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ASSESSING HYDROLOGICAL CHANGES 5th CONFERENCE OF THE EUROMEDITERRANEAN NETWORK OF EXPERIMENTAL AND REPRESENTATIVE BASINS (ERB)

FOREWORD

One of the main aims of the Euromediterranean Network of Experimental and Representative Basins (ERB) is to encourage international contacts and cooperation between scientists and research teams through data exchanges, scientific meetings and excursions. This volume includes papers presented at the fifth ERB General Assembly and Conference in Barcelona in 1994 on Assessment of Hydrological Temporal Variability and Changes. Since the ERB was established in 1996 conferences have been held at Aix en Provence (1986), Perugia (1988), Wageningen (1990) and Oxford (1992). ERB member countries are at present:Belgium, Czech Republic, France, Germany, Italy, Netherlands, Poland, Romania, Russia, Slovakia, Spain, Switzerland, and United Kingdom. A focal ERB tool is the basin inventory ICARE, which provides an informal database concerning the data of small research basins in ERB area. The ICARE inventory is located at CEMAGREF in Lyon, and it is accessible to scientists from ERB member countries.

The themes of the Barcelona Conference focus on examples of field experience from ERBs for the analysis of temporal variability and spatial differences in hydrological phenomena. Whereas previous ERB Conferences largely dealt with methodical and comparative hydrological problems which arise on a small basin scale, the present contributions are dominated by anthropogenic environmental impacts using the hydrological cycle and climatic issues as examples. Many of the studies presented in this special issue were conducted in Mediterranean environments, considering specific hydrological behaviour and effects such as soil erosion in one of the most degraded regions since early historical times. Some important ecohydrological aspects of land use pattern and change were demonstrated by field excursions to the Vallcebre study basins in the eastern Pyrenees, and the Prades research catchments situated in the Catalonian Pre-Coastal Range.

We hope that this collection of scientific papers will contribute to strengthen links between scientists working of ERBs, and to establish new scientific contacts. We are grateful to the organizers of the Barcelona ERB conference and editors of this proceedings volume, Dr. Pilar Llorens and Dr. Francesc Gallart from the Institute of Earth Sciences Jaume Almera (CSIC) in Barcelona, and to the publishers of Acta Geologica Hispanica for inclusion of this volume in this series.

Andreas Herrmann
ERB Coordinator

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Preliminary hydrological results from Sarennes glacier basin, French Alps

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ABSTRACT

Located in the French Alps, Glacier de Sarennes is a small glacier which has its mass balance, measured since 1948.

Now since 1992, the Snow Engineering and Avalanche control department of Cemagref Grenoble measures water level in the emissary of this glacier at the altitude of 2800 meters. The time step of these records is 10 minutes. The aim of this study is to realize the first hydrological assessment, despite having few records until now, to be able to control the new experimental discharge station, and to extract the first scientific results.

INTRODUCTION

The E.R.Bs. *(European Research Basins)* represent excellent tools for hydrologists: real open air often of small size, they offer powerful means to observe the water cycle and its processes *(Barbet D., Givone P., 1992)*. The Sarennes E.R.B., one of the recent French E.R.B.s, presents two special particularities:

- the length of some data: more than forty years for annual glacier balance *(Valla F., 1989)*.
- its specificity: the only French glacierized E.R.B.

This glacier located in the French Alps (30 km east south-east from Grenoble), at the northern border of the French Oisans massif, is studied for the mass balance since 1948, firstly by the "Eaux et Forêts" Service (De Crecy L., 1963), and now by the Snow Engineering and Avalanche control Department of Cemagref Grenoble (Valla F., 1994).

This very long set of data interested the glaciologists, who used it to strictly study this glacier, but also for the general typology of alpine glaciers, the relations between meteorological variables (precipitations, temperatures) *(Martin S., 1977)*, and the discharge forecasts for hydroelectric power generation *(Braun L.N. & al., 1993)*.

The Snow Engineering and Avalanche Control Department from *Cemagref* Grenoble satisfied the hydrologists, installing in 1992 a gauging site near the glacier outlet, on the Sarennes river *(Valla F. & al., 1993 –Tairraz V., 1992)*. The precipitation measurement was improved (with financial help from *E.D.F.-D.T.G. : Elecricitè de France– Direction Technique Gènèrale*), with telemetry of snow cover (daily time step, rain recorder). The aim of this study is to treat the first runoff data sets (1992, 1993, and the beginning of 1994), and to extract also the first hydrological results *(Barbet D., 1994)*.

Three different aspects are examined :
- Firstly an overview of the mass balance data set, and the main results about it.
- Secondly how the water depths are obtained in a mountainous basin, with metrological difficulties, and how discharge set data are issued from these data, how the missing data are filled in for the discharge set by multilinear regression between meteorological va-

riables. This aspect is concluded by a description of the Sarennes typology, with several time steps: annual discharge values, monthly discharge values (seasonal variations), daily discharge values (mean, instantaneous maximum, instantaneous minimum) and finally hourly discharge values (daily fluctuations), with also the description of some historical representative data sets during the melt period, and the study of the relation discharge-temperature with a hourly time-step. Finally the discharges-durations curves are constructed for the two complete years, and an estimation of the flood characteristic duration is also given.
Thirdly the correlation between discharge and meteorological parameters, with a 0 to 5 days lag was studied.

THE SARENNES GLACIER MASS BALANCE SINCE 1948

The Sarennes Glacier covers an area about 0.57 km² for a total area of 1.38 km² at the gauging station (see figure 1). Its coordinates are 45°07' N, 06°08' E. The elevation is between 3327 m. and 2800 m. The access is now easy from the top of Pic du Lac Blanc (top of the basin), using the highest cableway of the ski area known as l''*Alpe d'Huez*.

The main characteristics of this little glacier are :

- a South exposition. Sarennes is a relic of the last glacial extension.

