	76%	89%	95%
	3)	37%	**39%**	42%	49%	52%
EC in event water [µs]		0	20	**100**	200	300
event water with EC	1)	58%	60%	**66%**	77%	91%
	3)	30%	31%	**34%**	40%	47%

1), 3) see Tab. 6.3; bold: measured values

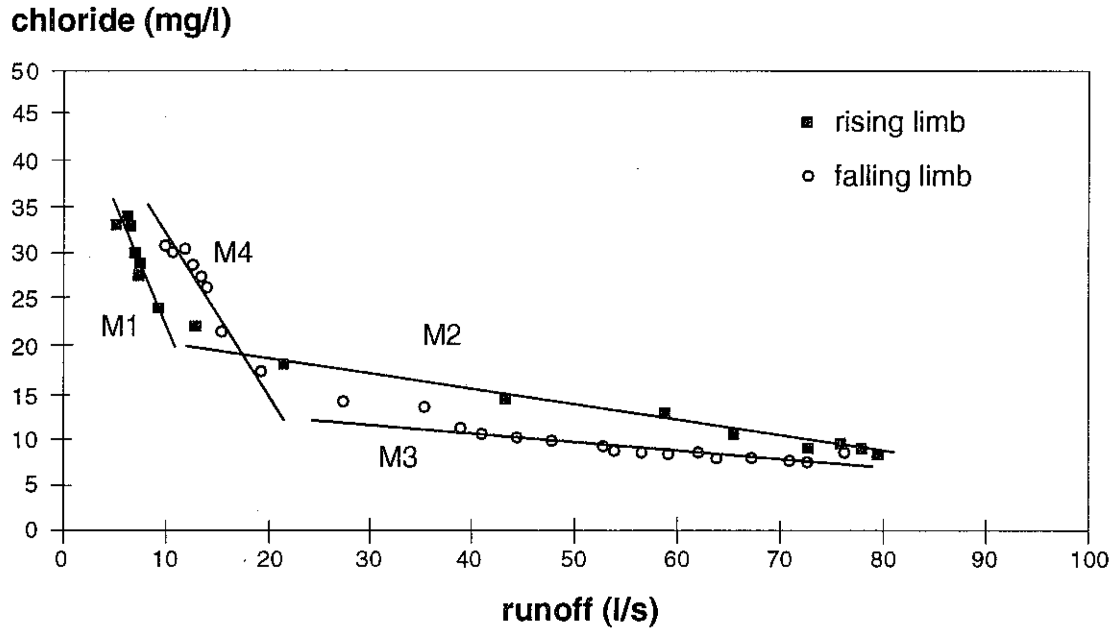


Fig. 6.8 Generalized version of the discharge-chloride relationship for storm event of September 19th, 1995, at Rippach (GÄSSLER 1995)

6.3 Runoff dynamics of the unsaturated zone

The application of the DIFGA model (Chapter 4) shows that direct runoff in the altered Loechernbach catchment plays an important role, in particular during the storm events. To understand runoff dynamics of unsaturated zone, two tracing experiments in the unsaturated zone are performed simultaneously (UHLENBROOK 1995).

Loechernbach

This tracer experiment was chosen in farmland with an artificial drainage system at the mid-point of the basin. The results of the experiment are briefly summarized below, and more details are given in UHLENBROOK & LEIBUNDGUT (1997).

A so-called "double-frame system", composed of two rectangular frames (17 and 8 m^2) and based on a double-ring infiltrometer, was constructed. Using this system, the tracers, 460 g of the fluorescent dye uranine and 12 kg of the salt sodium bromide, were applied on the inner frame. This method should minimize the loss in the unsaturated zone due to capillary suction (MEHLHORN et al. 1995). The first 12 hours recovered 90 % of the injected tracers. The total recovery rate amounted to 12.5 % for uranine and 13.4 % for bromide.

Three models were used to simulate the breakthrough curves of the tracer experiment: the convection-dispersion model (CDM, MALOSZEWSKI et al. 1984), the single fissure dispersion model (SFDM, MALOSZEWSKI 1994), and the transfer function model (TFM, JURY 1982). Both breakthrough curves have a similar shape with a small "pre-peak" (Fig. 6.9), a defined main peak, and a shoulder at the recession limb. The uranine concentration increased earlier than the bromide concentrations.

The first small peak may connect directly to mouse hole and rain worm-canal over a subpipe. The following major peak relate to macropore flow in the agricultural land. These macropores transported the most tracer during the first 11 hours. The third tracer peak dealt is believed to result from matrix flow. Compared with macropore flow, this component is small. The mean residence time of this third peak is 212 minutes, the corresponding K_f is 2.0×10^{-5} m/s.

Based on the tracing experiment, the mean residence time for rapid subsurface flow is less than 2.5 hours (Fig. 6.9). Excellent fittings were obtained for the modelling of the solute transport in the unsaturated zone, by simulating the flow in the macropore system with the SFDM and the flow in the micropore system with the CDM.

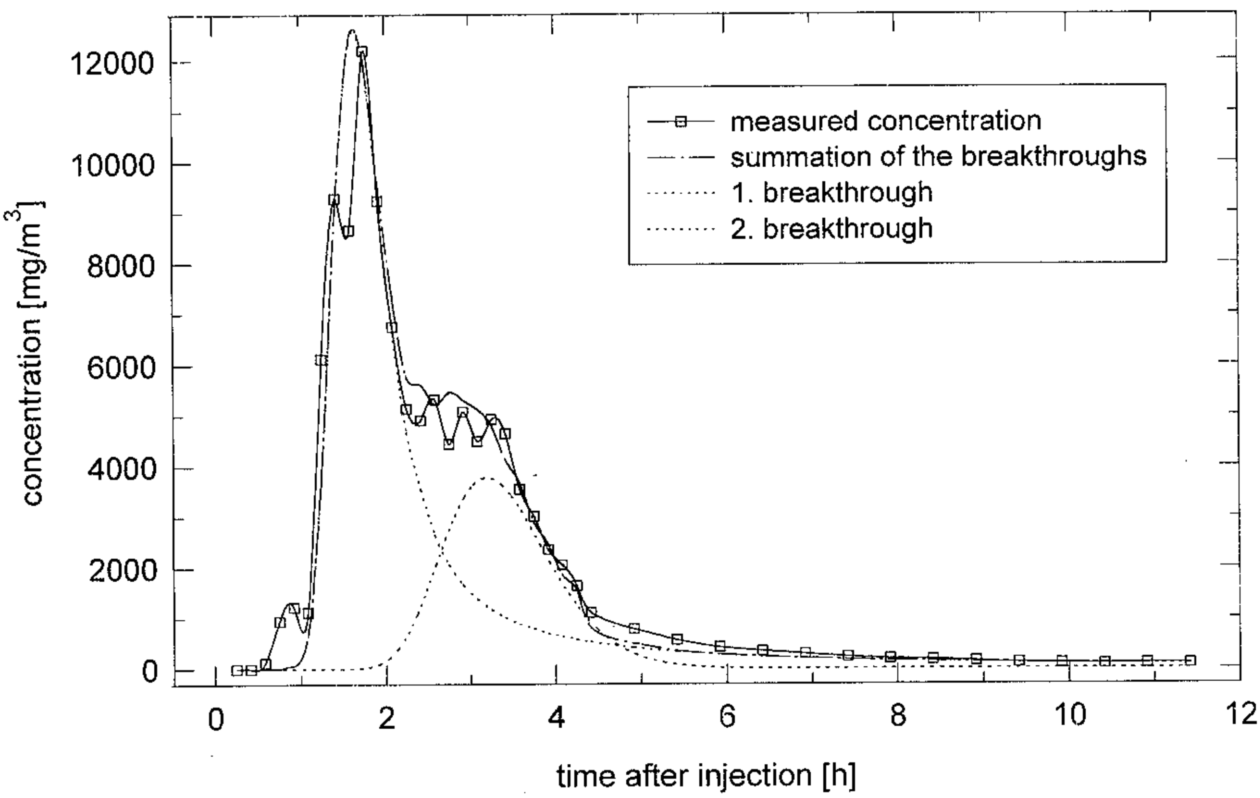


Fig. 6.9 Simulation of uranine concentration graphs for the tracing experiment in the unsaturated zone of Loechernbach on July 3, 1995 (UHLENBROOK 1995)

Rippach

In the Rippach basin, the tracer experiment was carried out on a small terrace (UHLENBROOK 1995). The boundary conditions and the methods were similar as for the experiment at Loechernbach. However, no tracer was recorded during the observation period of two months. The examination of soil samples revealed the absence of macropores . The different land use during the last few years seems to be responsible for the different macroporosity.

It is concluded that a long observation period is required for artificial tracer experiment of water movement in unsaturated zone. This is consistent with the result of a tracer experiment in unsaturated zone in April 1979 by KÄSS (1992), in which appreciable tracer concentration wasn't observed until some five years later after a strong storm event.

6.4 Discussion

The sensitivity analysis for storm events shows that the diluted salts of the roads and dissolution in the upper-layer of soil can increase the event water fraction. However, the observations demonstrated that EC is probably more sensitive to increased values in new water. Low EC values designed for event water is most likely the reason for the underestimation of event water contribution to stream.

A loess solution experiment by LANGHAMMER (1988) support the rapid dissolution of EC noted in this study. the influence of higher CO_2 on EC value is quantitatively determined for the silt soils in LANGHAMMER's studies. The EC with CO_2-degassing increased after 24 hours to 1000 µs, while the EC without CO_2-degassing increased only to 50 µs. It was suggested that lime solution was responsible for the high EC of pre-event water.

A further comparison of the solute concentrations of event water is listed in Table 6.5. The measured minimum solute concentrations during the storm events indicates that the maximum solute concentrations of event water can not go beyond these limits. Clearly, the probable dissolution of Cl^-, NO_3^- and SO_4^{2-} in surface runoff would only double the concentrations of event water, but the EC values might increase 8 times to that in rain water (from 35 µs to over 250 µs) in the study sites.

Assuming 65 % of event water contribution using $\delta^{18}O$ during peak period at the July event is correct, then the EC value in event water would be anomalous -38 µs; assuming 90 % of new water using Cl^- is correct, then the EC value would be 212 µs in event water. It indicates that event water estimation using $\delta^{18}O$ is critical in this rainfall event. Following question arises: is the EC value in event water really higher than 200 µs? The comparison for the event in September indicates that 100 µs of EC for event water in this catchment is objective (Tab. 6.5). This means that the EC value for event water in the Loechernbach catchment is higher than that in the Rippach. The explanation may be the contribution of macroflow via 20 km subpipes installed in this basin. For that, the event EC of over 200 µs for individual events in this altered basin is possible.

Tab. 6.5 The measured peak solute concentrations at output of stream for the observed events for July and September 1995

		Cl^- (mg/l)	$\delta^{18}O(^0/_{00})$	EC (µs)
Loechernbach	event water / pre-event water	2.2/37.3	-4.26/-8.08	35/868
July 3rd, 1995	runoff concentration during peak runoff	4.0	-5.61	278
	calculated concentration of event water			
	if event water = 65% (^{18}O)	*-13.8*	*-4.26*	*-38*
	if event water = 90% (Cl^-)	2.2	-5.63	212
Rippach	event water / pre-event water	0.5/33.0	-8.06/-8.26	24/830
Sept. 19th, 1995	runoff concentration during peak runoff	7.7	-8.23	263
	calculated concentration of event water		-	
	if event water = 78% (Cl^-)	0.5		104

Clearly, EC is not a conservative tracer for the runoff separation within the study sites. However, the 17 % difference of event water fractions obtained by EC (61 %) and by Cl^- (78 %) can be explained to be rapid dissolution of precipitation in near surface soil via subpipes.

The $\delta^{18}O$ content in rainfall on July 3rd is the average value of a whole rain sample, ignoring the influence of change in intensity of precipitation on $\delta^{18}O$. This simplification can reduce the accuracy of the estimation for the first peak event. This may explain different amounts of pre-event water based on isotope and hydrochemical tracers. It is believed for this event that event water fraction based on Cl^- rather than from stable isotope data is more reliable and to some degree, more accurate.

6.5 Summary

Hydrograph separations using $\delta^{18}O$ and hydrochemical tracers for two storm events are analyzed separately. The results shows a dominant event water fraction throughout the storm event on July 3rd in the altered catchment Loechernbach, while the event water only occurs during the peak phase of the storm event in the unaltered catchment Rippach. These results are consistent with the results obtained by DIFGA and other previous investigations in the study sites. It is noted that event water dominates short-term events, while old water dominates long-term period simultaneously at Loechernbach. After terracing, larger percentage of event water comprise individual streamflow hydrographs. In addition, the investigation show that there is no evidence for a third component (vadose water) during these events based on tracer or stable isotopic methods.

The sensitivity calculation shows that EC is very sensitive to increases of event water EC value. An EC value chosen too low for event water would result in marked under-estimation of the event water fraction to streamflow.

The mean residence time (20 minutes) of fast direct runoff at Loechernbach is consistent with the results of previous studies. This residence time is short due to the small area of the catchments, which allows rapid delivery of direct runoff into the stream network.

The results of the tracing experiment in the unsaturated zone in the Loechernbach catchment revealed a dominant macropore system and the influence of drainage pipes. This thereby explains why the system rapidly responses to a storm event. The mean residence time of 2.5 hours may be assumed as the average time of *QD2* (slow direct runoff) for this basin.

7 Comparison and discussion of the results obtained by various methods

This study of runoff generation has included both model and tracer methods. The comparisons between these methods provides not only insight into the application of the models for regional hydrological modelling, but also additional information about runoff generation, which is impossible to obtain by the use of any one method or tracer individually. These comparisons lay a foundation for the later combination of the results.

7.1 Comparison of the models applied in the study basins

Four different hydrological models have been used for the hydrograph separation in the study catchments. They include the HYSTATRI model (LUFT & MORGENSCHWEIS 1982), the USDAHL model (VOGELBACHER 1985), the DIFGA model (LEIBUNDGUT & CUI 1994) and PRMS model (BLAU 1996).

In HYSTATRI model 1980 daily and hourly hydrograph separations were used by means of statistically characteristic values (without direct inclusion of precipitation, LUFT & MORGENSCHWEIS 1982). Runoff is divided into direct runoff and base flow in the Rippach catchment. The model determines each hourly runoff value by means of a past 24-hour range and variation coefficient compared with threshold values of existing dry weather runoff or separation as base and direct runoff (LUFT 1980).

For the same purpose of runoff separation, the USDAHL model was applied by VOGELBACHER (1985) in the Loechernbach catchment. In the USDAHL model (U.S. Department of Agriculture Hydrologic Laboratory), special account was taken of impervious areas built up as asphalt roads. The soil moisture component of the model was tested separately in eight representative measurement sites. Runoff is separated into three components. The comparison of these components shows for the large-terraced basin an increase in surface runoff of 3.5 times of the unchanged basin, an interflow which is 2.5 time higher and a decrease in baseflow of about 10 %.

The modularity structured basin model PRMS (Precipitation-Runoff Modelling System, LEAVESLEY ET. AL. 1983, FLÜGEL & LÜLLWITZ 1993), was applied by BLAU (1996). The PRMS model accounts for the heterogeneity of hydrological systems, thereby permitting model adaptations to various hydrological regimes. To reproduce the physical reality of the hydrological system, each process of the hydrological cycle is expressed in the form of known physical laws or empirical relationships that have some physical interpretation based on measurable basin characteristics (LINDENLAUB et al. 1997). Runoff is separated into surface water, subsurface flow and groundwater.

Tab. 7.1 Comparison of the models applied in the study sites

	HYSTATRI	DIFGA	USDAHL	PRMS
type	continuous, without simulation and forecast	continuous, component model without simulation	continuous, deterministic and simulation model	continuous, deterministic and simulation model
parameters	runoff	precipitation, runoff	precipitation, evaporation, temperature and other parameters	precipitation, temperature, solar radiation
data	hourly data, without precipitation	daily values of more than 10 years	daily/hour values	daily values
concept	statistical method of separating discharge hydrograph by means of statistical characteristic values	parallel circuit of single linear-storages (maximum 4 storages), correct the separation using monthly water balance evaluation.	allow a horizontal subdivision for maximum 4 zones; vertical: maximal 5 storages	modular-designed, semi-distributed and physically based modelling with 3 storages
surface water *QD1*	the difference of discharge and baseflow	calculated as the remanant of the difference between discharge and other slower flows (*Q-QG2-QG1-QD2*).	water from sealed and saturated area, using $P\text{-}Q = \Delta D$, (*D*: mean surface runoff); 2 parameters: length of roads, falling gradient	water from sealed and saturated area, using a contributing area concept with nonlinear exponential function
rapid interflow *QD2*		same procedure as *QG1* using semilogarithmic plot (*Q-Q2-Q1*)	assuming: upper layer: only horizontal movement (20 cm); under layer: only vertical movement, 5 soil moisture parameters, 2 infiltration parameters; 5 evaporation parameters	as either a linear or nonlinear function of subsurface storage. Surface flow occurs when groundwater recharge exceeds an user defined rate.
slow interflow *QG1*		same procedure as *QG2*, using semilogarithmic plot (*Q-Q2*)	linear storage constants	
groundwater *QG2*	baseflow: based on the dry weather runoff using simple statistical characteristic values such as hourly variation coefficients and monthly threshold values.	linear, as envelope of the hydrograph using semilogarithmic plot during the low flow periods. Groundwater level for reference for determination of storage constant during dry periods	linear storage constant	as a linear function of groundwater storage

Table 7.1 summarized the results of the hydrograph separation using these models. All of them are continuous analysis of runoff components. HYSTATRI and DIFGA give no prediction of runoff behavior. They are for the purpose of water balance, and are not concerned with what happens in the system. The former is based on statistical methods, the latter based on graphical procedures. USDAHL and PRMS as physically-based models use subdivisions, having considered the spatial variations and different agricultural land use separately.

Surface runoff

In HYSTATRI and DIFGA, surface runoff is the difference of discharge and other components. For DIFGA, all errors which occur in the slower components are accumulated in the fast component, therefore the calculations are inaccurate for direct runoff. In contrast to this, USDAHL and PRMS allow a direct determination of this component. USDAHL uses a non-linear exponential function, and considers the catchment characteristics, e.g. the length of asphalt roads and the corresponding falling gradient. The results listed in Table 7.2 show that USDAHL and PRMS produce similar percentages of surface runoff for the Rippach catchment, but different percentages for the Loechernbach catchment. In the altered basin surface runoff is obviously sensitive to different treatment methods.

Interflow

Interflow is divided into rapid and slow parts. DIFGA consider only the temporal residence of the components. USDAHL considers evaporation, soil moisture and other characteristic values which are important for the basins with evaporation of 3/4 of the precipitation amount. A direct comparison of this component between USDAHL and other models is difficult because the definition of interflow by each model is quite different. USDAHL separates the soil into upper layer (20 cm deep) and lower layer (160 cm deep). The results obtained from the USDAHL model suggest that interflow be an important component in the research basins. In particular, 31 % of interflow in the Loechernbach catchment reflect the change of the ground structure after the terracing (KELLER 1985). Rain water in the Loechernbach catchment can rapidly enter the stream through the marked macropore flow via subpipes. This part of soil water does not distinguish greatly from overland flow in the time scale by DIFGA. This may explain the reason why the rapid interflow *QD2* in DIFGA is smaller than that obtained by USDAHL and other models.

Groundwater

All four models use the linear-storage concept to determine groundwater component. Hence, their results for this component are similar. DIFGA has an advantage for this component, benefited by the long-term time series of data (e.g. 10 years). Of course, the objective determination of groundwater storage constant by DIFGA often refers to catchment information, i.e. groundwater data. The HYSTATRI model also produced better results than other models. However, the separation by HYSTATRI consists of extending the baseflow recession curve, which occurs before the surface runoff (Fig. 2.1). This method is not suitable for study basins with large amounts of groundwater. The experimental investigations confirmed that the groundwater increased very quickly in the rising phase of the hydrograph. As shown in Table 7.2, the lowest baseflow values occur using PRMS, the reason possibly being that a part of the deep soil water has been divided in the second component.

Tab. 7.2 The results of components analysis from different models
Rp: Rippach, Lo: Loechernbach

Catchment	component	HYSTATRI (1972-1975)	DIFGA Rp (1972-1975/1980) Lo: (1977-1980)	USDAHL (1973 - 1980)	PRMS Rp: (1972 - 1980) Lo: (1977 - 1980)
Rippach	surface runoff	17%	20% / 23%	7%	8%
	rapid interflow			16%	22%
	slow interflow	83%	34% / 31%	77%	70%
	baseflow		46% / 46%		
Loechernbach	surface runoff		37%	19%	31%
	rapid interflow		4%	31%	40%
	slow interflow		38%	50%	29%
	baseflow		21%		

Overall, similar results obtained by different modelling methods are given for the Rippach catchment. Groundwater dominated streamflow in this basin throughout the entire year. It is therefore believed that the hydrological behavior is relatively simple. However, the modeling results of the Loechernbach catchment differ from each other. High direct runoff fraction and the complexities of the ground structure after terracing make the calculation difficult. A direct comparison of models is difficult because each model has different boundary conditions. Furthermore, each model component (i.e. interflow) is defined differently.

A conceptual model should be simple, and simulations should be as close as possible to the real physical situation. However, both demands are often at conflict. Physically-based models are over parameterized in a systems simulation sense (HORNBERGER ET AL. 1985). In a complex catchment it is questionable how representative these parameters are even if the results were obtained through experiments in several points. Both USDAHL and PRMS are physically-based models. The attempt of these models to use various parameters to describe the spatial variability of a catchment does not seem to be more successful than the simple DIFGA model. DIFGA is easily applied. For many basins with simple hydrological behavior the results using DIFGA yield a reliable approximation. The Rippach catchment is an example.

7.2 Comparison of the results of the different experimental approaches

Hydrochemical tracers may be altered in their flow path so that the soil chemical processes often make interpretation of most chemical parameters of stream waters too difficult for simple use (MAULÈ & STEIN 1990). In this study, SO_4^{2-}, NO_3^- and EC are believed not to be conservative during the observed long-term period. Theoretically, stable isotopes are not altered by the flowpath. They can, therefore, accurately distinguish new water (event water) from old water (pre-event water). Two factors influencing the accuracy of the separation estimate using stable isotopes are the easily occurring variations of the isotope values in precipitation due to evaporation, related to the change of intensity of rain, and the seasonal variation in precipitation. The latter refers to long-term separation.

7.2.1 Comparison of the long-term results obtained by oxygen-18 and hydrochemical tracers

To test how conservative Cl^- is, a comparison of old water fractions between both $\delta^{18}O$ and Cl^- for the weekly data is shown in a histogram (Fig. 7.1). Of the fifty-nine weekly samples contained in the histogram, 73% of the Rippach samples and 71% of those of the Loechernbach lie between ±0.10. All samples lain out of ±50% are due to the erratic $\delta^{18}O$. These results are satisfactory for long-term separation. As mentioned above, long-term separation differs from short-term separation. For the latter the erratic behavior of $\delta^{18}O$ is normally not obvious, therefore a better correlation of the results between Cl^- and $\delta^{18}O$ in short-term separation is given.

A further comparison of monthly runoff separations between $\delta^{18}O$ and Cl^- is summarized in Figure 5.6 (Chapter 5). When using $\delta^{18}O$ a more erratic proportion of old water is observed on a monthly scale, compared to Cl^-. The more sensible cases occurred in the Loechernbach catchments, in particular in spring and autumn seasons (even negative values occurred for September 1995). This is not surprising because of higher direct runoff in that area. The erratic behavior of $\delta^{18}O$ in spring and autumn is due to $\delta^{18}O$ concentrations between precipitation and groundwater being too close to each other, so that the corresponding monthly separation turns out to be problematic. In contrast to $\delta^{18}O$, the results obtained by Cl^- are consistent and reliable.

In principle, Cl^- can be used as an alternative tracer for the study basins. The chemical tracers could prevent the inaccurate measurement of $\delta^{18}O$ in soilwater and are particularly valuable when $\delta^{18}O$ of rainwater is too variable or too close to the $\delta^{18}O$ of the streamwater. It is noted that in spite of the disadvantage of $\delta^{18}O$ for the separation in spring and autumn, the results obtained by $\delta^{18}O$ for the average case of entire summer-and winter-half year are reliable. This is because stable isotope as ideal conservative tracer is indispensable in the study of runoff separation. Also, the average $\delta^{18}O$ of precipitation for the summer- and winter-half year markedly differs from that of groundwater.

In the present study, $\delta^{18}O$ and Cl^- allow for a two-component separation respectively. However, through the simultaneous measurement of $\delta^{18}O$ and Cl^-, it is possible to obtain further partitioning of stream than that afforded by the use of either tracer individually.