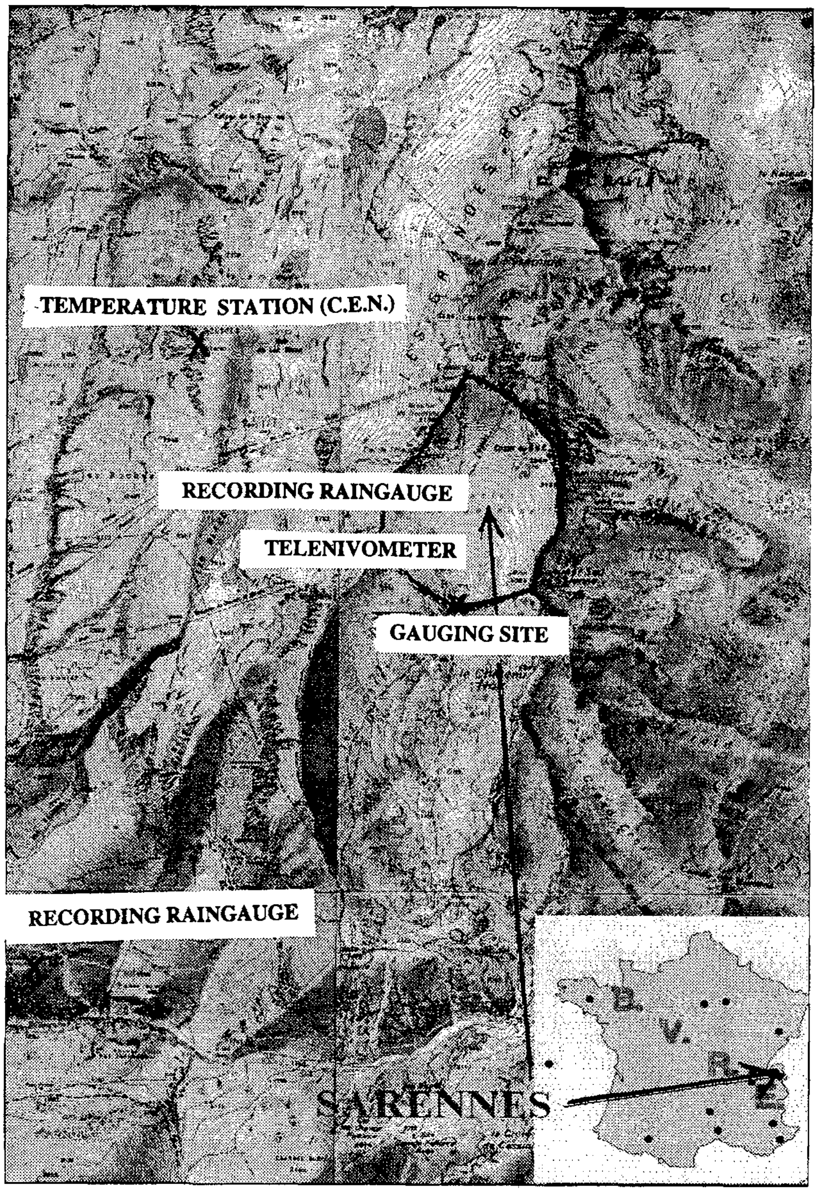


Figure 1. Situation map of the Sarennes glacier.

- a lack of movement as the oscillations are strictly vertical.
- a constant decrease of the glacierized area :
  - 1.09 km² in 1909.
  - 0.70 km² in 1981.
  - 0.57 km² in 1992.

Some historical data.

Despite of its small size, Glacier de Sarennes is known for at least one century. The first recorded visit was in 1891 by Prince Roland Bonaparte, a relative of Emperor Napoleon. He noted that the glacier front was more or less stationary. Fifteen years later, in 1905-1906, three researchers of the Grenoble University conducted important scientific works mainly in glaciology in the Oisans massif. They surveyed the glacier and drew a map at the scale of 1/10 000 with a high accuracy. The first photographs of Sarennes were taken in 1905 attesting the good health of the glacier. After that the Ministry of Agriculture financed several campaigns because the ministry well understood the fundamental role of glaciers in the hydrological behaviour of the alpine streams. As a result observations and data collections were made by the organization called "Eaux et Forêts", now relayed by *Cemagref.* Some photographs and locations of the glacier's front were published in 1927 and 1933.

45 years of mass balance.

With its lack of movement, this glacier is considered as a gigantic raingauge that stores the snow precipitation. Since 1948, four main parameters are measured: accumulation, ablation, mass balance and regime. The mass balance is the difference between accumulation and ablation, the regime is the sum of accumulation and ablation. The figure 2 presents the cumulated mass balance of the glacier since 1948. The graph shows a loss of 30 meters of water equivalent in 45 years, which corresponds to a mean annual rate of 0.66 meter of water equivalent. Because of the lack of significant movement of the ice of the glacier de Sarennes, figure 2 gives an idea of the variation of the ice level at the elevation of about 3000 m. In the center part of the glacier, the loss of ice is roughly 34 meters (= 30 m. / 0.9 ; 0.9 is the ice density.) *(Funk M. & al., 1993).*

FROM WATER DEPTH TO DAILY DISCHARGES.

At an elevation of 2800 m. a gauging station was built in 1992, to measure water level data.
The basin area covers 1.38 km², of which 0.57 km² glacierized. The water level is measured with two different probes: a capacitive water level probe, and an hydrostatic water level probe, in a rectangular measuring weir. The water level is observed at intervals of 10 minutes.

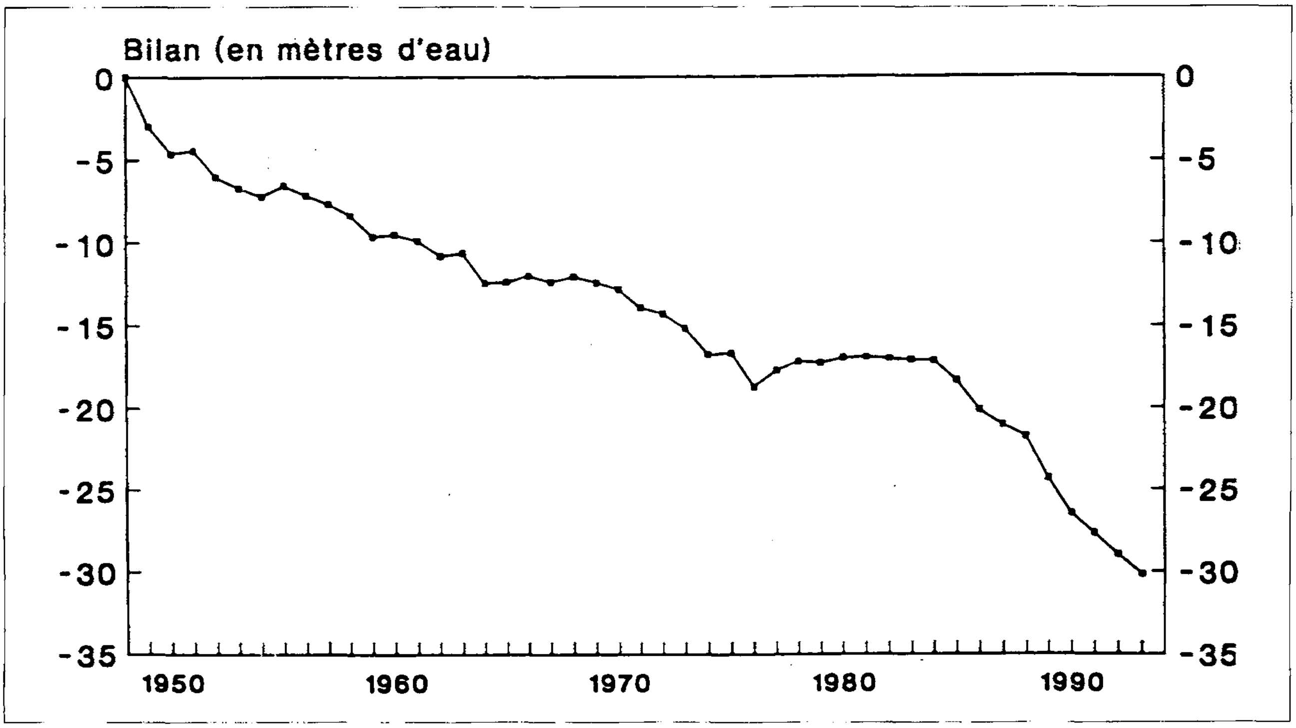


Figure 2. Accumulated mass balance on Sarennes Glacier since 1948.

Hydrometrical problems and their solutions.