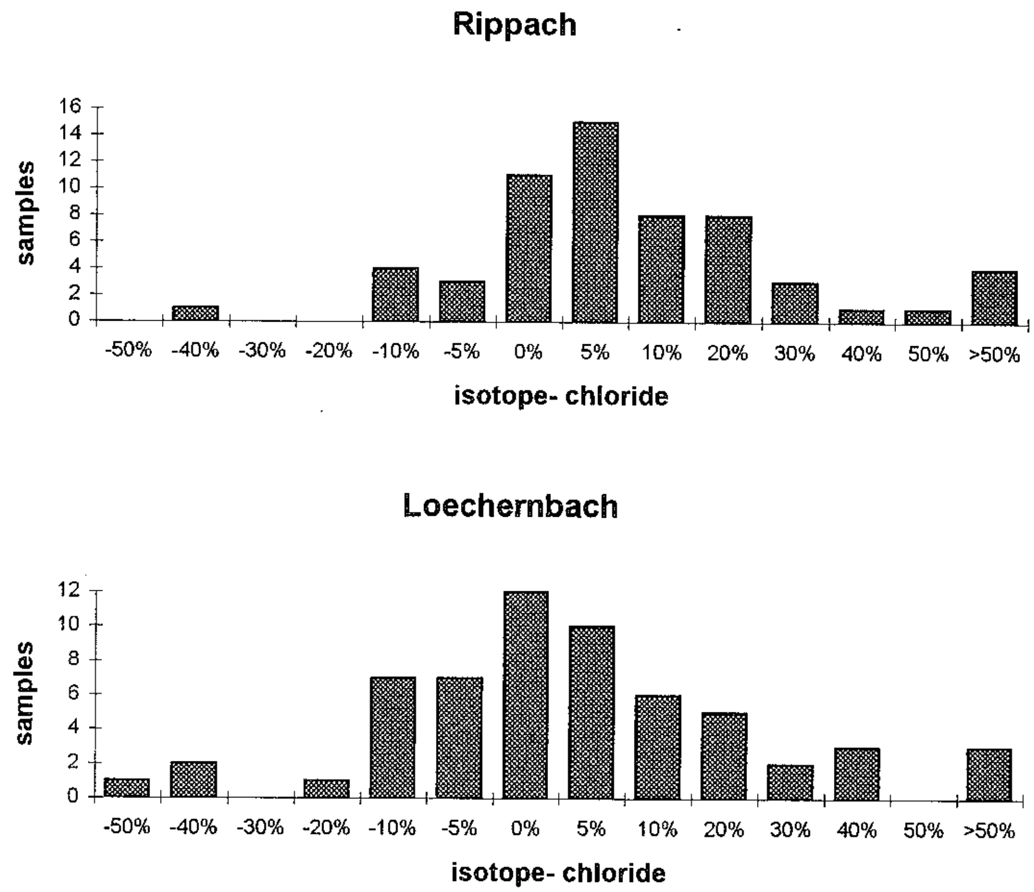


Fig. 7.1 Histogram of different old water proportion of total flow calculated by chloride and oxygen-18

The comparison between $\delta^{18}O$ and Cl^- indicates that old water fractions using chemical tracers are 8 % higher than those using $\delta^{18}O$ in the Loechernbach catchment (Tab. 7.3), while they were only 1% higher in the Rippach catchment. The present question is: where did this 8 % higher old water when using Cl^- originate from?

A most likely explanation of this difference of old water fractions between $\delta^{18}O$ and Cl^- is a conceptualization of partitioning. New water and old water components (using $\delta^{18}O$) can

Tab. 7.3 Comparison of yearly average old water contributions to the streams by different methods

Catchment	$\delta^{18}O$	Cl^-
Rippach	91%	92%
Loechernbach	83%	91%

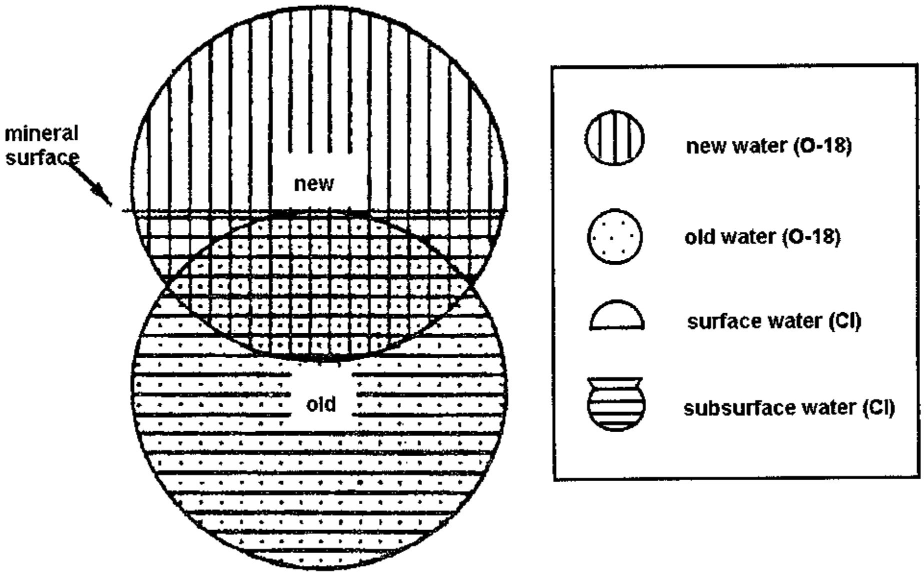


Fig. 7.2 Partitioning of waters according to the use of $\delta^{18}O$ and Cl^- (MAULÈ & STEIN 1990, modified)

be represented by two boxes that are connected to each other. To minimize confusion of terminology within this study, the definitions mentioned in earlier chapters have been utilized. Thus, using chemical tracers such as Cl^-, the surface and subsurface bisect, a horizontal line separates the new water component.

A Venn diagram is used to explain this concept (Fig. 7.2). This concept was presented by MAULÈ & STEIN (1990). They used silica and $\delta^{18}O$ to separate so-called recent subsurface flow (Fig. 7.2). New water is referred to by surface water and rapid interflow, old water is referred to by slow interflow (vadose zone water) and base flow. The amounts of new and old water are determined by $\delta^{18}O$. The rates of surface water and subsurface water (interflow + baseflow) may be determined by Cl^-. Here surface water is assumed to be very low in Cl^- concentration (similar to precipitation or somewhat higher considering the possible dust or fertilizer on the sealed areas), rapid interflow is assumed to be only new water. Old water can only be subsurface water, but subsurface water can include both new and old water (MAULÈ & STEIN 1990). On the other hand, surface water is always less than new water. According to this concept, the following results can be calculated:

rapid interflow = the difference of new water fractions using Cl^- and $\delta^{18}O$

i.e., rapid interflow = 92 % - 91 % = 1 % at Rippach

rapid interflow = 91 % - 83 % = 8 % at Loechernbach

It is noted that this calculation is only a rough estimate provided that certain conditions are fulfilled. The difficulty of this method is the interflow. In reality, rapid interflow is difficult to determine. In this study, rapid interflow can be understood as infiltrated water which continuously flows via macropores into a depth of soil up to 1.5 m, and is then transported by drainage pipes into the stream. The total course occurs within the order of minutes up to several hours. This explanation is logical, because the 20 km of installed drainage subpipes result in a marked contribution of this component, as cited in Chapter 6.

Macropore flow is dependent upon the antecedent precipitation and other factors. The average rapid interflow referred to here is only an average case of the year, including different rain characteristics and antecedent conditions.

Slow interflow can be understood as macropore flow through A and B horizons. This horizon movement of flow is very slow. The component is assumed to have the same $\delta^{18}O$ as groundwater. This proved to be reliable in the study basins. However, the case for Cl^- may be different. Chemical separation is based on the same assumptions as isotope separation: vadose water (here: interflow) has the same isotopic or solute content as "old" groundwater for the average case of the long-term observation period. In fact, this condition can not be entirely fulfilled. Cl^- contents of interflow could become lower if there was inadequate contact with the mineral soil or inadequate time for equilibrium. On the other hand, Cl^- contents of interflow could be higher if groundwater of higher Cl^- contents discharges into the upper layer. To reduce the probable error of the estimate, the current Cl^- contents in subsurface water is considered as changing, as presented in Chapter 2.

To explain the difference of runoff separation by using $\delta^{18}O$ and Cl^-, an interesting comparison of the sensitivity between the two tracers ($\delta^{18}O$ and Cl^-) is shown in Fig. 7.3. The data depict the summer separation of the Loechernbach. Giving a series of assumed Cl^- new water contents and a constant old water value, the corresponding new water fractions become a curve, according to Equation 2.13 (Chapter 2). Given a constant new water value and varying old water contents, another curve is given for the corresponding old water fractions. Using the same method, two curves for $\delta^{18}O$ are also given.

Figure 7.3 displays the following facts: the increase of Cl^- content from precipitation to new water does not cause a large increase of the percentage of new water. In contrast, the new water percentage is more sensitive to the change of $\delta^{18}O$ in new water. On the other hand, both Cl^- and $\delta^{18}O$ are sensitive to any change of the tracer concentrations in old water. This applied to the long-term period in the study catchments. These facts indicate following two points: First, any change of the Cl^- content in surface and subsurface flow

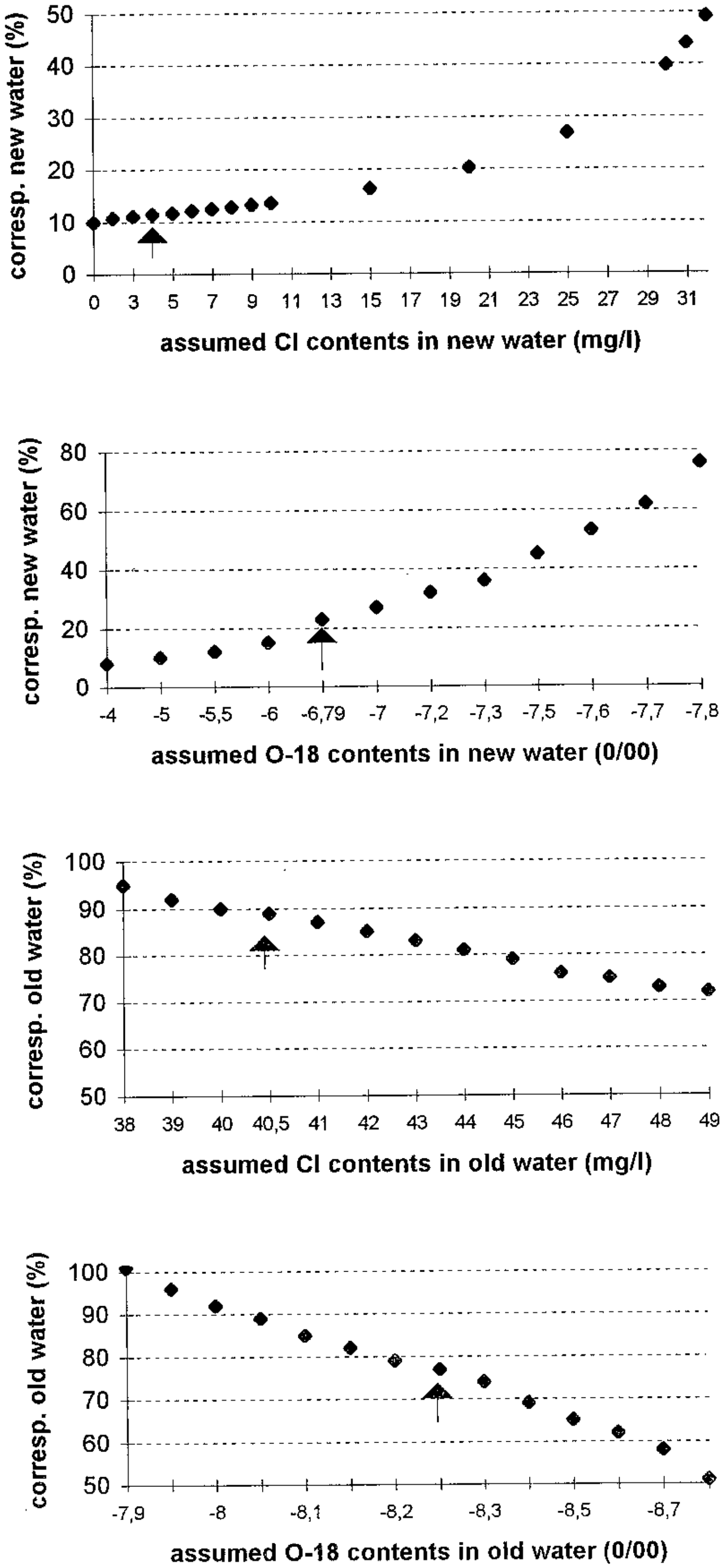


Fig. 7.3 **Comparison of sensitivity of chloride and oxygen-18**
(↑: measured values)

influenced the results of two-component separation only minimally. Second, it is important to consider old water as a varying sense, rather than in a constant sense. When this variation is neglected, a large error will result.

In fact, chemical tracers are easily altered in soil, so that the assumption of the two-component separation for the same chemical contents cannot be fulfilled in most cases. However, variability in isotope concentration in precipitation also reduces the accuracy of the separation. Therefore it is suggested that both isotope and chemical tracers are analyzed simultaneously.

It has to be noted that the rapid interflow results combined by $\delta^{18}O$ with Cl^- dealt with above are not indisputable from the hydrological aspect, though the computed percentages of rapid interflow are identical in order of magnitude with other catchment information and the results of the models. A further investigation is necessary to obtain accurate values of rapid interflow.

7.2.2 Comparison of the short-term results obtained by oxygen-18 and hydrochemical tracers

The comparison between isotope and chemical tracers for individual rainfall events is performed in Chapter 6.1 and Chapter 6.4. The results obtained by $\delta^{18}O$ proved to be critical when comparing them to the results obtained by Cl^- for the storm event in the Loechernbach catchment. For the Rippach event in September 1995, $\delta^{18}O$ had limited use for the separation because precipitation in the latter phase of the event was too close to that of groundwater.

7.2.3 Comparison between long-term and short-term results

Previous studies of runoff generation indicated that water stored from previous rainfalls volumetrically dominate the streamwater response to storm rainfall (e.g. FRITZ et al. 1976, SKLASH et al. 1979, RODHE 1981, HOOPER & SHOEMAKER 1986, HERR-MANN et al. 1986, MCDONNELL et al. 1990, OGUNKOYA & JENKINMS 1993). Indeed, the long-term results in either Rippach or Loechernbach catchments provide a consistent picture of dominant contribution of old water. However, for the short-term rainfall events, dominant event water fractions are observed in stream during the major phase of both short-term events. This suggests that event water may be dominant during individual events in the intensively agricultural use areas. The altered Loechernbach catchment is an example of this inconsistent picture between different time scales. On the one hand, the dominant old water contribution in the yearly scale reflects similar bedrock as in Rippach basin. On the other hand, event water dominated the streamwater, and this signal remained for a few days. Here the significant macropore flow via amounts of subsurface pipes are quite important.

Long-term and short-term separations are two different concepts. The former explain a statistical average case, in which old water relates to the weighted average water that exists previously in the catchment (or average pre-event water). In contrast to this, the latter has a clear physical explanation, here pre-event water differs from event water in most cases.

The change of solute concentrations in pre-event water is negligible during the individual events compared to the long-term period. It is therefore concluded that the estimation of pre-event water contribution based two-component model for individual events is more reliable than that for long-term calculation.

The comparison of the sensitivity analysis displays that chemical tracers are more sensitive to the increase of event water concentrations for short-term events compared with long-term period. The increase of Cl^- in new water from 2.2 mg/l to 10 mg/l results in only 3 % fluctuation for the long-term period, while 22 % fluctuation for the Loechernbach storm event and 19 % for the Rippach storm event were calculated (Tab. 6.4). This is because solute concentrations in stream water during individual events, in particular during peak events, are lower so that the difference of the solute concentrations between stream and event water is decreased.

7.3 Comparison of the components results obtained by the DIFGA model with the experimental results

A comparison of the results obtained by the application of the DIFGA model and tracer results for runoff concentrations is described in this section.

As mentioned above, the DIFGA model belongs to the traditional graphical methods and its application is simplified through the use of computers. This method does not need any parameter exception of P and Q, and is applied to a rough estimation of the groundwater contribution for a basin size of under 300 km^2. Therefore, this method is practical and easy to use. For regions with simple hydrological behavior the error is limited, whereas for a basin with large direct runoff the accuracy of the estimate will decrease.

The separation of rapid baseflow and slow baseflow with DIFGA is reliable for many cases, if storage coefficients *CG1* and *CG2* can be determined under considering long-time series data and groundwater data, as well as other information of the basins. The experimental results from isotope or hydrochemical tracers would be, of course, helpful to determine these storage coefficients.

However, the experiments using natural tracers confirmed that the mathematical models used in the study sites underestimate old water contributions to streamflow. Fig. 7.4 demonstrates that the underestimates of baseflow fractions with DIFGA are about 7 % at Rippach and 19 % at Loechernbach. The difference of 7 % in the Rippach catchment is acceptable. However, why do both methods significantly differ in the Loechernbach catchment (the underestimate was up to 27 % in summer)? Clearly, this deviation cannot be explained by the calculated error alone.

This underestimate of the modeling results is not surprising, because the graphical method demonstrates a sharp separation into direct runoff and indirect runoff. Previous studies by EDEN et al. (1982) and HERRMANN & STICHLER (1980) came to similar conclusions.

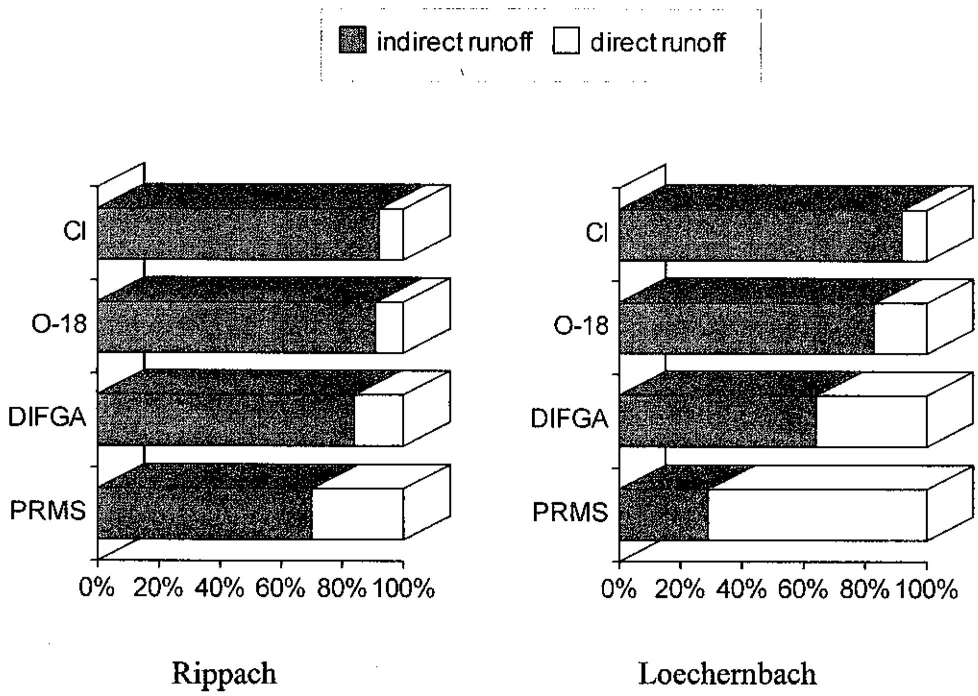


Fig. 7.4 Comparison of the components separations by different methods for the hydrological year 1995

The other reasons that may have caused underestimation of groundwater using the DIFGA model are summarized below:

(1) In the DIFGA model the soil reservoir is assumed to work only after the end of the surface water linear-storage. Correspondingly, the groundwater reservoir works only after the end of the soil water linear-storage. In fact, the groundwater contributes to stream even before the peak discharge, as proven in the field investigations. Furthermore, event water can be still a dominant contribution to the streamflow after the rainfall events. This is obvious in the storm event on July in Loechernbach. Therefore, the simplified assumption by DIFGA may result in some error. In complex catchments, the non-linearities inherent in the loss function would result in difficulties of simulation.

(2) The DIFGA model deals with an old concept of graphical separation. It allows for the separation for persistent events based on the estimate of the average recession coefficients. However, the connection line of the rising limb of the flood wave is more and less a speculation (DRACOS 1980, WETZEL1994). Therefore, the graphical separation is of uncertain nature.

(3) The separation between rapid direct runoff and slow direct runoff in the DIFGA model is problematic. The time scale of direct runoff is normally only a few hours.

(4) DIFGA model uses an error correction concept based on a monthly water balance. In the loess region soil storage capability is very large, so the water can be stored for a long time before entering the stream. Therefore, runoff separation based on the monthly balance will result in significant error.

(5) DIFGA only considers discharge and precipitation and neglects hydrological processes and catchment information, i.e. geology, morphology, or soil type of the basin. Therefore, the DIFGA model is only a black-box model based on water balance. Thereby DIFGA cannot solve the problem of determining the storage constants such as *CG2* and *CG1*.

In fact, not only the DIFGA model, but also the other applied models, have underestimated baseflow fraction in the altered basin. This is because hydrological models are extreme simplification of the reality, their application to complex basins is questionable. In contrast to this, tracer methods may overcome the disadvantages of models which do not consider natural runoff processes and allow for the objective and direct review of runoff processes. The runoff components are related to the physical processes instead of rough separation. Therefore the runoff components obtained by tracers are more reliable than those obtained by DIFGA in most cases.

As noted above (Chapter 7.1), the PRMS model, applied by BLAU (1996), also shows a overestimate of surface runoff percentage in the Loechernbach. After linking with the present isotope results, simulation of the annual hydrographs has been markedly improved. This indicates a possibility to link the tracers results with model results.

Of course, tracer methods are reliable with conditions. The two-component model applied in the experimental studies is just a simplification of the catchment process. In Equation 2.14 (Chapter 2) all new water reaching the stream is assumed to be originated from rainfall on saturated areas and asphalt roads. The isotopic contents of new water from asphalt roads usually are not changed along its path to the stream in small catchments. However, if overland flow passes the saturated areas on its way to the stream, the isotopic content may change in direction towards that of the groundwater (RODHE 1981). For that, groundwater can be overestimated. This also applied to hydrochemical tracers.

The boundary conditions of the DIFGA and tracer method are different. The former focuses only on temporal distribution of runoff processes, the latter focuses on both temporal and spatial distribution, in particular the spatial distribution of various soil zone in depth.

7.4 Comparison of the transit times determined by the DIFGA model and by the experimental investigations

The average transit times of each runoff component have been investigated using DIFGA, $\delta^{18}O$ data, and by tracing tests.

The storage constant given by DIFGA model for the slowest component is 495 days at Rippach, 74 days shorter than the corresponding residence time from $\delta^{18}O$, Their difference is acceptable (Tab. 7.4). However, a large difference between storage constant and residence time is given in the altered catchment. According to the result of $\delta^{18}O$, this value is at 639 days, a value of about 4 times longer than 160 days obtained by DIFGA. Groundwater reservoir volumes estimated by $\delta^{18}O$ and by conventional method (analysis of streamflow recession) are essentially different. Using $\delta^{18}O$, total mobile groundwater reservoir is estimated, whereas estimates based on DIFGA relate to active groundwater reservoir. This is discussed in next chapter in details.

Tab. 7.4 The comparison of the residence times obtained by $\delta^{18}O$ and DIFGA

catchment	component	DIFGA	artificial tracers	$\delta^{18}O$
Rippach	surface runoff	≅0	20 minutes	≅0
	rapid interflow	-	?	-
	slow interflow	12 days	-	624 days
	baseflow	550 days		
Loechernbach	surface runoff	≅0	20 minutes	≅0
	rapid interflow	3 days	2.5 hours	-
	slow interflow	9 days	-	639 days
	baseflow	160 days		

The interpretation of $\delta^{18}O$ data by mathematical flow models seems to be relatively accurate. Isotopic input is field well-distributed. The average transit times of the study sites are determined with streamwater data, the transit times of old water are computed with baseflow data. As far as fast components are concerned, the transit times obtained by the tracing experiments are essentially shorter than those obtained by the DIFGA model. The test fields are chosen half-way down the stream, but no in situ soil development exists in the undrained fields of the altered basin and thus, overland flow occurs easily after rainstorms. It is therefore difficult to use points input data from artificial tracer experiments to test storage times obtained by DIFGA.

7.5 Characteristics of runoff generation in the study sites

This study have indicated that the Rippach and the Loechernbach concern different mechanism of the hydrological behavior. While overland flow is the dominant origin of event water in the Rippach, rapid interflow is of equal importance in the altered Loechernbach.

In the Rippach catchment, groundwater dominates the stream throughout the year, so the runoff behavior to this region is relatively simple. Different models or tracers give the similar results. Concerning abundant pre-event water stored in the soil, this is due to the very high retention capacity of thick loessian soils. MORGENSCHWEIS (1984) indicates that a 2-3 m deep loess profile can store up to 700 mm which basically is mean annual precipitation depth of the area. The soil column therefore stores up to 99 % of the fallen precipitation. The valley site of the Rippach catchment demonstrates high capillary water due to higher groundwater table, which in turn increases water escapement through soil evaporation. Evaporation to this region may be even higher than total potential evaporation (MORGENSCHWEIS 1984). It is concluded that dominant precipitation infiltrates in the soil, and pre-event water is pushed out through the displacement effect during storm events. Most event water (overland flow) appearing in the stream is from the valley area and the asphalt roads.

The calculation based on isotope data indicates that infiltration rate of Rippach catchment is balanced throughout the year. One of the reasons for that is high crop cover in this basin (33 % agricultural area and 18 % forest) which promotes infiltration at the higher rates for longer periods by slowing down the rate of runoff on account of the greater resistance to surface flow.