An important part of this work was to solve several metrological and hydrometrical observation problems. But these problems were specific to this station: water depth beyond the measures range of the principal probe; differences between the measured depths for the gauging (dilution gauging or velocity-area method) and by the probes, water over the weir for some summer floods... We bypassed successfully the difficulties with classical calculation methods: linear regression, hydraulic classical formulas. A good rating curve was finally

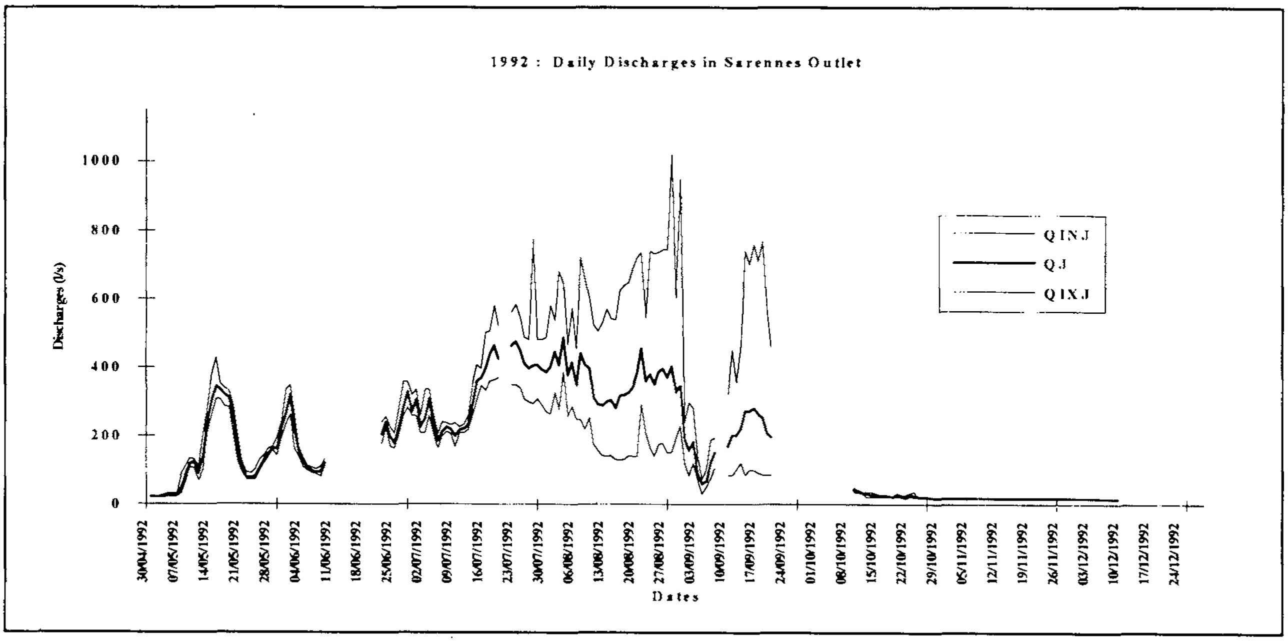


Figure 3. Annual discharge hydrograph (QJ, QIXJ, QINJ): year 1992.

obtained, for the period of the study (1992-1994). Then we could build the discharge year book for different values: daily discharge (called **QJ**), instantaneous maximal discharge for one day (called **QIXJ**), instantaneous minimal discharge for one day (called **QINJ**), the annual discharge, the monthly discharges, and the corresponding graphs. The figure 3 gives the discharge hydrographs (QJ, QINJ, QIXJ) for the year 1992.

Filling in the daily discharge blanks.

Because of a very difficult access (high elevation, snow, no pathway), some days of the years 1992 and 1993 were missing. An interesting part of the study was to fill in missing daily discharge. The method used *(Lang H. & al., 1985)* was an ascending (choice by Student T) multilinear regression between discharge and available meteorological variables (daily temperature and daily precipitations), with a 0 to 5 days lag, adjusted only for the deficient period (with around 15 days before and 15 days after). An example of the results obtained is shown in the figures 4 and 5 for the missing period of June 1992.

The correlation coefficient is equal to **0.89**. The regression equation is:

$$QJ = 136.9+22.6T_{J1} + 0.189P_{J1} + 9.49T_{J4} + j$$

(only for June 1992)

with :
QJ = Daily discharge for the day J.
TJ1 = daily temperature for the day J-1.
PJ1 = daily precipitation for the day J-1.
TJ4 = daily temperature for the day J-4.
φ = residual error

The parameters in the regression equation give some explanation about the hydrological behaviour of this glacier at the beginning of the melt period : in june, the snow cover is still strongly present, the runoff durations are low. Thus the precipitation and the temperature of the day J are not included in the equation. On the other hand, the concentration time between a rainfall event and its runoff seems to be equal to 24 hours (presence of the PJ1 parameter).

For the TJ4 parameter, it is more difficult to explain its signifiance. The recurrence interval between two atmospheric disturbances is roughly equal to four days. These disturbances would make the temperature goes down. This TJ4 parameter is probably not stable, and would disappear in a more general model.

In conclusion filling in this missing period, we find here a strong relation with temperature: the daily discharges decrease considerably when the temperature becomes negative.

Model) were available *(Durand Y., E. Brun & al., 1993)*. The figures 6 and 7 give daily values for temperature and precipitations for the year 1992.

The annual values.

The average values of the meteorological variables are shown in the table 1 for the two years of the study

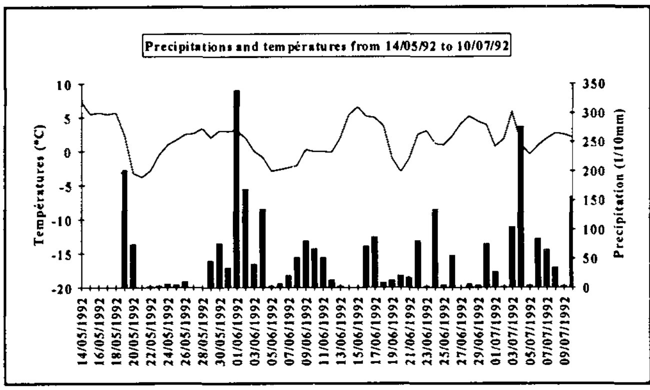


Figure 4. Daily precipitations and daily temperatures for the period of june 1992

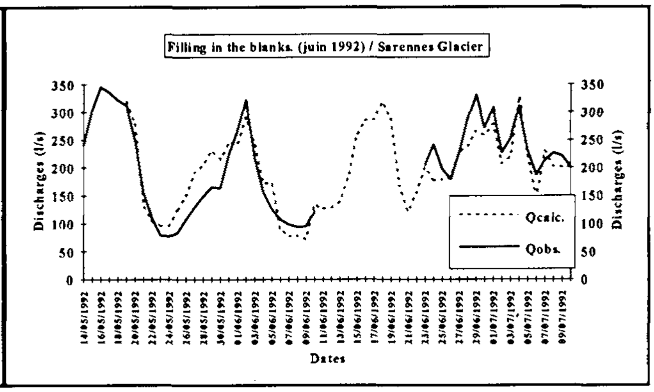


Figure 5. Observed daily discharges and calculated daily discharges: lacunal period of june 1992.

All the missing periods were filled in with the same approach and the same success (good correlation coefficients), but with other parameters following the different periods of the melt season.

To improve the results, the use of a third radiation parameter would seem a good choice in the future.

TYPOLOGY OF THE SARENNES RIVER: A GLACIAL RUNOFF REGIME.

After filling in the discharges blanks, the hydrological behaviour of this glacier is studied. We will precise it with its typology.

The other meteorological variables.

To build the typology of the Sarennes Glacier, meteorological variables are needed. In this study, data from the nearest stations were used.