The runoff behavior of the altered Loechernbach catchment is somewhat complicated. The drainage pipes to this region are an important factor influencing the hydrological processes. Except for the 6 % asphalt roads and solid surface, these pipes enable precipitation to enter the stream very rapidly through macropore flow, as proven in the individual event in July of 1995. This also confirms the finding of HILL et al. (1993) for dominant new water during the flood period. On the other hand, these pipes facilitate old water stored in the unsaturated zone entering the stream during the normally low rainfall or precipitation free period. These characteristics may be explained with Figure 7.5 appeared in BAUMGARTNER & LIEBSCHER (1990). This explains why the contribution of old water is important for long-term period in this basin and why the contribution of event water is important for short-term period. The small but significant contribution of pre-event water before the peak of an individual event shows the influence of the piston-flow effect.

The significant macropore flow is mainly the result of the agricultural activities. The pores formed by plant roots may be very effective in channeling water through the soil, even through unsaturated soils depending on the plant species and the conditions of plant growth (MOSLEY 1979). OMOTI & WILD (1979) have shown that such channeling or bypassing may also take place with capillary sized pores. These results strongly suggest that the Darcian principles may not adequately describe the infiltration and redistribution of water where the soil contains macropores. The pipeflow is in some areas also important, especially during a drought period, as proven in the July event. Drought will cause clay soil to crack, sometimes to a considerable depth, and may cause animals living in the soil to dig deeper in order to maintain a favorable environment for them (BEVEN & GERMANN 1982).

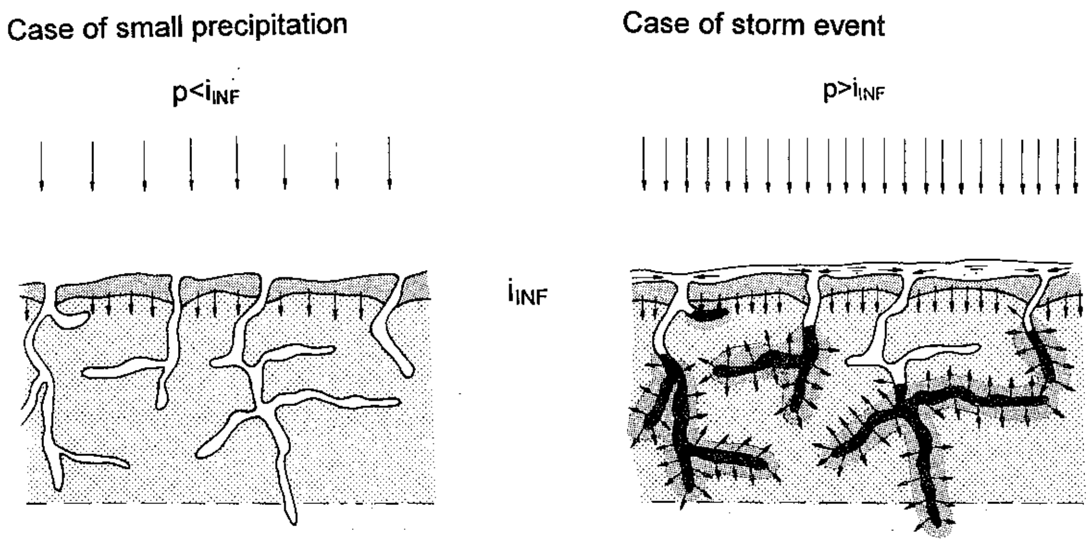


Fig. 7.5 Concept diagram of the influence of macropore flow (BAUMGARTNER & LIEBSCHER 1990)
P: precipitation; i_{INF}: infiltration rate

The density of some upper soil layers tends to increase and to destroy macropores through the use of heavy machinery in the altered Loechernbach catchment. This is also a reason why a number of subpipes in this basin are installed. Therefore, macropore flow contributing to the stream in the altered basin may be derived from a part of the total area. It is noted that macropores are similarly abundant in the Rippach basin, the bypass flow there, however, results in the deep infiltration of precipitation, because in this basin drainage pipes are absent, so the contribution of the macropore flow to runoff is not obvious.

The finding of the importance of macropores partly explained why in the present loess basins infiltration rates are especially large. The presence of macropores in the upper soil layer serves to increase infiltration rates because additional surfaces are made available for infiltration into the matrix at depth.

8 Concept model of the hydrological system in the study site

8.1 Introduction

The results obtained by different methods are combined in this chapter, to develop a concept model of the hydrological system in the study basins.

The attempt to introduce the isotope results into the DIFGA model is rejected. This is because:

(1) The DIFGA model deals with the temporal course of runoff processes, and does not concern where the waters are originating from. Isotopes and chemical tracers deal with not only temporal, but also spatial course of runoff processes. They have different objectives and different boundary conditions. For that, any introduction of the isotope results into DIFGA, such as the correctness of the storage constant for slow baseflow component, will raise the conceptual confusion.

(2) Storage constants of the DIFGA model are different from residence times obtained by isotope. The former concerns how runoff enters the stream, with the help of the recession analysis; the latter concerns how long infiltrated water is stored in the basin.

(3) The corresponding storage volumes between DIFGA and isotope data differ. While the former concerns only active groundwater, the latter relates to total mobile water reservoir. In general, the isotope reservoir is larger than the reservoir determined by recession analysis.

8.2 Determination of runoff components in combination of different methods

Direct runoff

The direct runoff (Q_d) is made by multiplying discharge and the new water fraction calculated using $\delta^{18}O$ data:

Rippach: $Q_d/Q_t = 8\%$, $Q_t = 115.90$ mm, $Q_d = 9.28$ mm (8.1)

Loechernbach: $Q_d/Q_t = 17\%$, $Q_t = 169.70$ mm, $Q_d = 28.85$ mm (8.2)

Here Q_t is the total discharge; Q_d is separated into two components: surface runoff Q_{d1} and rapid interflow Q_{d2}, which are determined by the difference of the new water fractions between Cl^- and $\delta^{18}O$ (Chapter 7.2).

Rippach: $Q_{d1}/Q_t = 7\%$, $Q_{d1} = 8.12$ mm (8.3)

$Q_{d2}/Q_t = 1\%$, $Q_{d2} = 1.16$ mm (8.4)

Loechernbach: $Q_{d1}/Q_t = 8\%$, $Q_{d1} = 13.58$ mm (8.5)

$Q_{d2}/Q_t = 9\%$, $Q_{d2} = 15.27$ mm (8.6)

Indirect runoff

Similarly, the indirect runoff (Q_g) is made by multiplying discharge and the old water fraction calculated by $\delta^{18}O$ data:

Rippach: $Q_g/Q_t = 92\%$, $Q_g = 106.72$ mm (8.7)

Q_{g+s}= 106.72 mm + 20.00 mm (winter spring) = 126.72 mm (8.8)

Loechernbach: $Q_g/Q_t = 83\%$, $Q_g = 140.85$ mm (8.9)

Here Q_{g+s} at Rippach is the sum total of Q_g and drinking water Q_s, which contributes to the water supply in winter months.

To separate Q_g into rapid baseflow (Q_{g1}) and slow baseflow (Q_{g2}), the ratio of rapid to slow baseflows is determined by the DIFGA data:

Rippach: $Q_{g1}/Q_g = 27\%/(27\%+57\%) = 32\%$, $Q_{g1} = 34.30$ mm (8.10)

$Q_{g2}/Q_g = 57\%/(27\%+57\%) = 68\%$, $Q_{g2} = 72.42$ mm (8.11)

Q_{g2+s}= 72.42 mm + 20.00 mm (winter spring) = 92.42 mm (8.12)

Loechernbach: $Q_{g1}/Q_g = 52\%/(12\%+52\%) = 81\%$, $Q_{g1} = 114.04$ mm (8.13)

$Q_{g2}/Q_g = 12\%/(12\%+52\%) = 19\%$, $Q_{g2} = 26.41$ mm (8.14)

The results dealt with above are summarized in Table 8.1.

8.3 Residence times in combination with different methods

As mentioned above, the storage constants obtained by DIFGA differ from the residence times of the natural system. Interpreted by the dispersion model DM, the mean residence times of the entire system (T_t) and times of baseflow (T_g) are from the average $\delta^{18}O$ data in stream water and baseflow respectively.

Under the assumption of the equivalent isotope content in groundwater and soil water for the two-components model (Equation 2.3), old water is the sum of rapid baseflow and slow baseflow. The residence times of slow baseflow (T_{g2}) are assumed to be the weighted average times of some groundwater measuring points determined by the simple procedure of the dispersion model with D/vx = 0.02 (Tab. 5.5). Rippach as an example, T_{g2} is considered to be the weighted average times of groundwater GR, drainage water DR6 and DR8. GR, GR6, and GR8 represent the average case of central valley area (11 % of the basin area), of marginal area (28 % of the basin area), and of high terracing area (60 % of the basin area), respectively. Similarly, T_{g2} in the Loechernbach is based on the data from groundwater GL, drainage water DL4 and DL7.

$$\text{Rippach:} \quad T_{g2} \approx (11\% \cdot T_{GR} + 28\% \cdot T_{DR6} + 60\% \cdot T_{DR8}) / 99\%$$

$$= (11\% \cdot 1.63 \text{ yr.} + 28\% \cdot 1.9 \text{ yr.} + 60\% \cdot 2.1 \text{ yr.}) / 99\%$$

$$= 1.99 \text{ years} \qquad (8.15)$$

$$\text{Loechernbach:} \quad T_{g2} \approx (20\% \cdot T_{GL} + 12\% \cdot T_{DL4} + 62\% \cdot T_{DL7}) / 94\%$$

$$= (20\% \cdot 1.42 \text{ yr.} + 12\% \cdot 2.0 \text{ yr.} + 62\% \cdot 2.0 \text{ yr.}) / 94\%$$

$$= 1.88 \text{ years} \qquad (8.16)$$

Considering the minimal variability range of the residence times in these measuring points (between 1.7 years and 2.0 years, Tab. 5.5), the assumption of these sampling representative of slow baseflow should be acceptable. The reason for D/vx = 0.02 is that the times with D/vx = 0.02 using simple procedure of the dispersion model resulted in values closest to the simulation using dispersion model (Chapter 5.5).

The residence times of rapid baseflow (T_{g1}) are made by the storage volumes (V_{g1}) divided by the corresponding discharge Q_g (Equation 2.2). V_{g1} is the difference between V_g and V_{g2}. Accordingly, 0.96 years of T_{g1} at Rippach and 1.72 years are given. These residence times are in a yearly order, while the corresponding times set by DIFGA are in a daily scale (8-10 days). This emphasizes that simple comparison of the model times with isotope results is not possible.

The reason for the huge difference between isotope and model times for rapid baseflow component is most likely that most unsaturated soil water or near-stream groundwater, which contributes to the rapid baseflow component, is pre-event water stored before in the catchment. Capillary rise in the valley areas of the study sites should play an important role for that. Therefore, the water related to this component displays double characteristics. On the one hand, the water responses the rainfall events for only a few days, mirroring the characteristic of runoff recession; on the other hand, the water in this zone has been stored before in the basins and occurs along with the feature of groundwater. This cannot be dealt with using a single approach, so the integration of model and tracer approaches is helpful for the better understanding of runoff generation in this zone.

The mean residence times of the direct runoff components were directly determined from tracing tests. The surface runoff is assumed to be 20 minutes (T_{d1}), according to the result of the tracing test on the asphalt road (approximately weighted point of the Loechernbach catchment) in July 1995 (Chapter 6.1.4). The 2.5 hours of residence times for slow direct runoff (T_d) is based on the data of the tracing test in an unsaturated zone at Loechernbach (Chapter 6.3). Actually, the contribution of this component is believed to be negligible in the Rippach because of the absence of drainage subpipes to this region.

Rippach: $T_{d1} \approx$ 20 minutes (8.17)

Loechernbach: $T_{d1} \approx$ 20 minutes (8.18)

$T_{d2} \approx$ 2.5 hours (8.19)

All results of the residence times are given in Table 8.1.

8.4 Storage volumes

8.4.1 Determination of storage volumes

According to the computed mean residence times of 1.60 years / 1.46 years using $\delta^{18}O$ for both catchment Rippach and Loechernbach, the mean discharge of the total mobile water volumes can be estimated to be $2.23\times10^5 m^3$ and $4.21\times10^5 m^3$, respectively. The corresponding groundwater storages of the catchment systems are $2.19\times10^5 m^3$ and $4.19\times10^5 m^3$, which are equivalent to water depths over the catchment area of 182 mm plus 35 mm storage supplied as drinking water at Rippach, and 247 mm at Loechernbach respectively.

The slow baseflow reservoirs are calculated by multiplying Q_{g2} and T_{g2}. For Rippach, the discharge of slow baseflow include the average of 20 mm drinking water depletioin during the winter months. The volumes of rapid baseflow (V_{g1}) are computed by the difference between total baseflow volumes (V_g) and slow baseflow volumes (V_{g2}). The computation procedure of these values are listed in Table 8.1. The reservoirs of rapid interflow are sufficiently small ($0.1\text{-}0.3\times10^5 m^3$) that they can be neglected.