For the precipitation values, the *Alpe d'Huez* (altiport) station –1860 m.–, with a daily time step records, was chosen. For the temperatures, the *Dôme des Petites Rousses* station, with an hourly time step records was taken. The particularity of this station (*C.E.N. Grenoble*) is that observed values from november to april, and calculated values for the other part of the year (*SAFRAN*

(1992 and 1993).

These values can be compared with some representative French glaciary rivers (Gaudet F, 1973) :

These values show that the Sarennes river has a specific discharge similar to rivers with a comparable rate of glacierized area (the scale effect disappears here with the specific discharge). It can be concluded that these specific discharges are high: the one of the Seine in PARIS is equal to 6.1 l/s/km≤ (interannual modulus).

The high values for Sarennes and the glaciary rivers resulted from :

- the elevation, that involves strong precipitations.
- the low temperatures, that involve a low evaporation.

The flow duration curve.

The flow duration curve is given in figure 8 for the two years of the study.

From the observations of these curves, it can be noticed that :

- the discharge values that are very low have a duration equal to six months for the two years. They are the low flows of the winter period, depending on the temperature and very regular from one year to another.
- The maximum daily discharges, exceeded less than

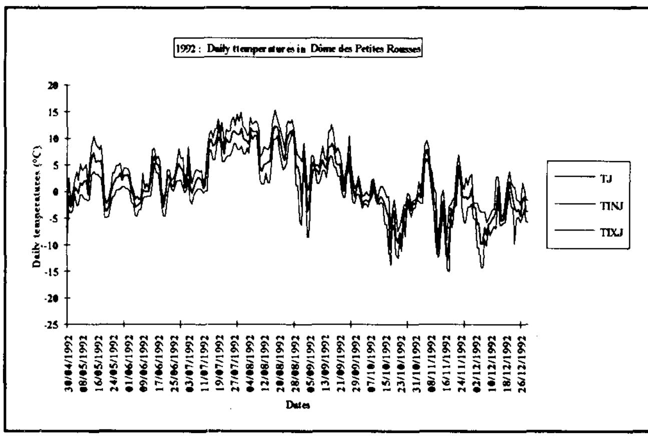


Figure 6. Daily temperatures: year 1992

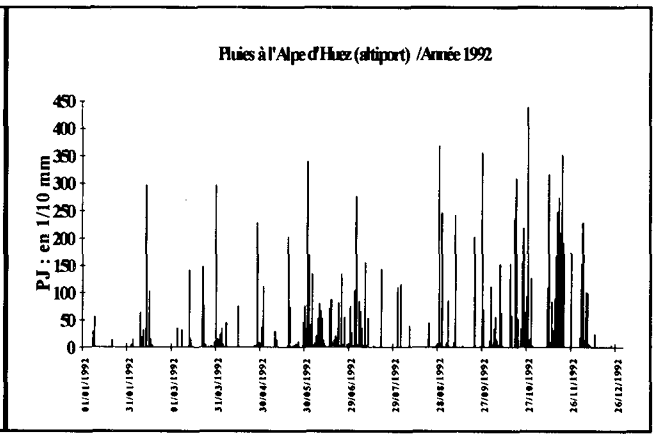


Figure 7. Daily precipitations : year 1992

Table 1. Some characteristic values of the meteorological variables.

	1992	1993
annual depth of runoff (mm)	**2473**	**1854**
annual discharge (l/s)	**108**	**81**
specific discharge (l/s/km²)	**78.2**	**51.3**
maximum daily discharge for the year (l/s)	**488**	**568**
minimum daily discharge for the year (l/s)	**2**	**2**
annual temperature (°C) (Dôme des Petites Rousses)	**-0.5**	**-1.0**
annual precipitation (mm) (Alpe d'Huez / Altiport)	**1223**	**1154**

Table 2: Some characteristic variables of French glacierized basins. (Gaudet F., 1973)

Station	area (km²)	%glacier	elevation (mean in m.)	modulus of annual disch. (l/s)	Specific modulus. (l/s/km²)	annual depth (mm)	Période
Arveyron MER de GLACE	77.98	51	2780	5.16	66.17	2103	1950-71
Arve aux FAVRANDS	205	33		13.5	66	2090	1961-68
Tré-la-Tête	20.9	42	2890	1.53	73	2303	1958-67
Isère à VAL d'ISERE	43	17	2695	1.82	39.6	1250	1951-70
Arc à BONNEVALl	81	29	2754	3.61	44	1400	1948-58
Romanche à PLAN l'ALPE	43	(16)		1.62	37.7	1190	1951-70

one month per year, are very irregular.

It is also shown that the year 1992 is warmer than year 1993: the discharges are higher than in 1993.

In conclusion the annual discharge of the glaciary regime shows a constant annual level, and a strong seasonal variation.

Discharge - mass balance relationships.

The dependance between mass balance and annual discharge has been examined. The relation between both is the following :

$$PA - QA - EPA = VA$$

where :

PA = annual precipitation on the glacier (mm).

QA = annual discharge in the outlet of the glacier (mm).
EPA = evaporation (mm).
VA = variation in volume (mm)
.

The meteorological values for the glacier area were adjusted depending the elevation . Without other data, simple assumptions were made:

- a rate of -0.65°C per 100 m. of elevation for the temperature.
- a increase of 60 mm per 100 m. of elevation for the precipitation.
- a constant evaporation value: 200 mm for a year (minimal value, the maximal value would be near 400 mm).

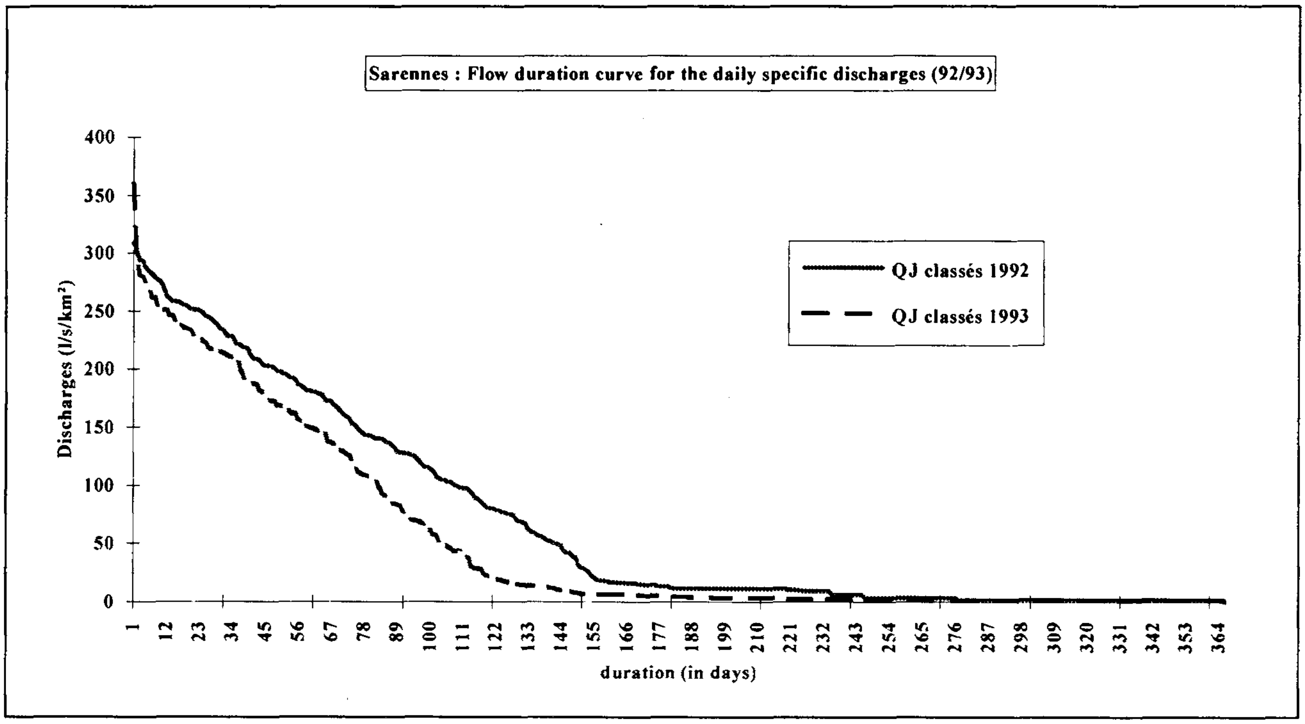


Figure 8. Flow duration curve for the two years 1992 and 1993 (from daily specific discharges)