8.4.2 Discussion of the results

According to tracing tests, LUFT & MORGENSCHWEIS (1982) computed the effective porosity of the deposits to be 0.2 in the study sites. Assuming this porosity can be represented as the average value of the entire basin, the resulting average depth of the total groundwater reservoir would be:

$$H\,(\text{Rippach}) = H\,(\text{groundwater}) / \text{effective porosity}$$

$$= (0.182\text{ m} + \text{drinking water } 0.035\text{ m}) / 0.20 = 1.085\text{ m} \quad (8.20)$$

$$H\,(\text{Loechernbach}) = 0.247(\text{m}) / 0.20 = 1.235\text{ m} \quad (8.21)$$

where H is the average depth of total groundwater reservoir.

This means that the average depth of total groundwater reservoirs would be 1085 mm and 1235 mm respectively. The annual average discharge from the basins is about 15 % of the total precipitation (annual precipitation 700 mm). 85 % of the annual precipitation are evaporated. According to the soilwater measurement (MORGENSCHWEIS 1984), a loess profile of a depth of 2-3 m in the Kaiserstuhl could store up to 700 mm of precipitation water, which represents the average annual volume of precipitation of the area. Hence, the computed groundwater reservoirs in both catchments are lower than what would have be expected from the site characteristics.

The reason for underestimating the groundwater reservoir is not known. Similar under-estimates of such groundwater reservoirs also appeared in other studies, e.g., from MALOSZEWSKI et al. (1992), in which water volume computed from isotope data was only about half that estimated from hydrological data.

A possible explanation for the small volume of the reservoir could be the presence of a large immobile water reservoir. In other words, the mobile water could be only a part of the total groundwater reservoir. Using $\delta^{18}O$ in the study basins, only mobile and a part of the immobile water can be determined. In the study basins, the thick layer of loess allows for deep-lying groundwater, which moves minimally (quasi stagnant) and thereby contribute to the immobile water reservoir. Also, destruction of the soil through terracing in the altered basin has increased the immobile water reservoir. In addition, the presence of macropore in loess results in a double porosity media. In double porosity media, both mobile and stagnant water can exist simultaneously (HERRMANN et al. 1990).

According to HERRMANN et al. (1990), the residence time of the dispersion model DM is suitable for the case of porous aquifers. For double porosity, e.g. in the case of soils with macropores in or fissured rock aquifers with a porous matrix, diffusion of tracer in the matrix has to be considered as a second relevant process. In other words, the reservoir in this case includes mobile and stagnant water volume of a system. The corresponding T_t is larger than T_0 (MALOSZEWSKI et al. 1985). The double porous media of macropore flow and matrix flow have been confirmed in the tracing investigations in the unsaturated zone (UHLENBROOK 1995). Therefore, the residence times and corresponding water

Tab. 8.1 **Computation of the runoff values, residence times and storage volumes for each runoff component**

No.		components (mm/year)		calculation method
		Rippach	Loechernbach	
1	P	799.00	842.00	hydrometric data (precipitation)
2	Q_t	115.90	169.70	hydrometric data (runoff)
3	Q_d	9.28	28.85	isotope data (new water fraction)
4	Q_g	106.72	140.85	isotope data (old water fraction)
5	Q_{d2}	1.16	15.27	isotope data and chloride data
6	Q_{d1}	8.12	13.58	difference of Q_d and Q_{d2}
7	Q_{g2}	72.42	26.41	DIFGA data
8	Q_{g1}	34.30	114.44	DIFGA data
9	Q_s	20.00	0.00	estimate value, according to LUFT (1980)
10	$Q_{(g2+s)}$	92.42	26.41	sum of Q_{g2} and Q_s
11	$Q_{(t+s)}$	135.90	169.70	sum of Q_t and Q_s
12	T_t	1.60 yr.	1.46 yr.	isotope data (streamwater, using DM)
13	T_d	0.00 yr.	0.00 yr.	tracing test
14	T_g	1.71 yr.	1.75 yr.	isotopic data (baseflow, using DM)
15	T_{d2}	20 min.	20 min.	tracing test
16	T_{d1}	-	2.5 hrs	tracing test
17	T_{g2}	1.99 yr.	1.88 yr.	isotopic data (spring, using simplicated DM)
18	Tg1	0.96 yr.	1.72 yr.	calculated value, according to No. 8 and No. 25
19	V_t	217.60	247.76	calculated value, according to No. 2 and No. 12
20	V_d	0.00	0.00	difference of V_t and V_g
21	V_g	182.49	246.49	calculated value, according to No. 4 and No. 14
22	V_{d2}	0.00	0.00	difference of V_d and V_{d1}
23	V_{d1}	0.00	0.00	calculated value, according to No. 6 and No. 16
24	V_{g2}	144.11	49.65	calculated value, according to No. 7 and No. 17
25	Vg1	32.95	196.84	difference of V_g and V_{g1}
26	$V_{(g+s)}$	216.69	246.49	calculated value, according to No. 4, 9 and No. 14
27	$V_{(g2+s)}$	183.91	49.45	calculated value, according to No. 10 and No. 17

volumes of groundwater obtained by $\delta^{18}O$ in the altered catchment may be larger than those occurred only in the simple porous aquifer.

MALOSZEWSKI (1996) suggested that tritium as tracer results in total groundwater reservoir (mobile and stagnant water reservoir). Stable isotope, however, usually does not include total, if any, immobile water reservoir. This is because stable isotopes signals cannot determine the transit time which are over 4 years. There has yet no clear limit for this concept in previous studies. In the present study basins, the water reservoirs calculated using $\delta^{18}O$ data can be understood as active water reservoir, which contributes to the streams, in which insignificant, if any, contribution of immobile water may be concluded.

Besides for the above reasons, the effective porosity of 0.2 determined by LUFT & MORGENSCHWEIS (1982) may be overestimated. This value applies to the central valley region with clayey to weak clayey silt farm. On high terracing land, however, such as the Mahlkuenzig station where the groundwater table is located 20 m deep below the surface (MORGENSCHWEIS 1984), the effective porosity could be significantly lower.

8.4.3 Comparison of the storage volumes obtained from DIFGA and isotope

A comparison of the storage volumes obtained from DIFGA and $\delta^{18}O$ is shown in Tab. 8.1. For the convenience of comparison, the spring water of 20 mm/year in the Rippach catchment has not been included in the calculation. Accordingly, The total groundwater storages of the Rippach catchment system was $2.23\times10^5 m^3$ by $\delta^{18}O$ and $1.40\times10^5 m^3$ by DIFGA. This difference is limited.

However, a huge difference between both storage volumes in the Loechernbach has been observed (Tab. 8.2). Using $\delta^{18}O$, a similar storage volume as for Rippach is calculated (247 mm), while the total volume as according to DIFGA is only 8.8 mm, and thus lower than that by using $\delta^{18}O$ by a magnitude of two orders. The negligible storage volumes from DIFGA can be explained as the result of a small Q_{g2} and T_{g1}. On one hand, when the slow baseflow at Loechernbach is only a small part (8%) of the total runoff, the small Q_{g2} results in a small storage volume ($Q_{g2} \cdot C_{g2}$) by DIFGA. On the other hand, the large proportional contribution of Q_{g1} with a small storage coefficient C_{g1} (10 days) does also results in only a small storage volume of this component.

8.4.4 Concept of different storage volumes

To explain the difference in the reservoir volumes according to the model and the experimental results in Table 8.2, a concept of energy reservoir is introduced. This concept is introduced by BERGMANN et al. (1996). They considered the underground water as two distinctive volume marks (Fig. 8.1). The first mark relates to the total active reservoir (V_s). This reservoir is well-known, and has been discussed in Chapter 2 (Equation 2.21). Another mark is so-called energy reservoir (V_e). This reservoir is a hypothetical one, and is

Tab. 8.2 Subsurface water and groundwater storages of the catchment systems*

	total reservoir		groundwater reservoir	
	Rippach*	Loechernbach	Rippach*	Loechernbach
area (km^2)	1.20	1.70	1.20	1.70
mean residence time (years)	1.60	1.46	1,71	1,75
1995				
discharge (mm/year)	115.90	169.70	106.60	140.90
volumes by O-18 (mm)	182.49	247.76	182.49	246.58
volumes by O-18 ($10^5\ m^3/s$)	2.23	4.21	2.19	4.19
volumes by DIFGA (mm)	116.81	8.76	116.00	8.76
volumes by DIFGA ($10^5\ m^3/s$)	1.40	0.15	1.39	0.11
1977-1980 + 1989-1995				
discharge (mm/year)	118.40	152.40	108.93	126.49
volumes by O-18 (mm)	187.07	222.50	186.27	221.36
volumes by O-18 ($10^5\ m^3/s$)	2.24	3.78	2.23	3.76
volumes by DIFGA (mm)	89.39	12.54	88.42	11.04
volumes by DIFGA ($10^5\ m^3/s$)	1.07	0.21	1.06	0.19

* without including drinking water depletion (20 mm/year) in the Rippach catchment

assumed solely as a measure for the energy caused by the underground water movement. According to the Darcy law, this V_e has a linear relationship with discharge Q:

$$V_e = k \cdot Q \tag{8.22}$$

or

$$h_e = V \cdot q \tag{8.23}$$

$$Q = \frac{dV_e}{dt} = -k\frac{dQ}{dt} = Q_0 \cdot e^{\frac{t}{k}} \tag{8.24}$$

where h_e is depth of the energy reservoir; k is the storage constant, determined according to mean discharge and mean monthly low discharge data of different subareas (BERGMANN et al. 1996). The k can be compared with the value C in the DIFGA model (Equation 2.1).

BERGMANN et al. (1996) indicated that the total volume is, in reality, many times larger than that of energy storages (factors 10 - 75). This energy reservoir may represent the average storage volume from DIFGA model. With this theory, the difference of DIFGA and isotope results can be explained. In the Loechernbach catchment, the active reservoir volume is similar to that of the Rippach catchment, yet the energy reservoir in the Loechernbach catchment is quite small. According to the above calculation, the total volume is 30 times that of the energy storage volume.

There are various reasons for the large difference between the energy or the dynamic reservoir and the total reservoir in the altered catchment. On one hand, during the terracing the soil structure had been destroyed by compaction and through working with mechanical scrapers and graders. This reduced the hydraulic conductivity and enlarged ratio of capillary moisture. Consequently a reduced dynamic of soil water movement resulted (LUFT et al. 1983). As the result, the isotope content of the soil water is similar to that of the water in the subsequent layer. On the other hand, deep ploughing and mole drainage by using broken styrofoam in more than one-quarter of the large terraces allows for the rapid displacement of old water and for this old water entering the stream. As the result, the hydrograph reflects a rapid recession curve.

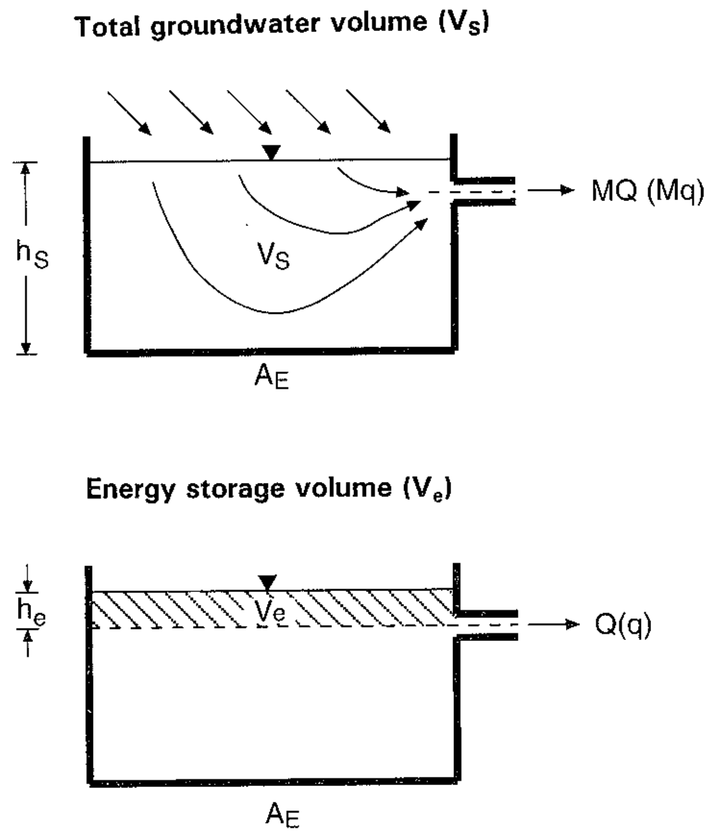


Fig. 8.1 **Concept diagram of total volume and energy storage volume** (BERGMANN et al. 1996)