With these values, a difference between the mass balance and the hydrological balance was found equal to **- 225 mm.** for 1992 and **+ 280 mm.** for 1993 (including the ice thickness lost by the glacier during the icemelt period). These differencies could be explained by :

- an uncertainty about the karstic supply.
- an uncertainty about the precipitations values.
- an uncertainty about the evaporation data.
- and of course, an uncertainty about the measure of water high and the rating curve.

These differencies between both balances can be accepted, with a relative error comprised between 10% and 15%, which is finally correct. But this hydrological balance contains uncertainties.

Seasonal variations.

The glaciary regime is a regular and simple regime, having a melt period with high discharge in summer (june to september), and a long period of low discharges in winter.

Monthly values of discharges, and relation with precipitation and the temperature.

The monthly values of discharge, precipitation (runoff wave) and temperature are indicated in figure 9 for the year 1992.

This figure shows a strong variation of monthly discharges over the year, the occurrence of the highest discharges do not correspond with that of the precipitation which is the proof of a strong snow retention. Three periods of typical discharge regime can be observed :

- the summer (june to september)
- the intermediate months (april, may, october, november).
- the winter (december, january, february, march).

The curve of the temperature is similar to that of the discharge, showing a real dependance between discharge and temperature for this time step. This parameter is the most important to explain the runoff, because a precipitation event can produce no discharge if the temperature is low. In addition to that, precipitation often decreases temperature, and subsequently decreases the discharge, not as usual regimes.

Daily fluctuations of summer discharges.

There are daily fluctuations of the discharges for all the glaciary rivers, due to the daily fluctuations of temperatures *(Obled, 1971 - Power J.M., 1985 - David & al., 1984 - Ferguson R.I., 1985 - Fountain & al., 1985).* These variations were studied for several periods during the melt season, from may to october.

Figures 10 to 14 show the influence of the hourly temperature in the form of the daily hydrograph. To sum up the results, it can be concluded that, except precipitation, hourly discharges and hourly temperatures have roughly the same shape. The drawing is more and more regular as we progress in the melt period: in the beginning of the melt season, the seepage of water through the snow cover is more difficult than through the internal circulation system of the glacier ice at the end of the season. Furthermore the ice stronger reacts than snow to the radiation, because the albedo of the snow is higher.

Finally the concave or convex form of the daily hydrograph can be observed. The ascending curves are straight, and descending curves rather convex, and all the more because the minima are low when the time gets closer to the evening and temperatue drops, then the melt area decreases also, and the melting decreases twice.

Figure 15 shows observed discharged and calculated discharges for august 1992. The calculated discharges are obtained by a multilinear ascending regression between hourly discharges and hourly temperatures with a 0 to 12 hours lag. The parameters of the regression equation are:

TH4, TH9, TH12 (THi is an hourly temperature with a i hours lag). The equation is in log-log. The TH4 parameter corresponds to the four hours between the peak of the hourly hydrograph and the peak of the hourly thermograph. The other two parameters can perhaps be explained by the asymetric form of the discharge curve. The correlation coefficient is equal to 0.99.

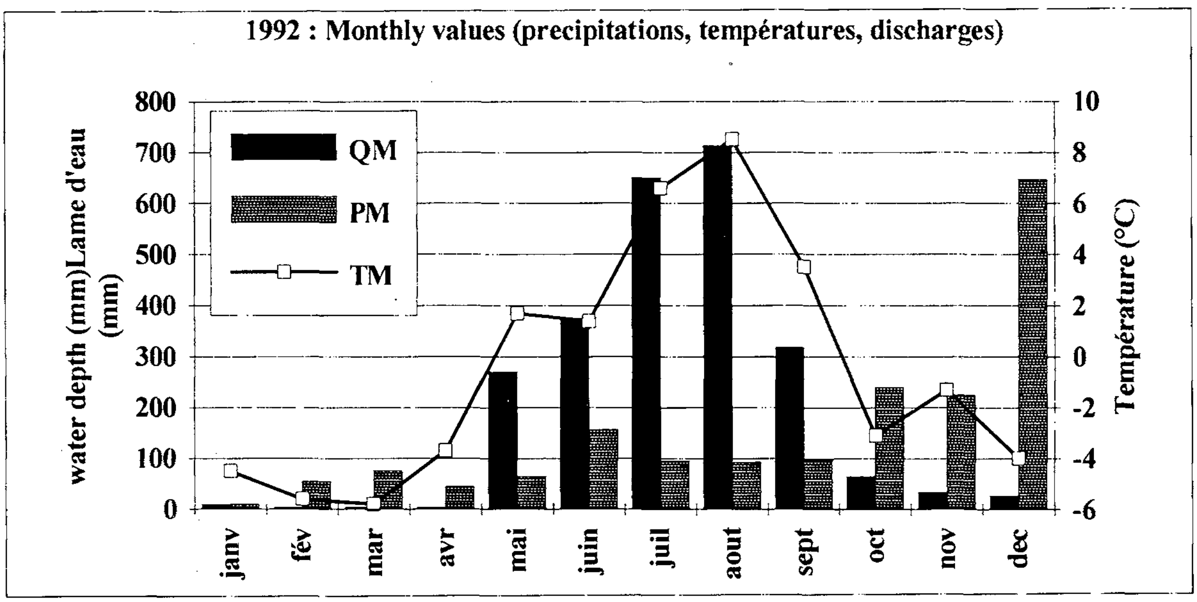


Figure 9. Monthly values of meteorological variables : year 1992.

The discharge-duration curves (with an instantaneous time step) and their interpretation.

The discharge-duration curves *(Galèa G., 1989)*, with an instantaneous time step, between instantaneous discharge and the continued duration when a threshold discharge is exceeded are indicated in figure 16 for the two years of the study. The graph gives also the number of floods for each threshold discharge.

For the discharge-duration curves, there is a break near the 350-400 l/s., this value corresponds to the threshold of the daily floods of the summer. The other curve (number of floods for each threshold discharge) ve the half of the maximal instantaneous discharge) was made for the two years. Except for some outliers occurred during complex floods, a value around 12 hours is found, corresponding to the daily summer floods.