(a) h_s: total storage height, MQ: mean discharge, Mq: mean runoff coefficient, A_e: site area

(b) h_e: energy storage height, Q: discharge, q: runoff coefficient

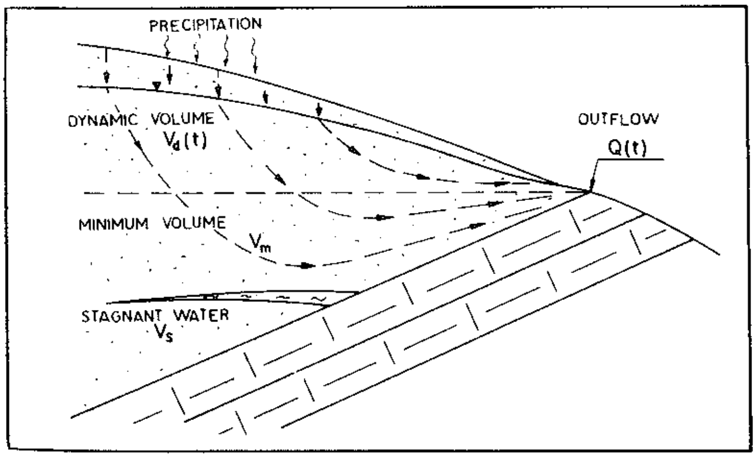


Fig. 8.2 Concept of dynamic volume, minimum volume and stagnant volume (ZUBER et al. 1986)

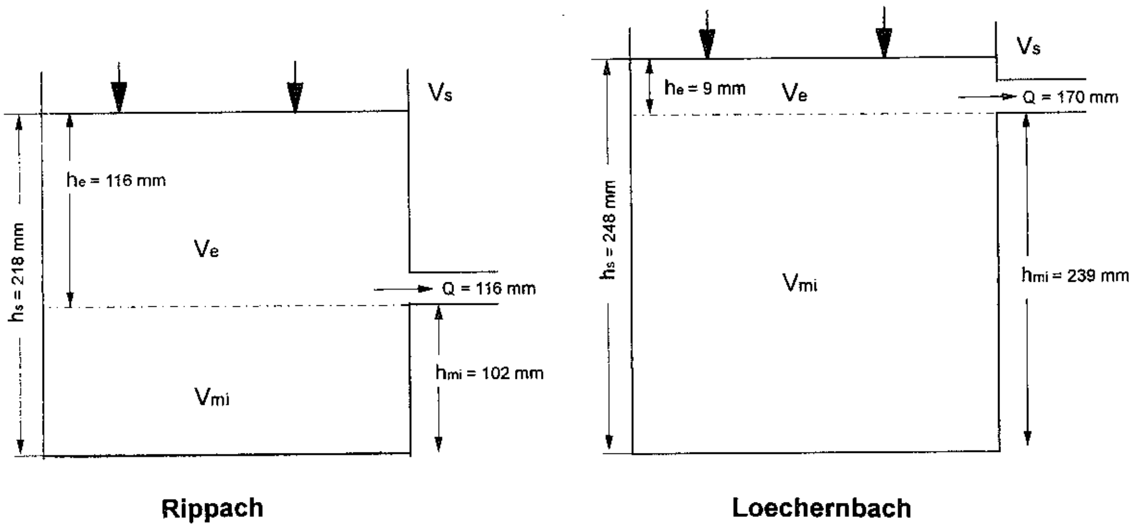


Fig. 8.3 Storage concept in the research sites

V_s total mobile volume determined by oxygen-18 data
h_s total storage depth
V_e energy (dynamic) volume determined by the DIFGA
h_e energy storage depth
V_{mi} minimum volume determined by the difference of V_s and V_e
h_{mi}: minimum storage depth
Q mean annual discharge

The capillary fringe effect, which limits storage and models rapid discharge of stored water, has been clearly demonstrated applying laboratory models (ABDUL & GILLHAM 1984, STAUFFER & DRACOS 1986). Some hydrologists suggested mathematical simulations which has, however, not really been reproduced in the field (e.g. MCDONNELL 1990). The present results in the altered Loechernbach support the presence of this capillary fringe effect as found in laboratory models. Due to drainage pipes, rapid discharge of stored water is allowed for displaying small energy reservoir values. However, this water mainly originates from the deep-lying reservoir - displaying characteristic of deep groundwater.

A physical explanation of this energy storage was not given by BERGMANN et al. (1996). The results from this study suggest that the energy may be from the lower layer of mobile water produced by capillary rise of groundwater. This is based on the fact of large difference of the residence times for rapid interflow between DIFGA and $\delta^{18}O$.

Energy reservoir V_e can also be compared with dynamic volume introduced by ZUBER (1986), who divided total reservoir volumes into three parts: dynamic volume which influences the outflow rate, minimum volume which influences the tracer movement but has no bearing on the outflow, and stagnant volume which practically is not accessible to tracer (Fig. 8.2). ZUBER (1986) indicated that the difference between the turnover time and the mean transit time of tracer is accounted for in the final formula. In this concept, minimum volume is similar as h_s-h_e in Fig. 8.1, or difference of total storage volume determined with $\delta^{18}O$ data and with the DIFGA model.

The combined storage volumes for the present study sites are illustrated in Fig. 8.3. The influence of the terracing on the hydrological processes is appreciable. Detailed mechanism of how the water flows under ground in this altered basin remains to be understood and demands further investigation.

8.5 Conceptual model of the study catchments

The principal results from the experiments using natural and artificial tracers as well as DIFGA are compiled in Table 8.2, where the inputs and outputs (mm/year) are from hydrometric data. The other parameters have been discussed above. The subsurface reservoirs are separated into four zones as shown in Figure 8.4. Here the underground runoff, approximately 5 mm/year (LUFT 1980), are neglected.

Basic findings, such as 92 % / 83 % groundwater fractions and corresponding residence times of 1.7 years for the groundwater system, are not singular but fully compatible with previous studies in this field. These results support the results of the previous investigations of over the past two decades in the study basins, and provide additional information about the study catchments, which are useful for the better understanding of the hydrological processes in the study sites.

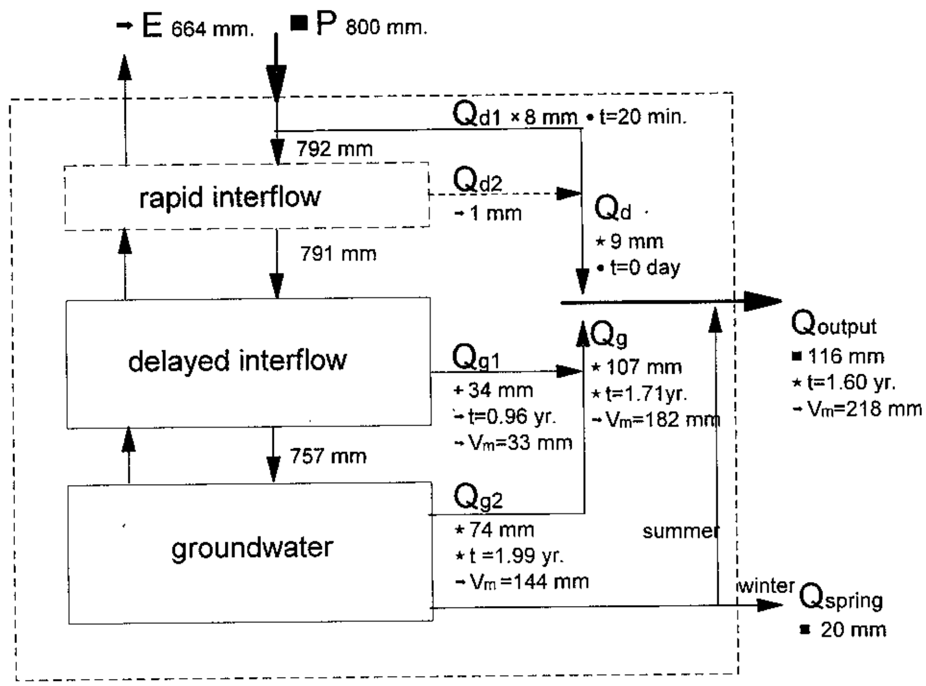
It is mentioned that this concept model derived from a combination of various methods is based on a series assumptions. These assumptions have been discussed in the previous sections. Overall, they include:

(1) The separation of two baseflow, which are the relative proportion of the fast baseflow and the slow baseflow, can be determined using DIFGA results.

(2) The average residence times of groundwater samplings can be taken for those of the slow baseflow.

(3) The residence times of two fast components can be performed by those obtained with the tracing tests.

Rippach



Loechernbach

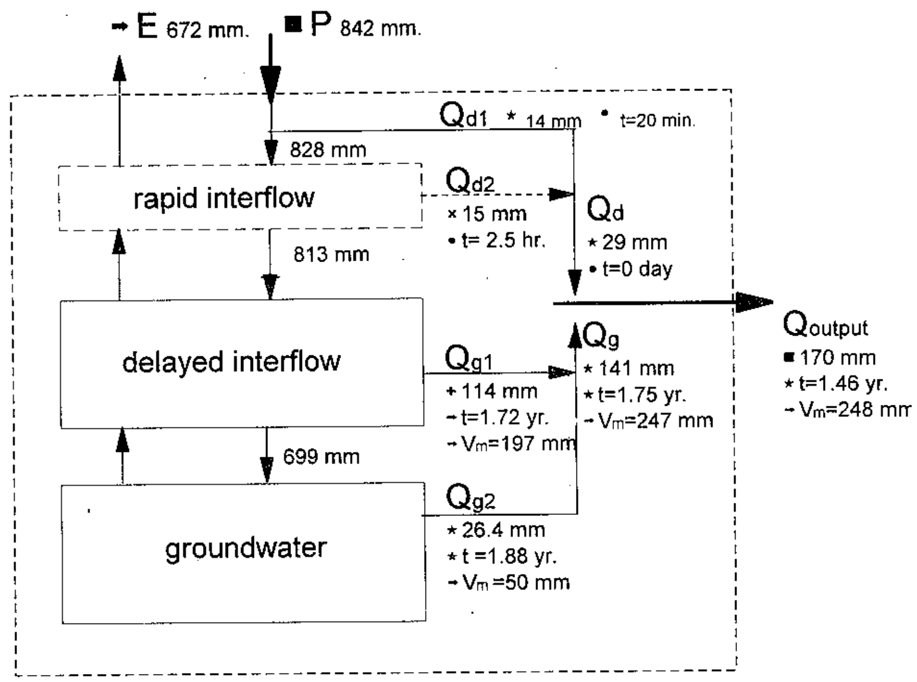


Fig. 8.4 Concept model of the study sites, a combination of the results obtained by isotope, hydrochemical tracers and the DIFGA model

(■ Hydrometric data; ★ oxygen-18 data; ® chloride data; ⊄ DIFGA data; ● tracing test data; □ calculated data)

9 Conclusions

The objective of this study was to contribute to the understanding of runoff generation by means of a recession analysis model and tracer studies. The analysis was completed for seasonal periods as well as for individual runoff events.

The conclusions of this study are summarized as below:

(1) As an altered form of the traditional graphical approach, the DIFGA model provides a possibility to separate long time series of recorded daily discharge into flow components which are described as single linear storages. The calculations verify a small proportion of effective precipitation versus total precipitation at both study sites, and show the dominant contributions of long-term baseflow in the Rippach catchment and of short-term baseflow in the Loechernbach catchment. Compared with isotope data, the direct runoff percentage for the small-altered basin is correct in the order of magnitude. The corresponding percentage for the altered basin is, however, underestimated by approximately 20 %, The failure of the models mainly is believed to result from non-linearities inherent in the loss function for interflow. For this component Darcian principles may be not applicable in the altered basin. The storage constants of the fast components determined by the DIFGA method are inaccurate. The tracing experiments showed the very short average transit times of overland flow (20 minutes) and those of rapid interflow (2.5 hours). These times are much shorter than those obtained by DIFGA.

(2) The long-term experimental investigation incorporating isotopic tracer demonstrates a dominant old water fraction in the streams (92 % at Rippach and 83 % at Loechernbach). The 92 % of old water fraction in the Rippach is one of the highest groundwater fractions reported so far. The results reflect the important characteristic of loess to store high amounts of rainwater in undrained areas. The old water fraction in summer could be overestimated, because some peak discharge may be missed despite a high sampling frequency. The probable error range of the estimate for old water fraction using $\delta^{18}O$ is 5-8 %.

(3) The minimal variation of the Cl^- concentration in groundwater in the study basins allows for the separation of runoff components. Old water fractions obtained by Cl^- are larger than those obtained by $\delta^{18}O$, which is explained by the dilution of the solute in the soil.

(4) The study emphasized that the comparison and combination of isotope and hydrochemical tracers are necessary. This is because the results of hydrograph separation using either chemical tracers or isotopes may be erroneous. Hydrochemical tracers such as Cl^- may not always be conservative, depending upon the soil characteristics to each watershed and biological activity of the stream in question; and the large fluctuation of stable isotopes in precipitation also reduces the accuracy of the calculation on a weekly scale. It was possible to estimate the order of rapid subsurface flow by combining the results derived using both $\delta^{18}O$ and Cl^- based on the two-component model, thereby to conclude the negligible rapid interflow at Rippach and 8 % of this component at Loechernbach.

(5) The mean residence times, determined by the dispersion model (DM), were 1.60 years in the Rippach and 1.46 years in the Loechernbach catchment. The groundwater reservoirs showed mean residence times of about 1.71 years and 1.75 years respectively, implying the mobile water storage in the order of 217 mm at Rippach and of 247 mm at Loechernbach, about twice the annual runoff on average. The best fittings were found to be by baseflow. The use of isotope data also enables the determination of the infiltration coefficients between summer and winter: 0.98 at Rippach and 0.68 at Loechernbach. These results confirm underestimating of the values obtained using hydrological data.

(6) The short-term investigations demonstrate that event water dominated the major phase of the storm events in both basins. 78 % of event water which occurred on July 3rd, 1995, in the altered basin is larger than those reported in the literature, and indicates that event water can dominate an event, depending upon the basin characteristics. Independently, the tracing experiments also confirmed a rapid reaction of overland flow from sealed areas and a rapid response of subsurface flow to rainfall. The contribution of rapid subsurface flow may be larger than overland flow, according to the calculation data. It is thereby concluded that macropores, mainly due to agricultural activities, allows vertical bypassing of the unsaturated matrix very rapidly. The drainage networks (artificial subpipes) conduct considerable amounts of runoff during storms, and enables rapid subsurface flow to enter the stream. The insignificant contribution of rapid subsurface flows to stream in the Rippach is due to the absence of drainage subpipes, thereby confirming that in this basin macropore flow via natural paths entering into stream is minimal.

(7) A significant result in this study was: The very different residence times and storage volumes that are documented by using either the isotope method or graphical method are mainly due to the different boundary conditions of both methods, rather than being only due to the uncertainty and unsatisfied interpretation of single-linear storages explained by graphical methods. This suggests that the criticism to traditional graphical method through isotope studies (e.g. SKLASH 1979, EDEN 1982, HERRMANN & STICHLER 1980) may not be correct in every respect. This is because:

First, the residence time obtained by $\delta^{18}O$ is significantly longer than that obtained by the storage constant. An excellent example is the altered basin of Loechernbach, where the fast baseflow component using $\delta^{18}O$ is in a yearly order of residence time, while the daily scale set by DIFGA for fast baseflow is lower by two orders of magnitude. This points out that isotope yields the time of how long the water is stored in the basin (residence time), while the storage constant provides information on how fast the water flows into stream (response time).

Second, previous studies did not strictly distinguish between the two different types of groundwater reservoirs obtained from models or graphical method and from isotope data. The results of this study suggest that graphical methods relate to a energy or dynamic reservoir, which is smaller than mobile water reservoir determined with isotopes.

In summary, the graphical method relates to temporal variation of runoff, while isotope method relates to both temporal and spatial variation, in particular the displacement effect at depth. Therefore, simple comparison of the model with the isotope results is not possible and is best avoided.

(8) For two-component model, baseflow, rather than individual groundwater data, is suggested to be appropriate as old water. Weighting tracer concentrations for both input and output should be taken carefully in account too. In addition, the uncertainty or sensitivity analysis of two-component model is necessary, especially for long-term runoff separation. Based on the analysis, the uncertainty in the large-terraced basin is larger than that in small-terraced basin; and the uncertainty in summer larger than in winter. Moreover, new water results in larger uncertainty compared to old water.

The sensitivity analysis for storm events suggests that diluting salts and their dissolution in the upper-layer of soil can largely increase the event water fraction. EC is suggested to be quite sensitive to possible increase of the value in new water. The field observations indicated the EC value for event water should be considered between 100-200 μs, rather than 15-35 μs in precipitation. It means EC value for event water should be empirically determined on a basin by basin basis. Obviously, EC value set too low for event water results marked underestimate of event water fraction to stream. In the present Loechernbach basin this underestimate could be by 13 %.

In present hydrology, the identification of runoff sources and paths is still a major topic. Clearly, the relative importance of variable sources and their temporal nature is still not fully understood.

Previous studies cautioned the use of graphical methods for hydrograph separation in the absence of experimental data. Indeed, the tracer studies demonstrated when these methods are valid and sometimes, reveal unexpected hydrological phenomena, thereby contributing to the development of theory. Although the present study emphasized isotopic procedures, they cannot and should not replace empirically-derived data and other methods. Particularly, for many hydrological problems, one of the main concerns is how water moves within the hillslope system rather than its origin.

This study is based on the investigation data of one year. This period is long enough for the identification of runoff sources: new water and old water fractions. However, this period is still too short for the accurate determination of residence time using flow models and of corresponding groundwater reservoirs for the study sites. Therefore, the estimate of the residence times in this study is only correct to within an order of magnitude.

Three facets of runoff generation demand further study:

(1) Macropore flow and pipeflow. The detail mechanisms of these flows under surface are still not clear. Further investigation, in particular the research of the storm mechanisms, will be very helpful. The amount of installed drainage subpipes in the present altered basin Loechernbach provide an excellent condition for the research, therefore this basin remains an interesting test catchment.

(2) Coupled models of runoff generation processes and associated chemical transport. This coupling has had limited success so far in simulating the observed chemical behavior of streams. In addition, uniform explanation of soilwater definition and determination of different boundary conditions are particularly helpful when hydrologic models are linked to chemical transport relationships.

(3) Storage reservoirs. Different methods or tracers may obtain different reservoirs. In some cases they are only a part of the total groundwater reservoir, neither total mobile nor mobile plus immobile water reservoir. The connection and difference of these reservoirs such as dynamic water, minimum water, mobile water, and immobile water reservoirs should be investigated.

One important direction of hydrology in the future should integrate the various concepts of models and tracer methods into one theory, rather than focusing only on calibration of models or providing only additional information about the watershed. If such a theory is possible, the future of study to runoff generation will be widened.

Notation

a	parameter of relationship of baseflow and groundwater depth (DIFGA)
A	surface area of the river basin in km^2
Ain	amplitude of input sine curves
A_{out}	amplitude of output sine curves
b	parameter of relationship of baseflow and groundwater depth (DIFGA)
B	time conversion constant of DIFGA model (86.4)
C	storage constant in days
c	tracer concentration
c_d	tracer concentration of new water
$CD1$	storage constant of rapid direct runoff (DIFGA) in day
$CD2$	storage constant of delayed direct runoff (DIFGA) in day
Cd_i	concentration of oxygen-18 fractionately collected for a certain time
$\overline{C_d}$	isotopic contents of the weighted average precipitation
Cg	tracer concentration of old water
$\overline{C_g}$	isotopic content of the average base flow in the whole observation period
$CG1$	storage constant of short-term baseflow (DIFGA) in days
$CG2$	storage constant of long-term baseflow (DIFGA) in days
C_i	monthly oxygen-18 content of precipitation
$\overline{C_i}$	long-term average oxygen-18 content of precipitation for the total observation period
C_{in}	input concentration
C_{out}	output concentration
$\overline{C_{out}}$	long-term average oxygen-18 content of output for the total observation period
$\overline{C_t}$	isotopic content of the weighted average stream water
Ct_i	oxygen-18 content in streamwater for a specific time
ΔC_d	errors or uncertainty in the oxygen-18 of old water
ΔC_g	errors uncertainty in the oxygen-18 of new water
ΔC_t	errors uncertainty in the oxygen-18 of streamflow
D/v	dispersion constant
D/vx	dispersion parameter of the dispersion model
ETR	evapotranspiration in mm
$g(t)$	weighing function of the mathematical flow model (age distribution function)

H	average depth of total groundwater reservoir
h_e	energy reservoir depth
h_s	total mobile storage depth
h_{mi}	minimum storage depth
k	storage constant
K_r	recession constant whose value is dependent on the units of t
L	groundwater level in m
P	sum of precipitation in mm
P_i	amount of precipitation collected for a certain time in mm
P_s	average precipitation for summer months in mm
P_w	average precipitation depths for the winter months in mm
Q	volumetric flow rate through the system
Q_0	intitial discharge
$q(t)$	intergrate discharge from time 1 to t
$Q(t)$	total flow at time t
QD	sum of flow concentration of the direct components ($RD1+RD2$) in mm (DIFGA)
Q_d	discharge of new water in mm
$QD1$	sum of flow concentration of fast direct runoff in mm (DIFGA)
Q_{d1}	discharge of surface runoff in combination of DIFGA with tracers in mm
$QD2$	sum of flow concentration of delayed direct runoff in mm (DIFGA)
Q_{d2}	discharge of rapid interflow in combination of DIFGA with tracers in mm
Q_g	discharge of old water
$QG1$	sum of short-term baseflow concentration in mm (DIFGA)
Q_{g1}	discharge of slow interflow in combination of DIFGA with tracers in mm
$QG2$	sum of long-term baseflow concentration in mm
Q_{g2}	discharge of long-term baseflow in combination of DIFGA with tracers
Q_{g+s}	sum of Q_g and spring runoff Q_s
$Q_i(t)$	flow discharge at time t (DIFGA)
$Q_i(t_0)$	starting value for Q_i (DIFGA)
Q_O	initial discharge
Q_S	drinking water depletion in winter months in the Rippach
Q_t	total discharge of streamwater in mm/year
Q_{t_i}	average discharge value for a specific time

Q_w	average discharge depths for the winter months in mm
$\Delta QG2_i$	increase of long-term baseflow *QG2* during step *i* in m^3/s
$\overline{P_i}$	mean monthly precipitation
r	correlation coefficient
RD	sum of flow formation of the direct components (*RD1*+*RD2*) in mm
$RD1$	sum of flow formation of rapid direct runoff in mm
$RD2$	sum of flow formation of delayed direct runoff in mm
$RG1$	sum of short-term baseflow formation in mm
$RG2$	sum of long-term baseflow formation in mm
R_{sample}	oxygen-18 in sample
$R_{standard}$	oxygen-18 relative to the international standard (SMOW)
s	parameter of EPM model
T	mean residence time of the system
T_a	temperature in °C
t_o	turnover time
v	mean transit velocity of water in the system
V	total storage volume in the system in m^3/s or mm
Ve	storage volume of energy reservoir in the system in m^3/s or mm
Vg	average storage volume of DIFGA in mm/a
Vm	storage volume of mobile water in the system in m^3/s or mm
V_{max}	maximum storage volume of DIFGA in mm/year
V_{mi}	minimum volume
V_{min}	minimum storage volume of DIFGA in mm/year
Vs	total groundwater volume
V_{st}	stagnant volume
W	sum of residual, containing real storage in mm
X	absolute uncertainty of tracer
x	length of the lines of flow
ΔX	inversely proportional to C_g-C_d for a certain value of X
ω	angular frequency of variation
φ	phase lag of variation
α	recession coefficient
α_i	monthly infiltration coefficient
$\overline{\alpha_i}$	mean monthly infiltration coefficient

Abbreviation

DIFGA	hydrograph separation model (**DIF**ferenzen-**G**anglinien-**A**nalyse)
DL1	drainage sampling at Loechernbach
DL2	drainage sampling at Loechernbach
DL4	drainage sampling at Loechernbach
DL5	drainage sampling at Loechernbach
DL6	drainage sampling at Loechernbach
DL7	drainage sampling at Loechernbach
DL8	drainage sampling at Loechernbach
DL13	drainage sampling at Loechernbach
DM	dispersion model
DR1	drainage sampling at Rippach
DR5	drainage sampling at Rippach
DR6	drainage sampling at Rippach
DR7	drainage sampling at Rippach
DR8	drainage sampling at Rippach
EC	electrical conductivity
EM	exponential model
EPM	exponential-piston flow model
GL	groundwater measurement point at Loechernbach
GL2	upstream spring at Loechernbach
GR	groundwater measurement point at Rippach
GR2	upstream spring at Rippach
HYSTATI	statistic model (**HY**drologic **STATI**stics)
LO	Loechernbach catchment
PL1	rain gauge at Loechernbach
PL2	rain sampling at Loechernbach
PR1	rain gauge at Rippach
PR2	rain sampling at Rippach
PRMS	conceptual model (**P**recipitation-**R**unoff **M**odelling **S**ystem)
QL	streamwater gauge station at Loechernbach
QLM	weekly average streamwater in gauge station of Loechernbach
QR	streamwater gauge station at Rippach

QRM	weekly average streamwater in gauge station of Rippach
RP	Rippach catchment
USDAHL	physically-based model (U.S. Department of Agriculture Hydrologic Laboratory)

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