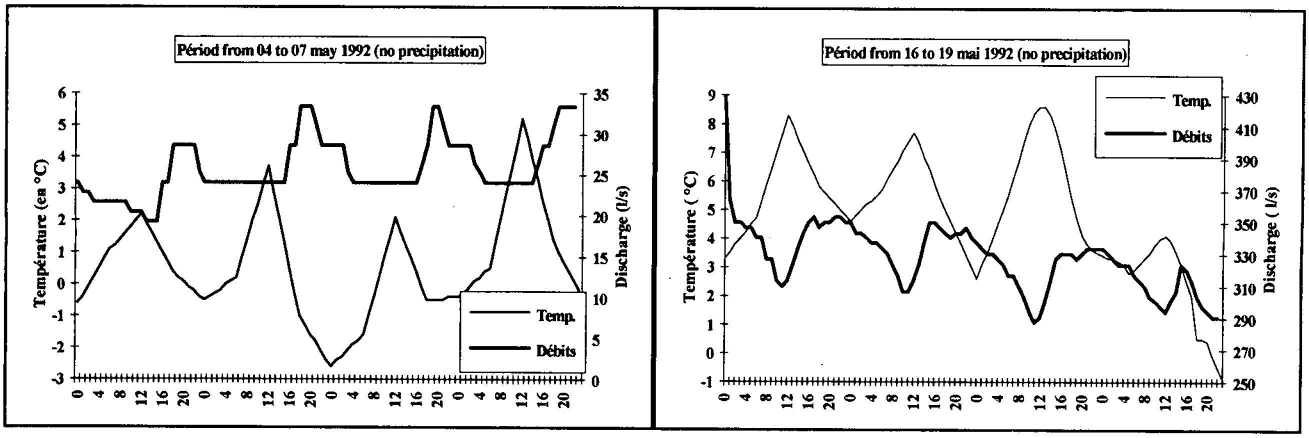


Figure 10 and 11. Link between hourly discharges and hourly temperatures for the periods of may 1992.

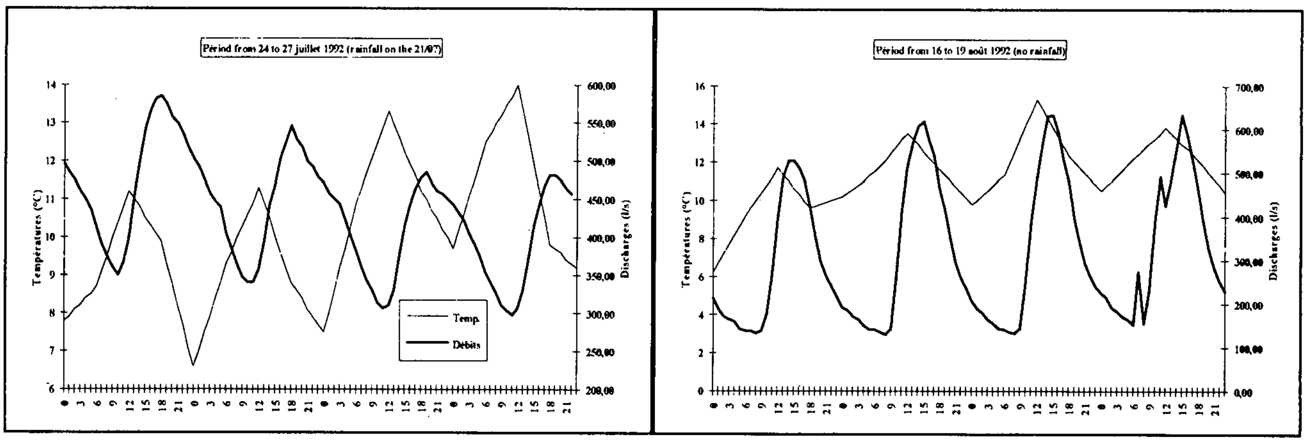


Figure 12 and 13. Link between hourly discharges and hourly temperatures for the periods of july and august 1992.

shows that the maximum corresponds to the same value.This 350 l/s value can be interpreted like the "minimum of the maximum summer discharge".

A very few floods are exceeding 700 l/s : beyond this value, we meet rare floods (storm during icemelt).

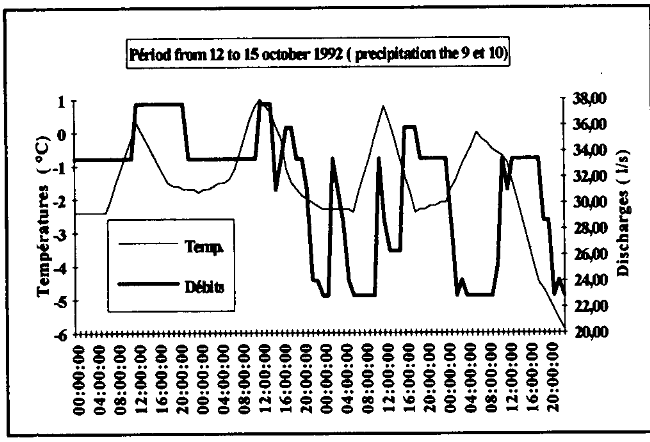
For the number of floods by threshold-discharge, there is another small peak near 25 l/s. This value corresponds to the discharges during the autumn or the spring, where the temperature is near zero degree, sometimes under, sometimes above, with the discharges that vary around this 25 l/s value.

The research of the characteristic duration (basin characteristic : it is the duration when the discharge is abo-

CORRELATIONS BETWEEN METEOROLOGICAL VARIABLES.

Correlations between discharge and meteorological variables

In this study of the Sarennes glacier the method of multiple correlation is used, according to the method applied for the Grande Dixence dam (Lang & al., 1985). The aim was not exactly to find a model, but to observe the simple and partial correlations between the discharges and the meteorological variables (precipitation P, temperature T, and P*T, with a 0 to 5 days lag).
The correlation coefficient are indicated in the figures 17 and 18.
The melt season is divided into three periods : begin



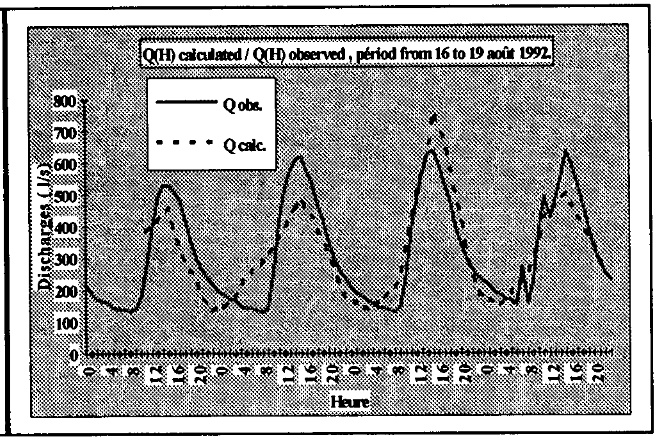


Figure 14. Llink between hourly discharges and hourly temperatures for october 1992.

Figure 15. Observed discharges and calculated discharges with a hourly time step for august 1992

(half may to half july), middle (half july to the end of august), end (september to half october). The two years are grouped together.

The analyse of the results shows for the simple correlation:

- a strong auto correlation for the discharges.
- a strong correlation with the temperature (the best is for T(J-1))
- a low but negative correlation with the precipitations.

The analyse of the partial correlation shows some other results :

- always a strong relation with the temperature.
- the interest to take the P*T variable (good coefficient).
- always the negative influence of the precipitation : when it rains, there are clouds, the total radiation increases, so the melt.

For the "begin" and "end" periods, there are high multiple correlation coefficients (0.80), but worse for the "middle" period probably because of the changing melt from july to august : the icemelt appears after the snowmelt, with a lot of differencies between the two runoffs.

CONCLUSION

Conclusion on measurement.

Measurements in the high mountain are difficult (elevation, strong slopes, snow, no access..). But after these problems have been solved, the results are interesting: Glaciary regimes are not often studied for a long period. For the future, the rating curve must be improved with new dilution gaugings, and new probes have to be installed that accept higher water level.

For the representativity of the meteorological variables a new meteorological station on the glacier must solve this important problem.

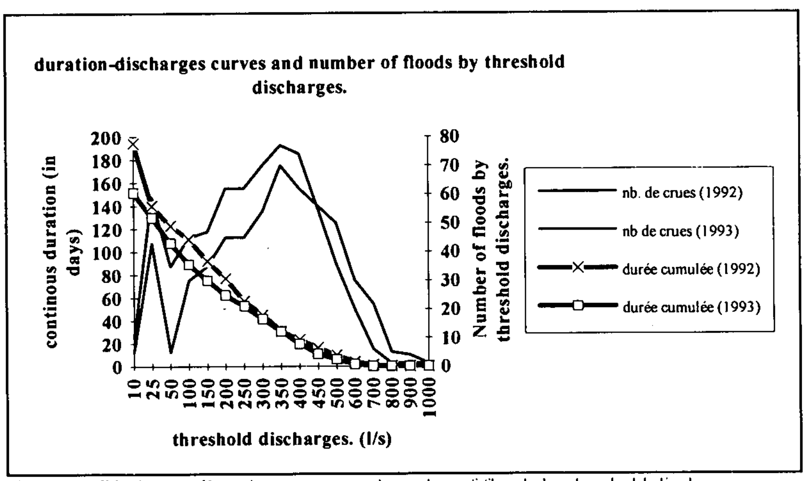


Figure 16. Discharges-Durations curves and number of floods by threshold discharges.

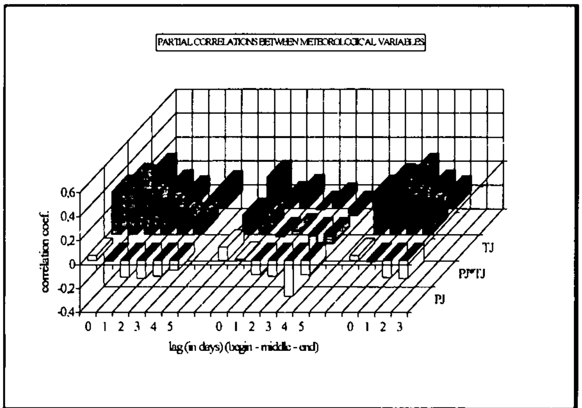


Figure 17. simple correlation coefficient between QJ and meteorological variables (QJ, P, T, P*T, lag 0 to 5 days)

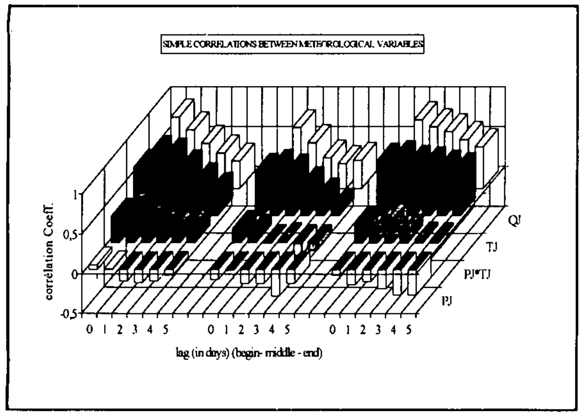


Figure 18. partial correlation between QJ and each separate variable (P, T, P*T, lag 0 to 5 days).

A new gauging station on the same river but more downstream in the valley will help to describe the scale effect and to study the difference between the glaciary regime and the snow-glaciary regime.

Conclusion on typology.

The described typology shows a normal glaciary regime(Parde M., 1933), with the maximum discharge in august or july, depending on the year. The annual runoff/mass balance relation leads to a different runoff total of about 10 to 15%, which can be accepted when the uncertainties of the measures are being taken into account..

The description with different time steps (annual, monthly, daily, hourly) allows to precise the values of annual specific discharge, the form of the monthly hydrographs, in relation with those of the monthly precipitation and the monthly temperature. The relation with temperature appears to be strong as it is a glaciary regime characteristic. The daily fluctuations of summer discharges were analyzed depending on the three melt periods, each period presenting a characteristic form that depends on the runoff conditions for that period. A calculation with an explanatory model has been done on the icemelt period where the daily fluctuations are the more regular, with a multilinear regression between hourly discharges and hourly temperatures, with a 0 to 12 hours lag. Then the discharges-duration curves were established to find again the summer floods (around 400 l/s) and the floods concerning the intermediate months, when the temperature appoaches the zero degree. With the next data sets of discharges, we would be able to find the discharge-duration-frequency curves (Q-d-F model, *Galea G. 1989*).

Concerning the explanatory variables and the modeling of the daily precipitation and the hourly temperature, good results were foundfor the period of "beginning " and "end" (correlation coefficient equal to 0.8), but less good for the "middle" period, which is more heterogenous (passage from snowmelt to icemelt). We observe with these partial correlations that the relation is strong between discharge and temperature, good with discharge and the product temperature with precipitation, low and negative with the precipitation.

The continuation of collecting instantaneous water level data on the two gauging stations, and data from the new meteorological station must permit to improve these different points.

The first data sets are encouraging to continue the study and to apply different models, to study floods, to find other explicative variables (radiation)... The future research in this E.R.B. will be interesting for the study of the climatic change.

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High mountain basins in northern Chile: water balance problems in an arid volcanic area

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ABSTRACT

In, the arid volcanic area of northern Chile, over 3,100 m, the overflow layer is very high and shows little variation in time. During dry years, it may exceed the precipitation layer and it seems independent from rainfalls. The natural overload is very scarce and the resources that actually are developed could be considerated as fossil groundwaters.

To explain these anomalous values of the water balance, the most likely hypotheses are three: 1 - missestimation of actual precipitations (infiltration of snow cover); 2 - existence of external contribution (drainage from the east below the volcanic range); 3 - discharge of powerful aquifers (1st, the climatic conditions in the Holocene have been much wetter than previously considered; 2nd, thick detritic formations are interstratificated in the volcanic complex).

1 - INTRODUCTION

Located be° 30' W long., on either side of the Tropic of Capricorn, the region of Antofagasta rises from the Pacific Ocean level up to more than 6000 m, elevation of the highest volcanoes of the Andes (Fig.1).

The region is characterized by its extreme aridity. Hydric resources are very limited but the mineral resources are abundant and they need important water volumes for present and future exploitation. The increase of hydric requirements for mining causes acutes conflicts with other consumption sectors and makes up a drawback for regional development.

2 - MAIN CHARACTERISTICS OF THE REGIONAL ENVIRONMENT

Rainfall and geology are the two main components of geographic environment that play an important role in the genesis of water resources.

2.1 - Pluviometric framework

The three main causes of the generalized regional aridity of northern Chile are: the almost permanent influence of the Southeastern Pacific Anticyclone, the proximity of the cold Humboldt Current and the presence of the Andean Range.
Along the coastline, highly influenced by the first two factors, the climate is stable. There are no large thermal variations, rainfall is almost totally lacking and water resources are particularly scarce.

To the east, the climate is controlled by the conditions proper to the Highlands, and particularly by the fact that the Range constitutes a barrier to Atlantic influences. Alternatively, the higher parts are influenced by Pacific air masses during austral winter and by continental air masses from December to March, period during which wet hot ascending air masses originated in the Paraguayian and Argentinian Chaco are pushed to the west. They produce rainfall and thunderburst, and peaks can be seen covered by the snowy layer characteristic of the bolivian or altiplanic winter. In relation to temperatures, high daily or seasonal thermal variations are detected.

Quantitative level: the available data records reflect the density of rainfall and climatological measurement stations. These are only a few, due to the great difficulties for access and because an extensive part of the territory is completely isolated during several months a year. The precipitation network, consequently gives an imperfect representation as compared with regional extension heterogeneity, on the one hand, and with respect to the desert conditions that are characterized by an extreme irregularity, on the other hand. It does not show the real number of rainfall events and does not allow a proper estimation of the space distribution of rain or of the water-equivalent of the snow precipitation in the highest areas. Anyway, it is still too early to define a synoptic pattern which can lead to forecast or which can even correlate some meteorological characteristics to snow or rainfall events, that are randomly occuring.

Rainfall data records are reliable only at a local scale and show that summer precipitation has a high interannual irregularity and make up about 90% of the pluviometric amount, meanwhile the other part of the year is normally dry.

2.2 - Geodynamical context

The subduction of the Nazca oceanic plate under the western edge of the continent, that is responsible for the seismic activity and active volcanism of the east zone, has also an influence at a structural level. Major parallel faults run approximately northeast-southwest and control a block-tectonic system which defines different morphostructural compartiments (Fig. 2). In relation to the sequence and constitution of the geological strata of the area, the traditional interpretation proposes that most outcropping rocks have a relatively recent igneous origin (plutons and volcanites). Also, in relation to detritic formations, the importance of which is relevant owing to their aquiferous potential, it can be quoted "...durante este período (Mioceno-Holoceno), imperó un clima hiper-árido, con escasa erosión, aunque se formaron extensas cubiertas aluviales y algunos depósitos lacustres y salinos, que constituyen las actuales planicies (pampas) y salares localizados en la Depresión Intermedia y Cuencas Intramontanas" (Boric et al., 1990)[1].

Quantitative level: the reliability of this thesis is questionable, but this may be because it does not provide the necessary details. This drawback is mainly due to the presence of a wide recent ignimbrite layer that covers numerous geological limits and tectonic accidents. It also reflects the fact that geological surveys were done almost exclusively from a mining viewpoint. It is interesting to note that in spite of the unavoidable erosion-sedimentation cycles resulting from the Andean orogenesis, such a pattern does not take into account the detrital deposits. It also implies that arid conditions have been permanent for about ten million years; thus, this interpretation excludes the hydric processes from the formation and deposition of thick layers of continental sediments existing in the intermediate region.

The whole study area corresponds to an extremely complex volcanic environment. The great number of episodes of magmatic-volcanic activity is shown through a sequence where volcanic lavas and tuffs, with diverse composition and large lateral heterogeneity, alternate. Besides, detrital formations resulting from altitudinal adjustments of the Andean uplift are interstratified and the whole sequence is often covered by recent strato-volcano cones and lavaflows.

3 - PRESENT RESEARCH PROGRAMME AND OBJECTIVES

Scientific cooperation between UCN and ORSTOM have as a purpose to remedy the water scarcity since 1991. One of the aims of the program is to understand the mechanisms that control the formation and circulation of surface flows and groundwaters.

In this perspective, the study proposes to quantify the value of the different components of the simplified water balance equation, built on the global model:

- 1 - INPUT, production function, or entrance to the hydrological system: This component corresponds essentially to the meteoric water; it is measured by series of observed rainfalls;

- 2 - TRANSFER, exchange function in the system: This component correspond to the physical/geological context; it deals with the separation of the runoff, the storage capability and the circulation velocity;

- 3 - OUTPUT, restitution function, or exit from the geographical framework: This component corresponds to the amount of water that is produced as a resultant of the previous functions. It is measured at the hydrometric stations located at the exit of the system.

Because of complexity of the geographical environment, and of the scarcity of available data, the study was primarily focused on:

- an inventary of the geological characteristics linked to water production;

- a wide sampling of meteoric waters, as well as superficial water and groundwater, for physico-chemical and isotopic analyses;

- the collection of basic field information and the establishment of a preliminary hydro-climatological network. The study areas included the endorheic watershed of Rio Zapaleri (in the altiplano, or puna), and the upper watershed of Rio Loa (the only exorheic river in the area);

- a critical analysis and processing of the existing data like the Rio Salado (an affluent of Rio Loa, Fig. 3) records. This watershed has been selected as one of the most representative basins in order to establish the first water balance available in this region. Analysis of these previous data (J. Köhnenkamp, [2]) showed several anomalies that we shall now examine.

4 - ANALYSIS OF THE RESULTS

4.1 - Preliminary results on a regional scale

Systematical geological reconnaissance led to re-examine some concepts. For instance, in the highest part of the watersheds, these studies led to identify some thick and wide continental units of post-Miocene age (Fm. El Loa conglomerates). Besides, detritic sediments and unconsolidated tuffs were found interstratified in some ignimbrite sequences (e.g. Toconce Fm.).

From the isotopic analyses, two types of results should be emphasized:

- snow profiles sampled in june 1993 indicate clearly that the water had a continental source;

- the almost total lack of 3H (confirmed by B. Messerli et al., 1993, [3]) in the water of the drainage net means that the water is older than 1945.

PRELIMINARY CONCLUSIONS: In first place, it can be inferred form the fact that some detritic formations results from strong hydric origin, that climatic conditions have been much wetter than previously considered. In this respect, the thick layers that show typical aquifer characteristics could be recharged at that time and constitute major reservoirs. As expressed by B. Messerli et al., 1993, [3]: "The early Holocene (11-7 kyr B.P.) experienced wetter conditions ... After about 3,000 B.P. conditions became drier; ...", one should forget the prevailing concepts about post-Miocene hyper-aridity.

On another hand, the results of the isotope analyses, which still need to be reconfirmed, point to the predominant role of the continental air masses. These results suggest that the rainfall mechanisms are rather complex. Finally, they indicate that overflow water in the drainage net is mainly of groundwater origin.

4.2 - Water balance in Salado River basin

Owing to its importance for water production (470 l/s for Antofagasta tap water plus 1,090 l/s to meet some of the requirements of Chuquicamata copper mine) the Salado River has been equiped with one of the oldest hydroclimatological observation networks of the region. This network gathered useful data even if some stations provided only episodical records.

This basin is located between 22° and 22°30' S lat., and 67° 45' W long. (Fig. 3), at elevations between + 2,530 and + 5,960 m and extends on 2,390 km .

The interesting area that controls the cordilleran system of water genesis in which are faced the major problems to equilibrate the terms of the hydric balance, is located over 3,100 m. This area, called upper basin involves the Sifon de Ayquina hydrometric station and measures 748 km .

Annual means of temperatures vary from 12° to 2°C according to the altitude (altitudinal gradient of 0.42°C/100m), with extreme values close to 30°C and lower than -15°C. The O°C isotherm, snow position index, is a function of the season of the year: it is fixed at 4,700 m (average temperatures) and 2,800 m (minimum temperatures) in July, while in January it rises up to 5,500 m and 4,200 m, respectively.

Although pluviometric data exist since 1968, a reliable statistical treatment can only be applied to series after 1974 (table 1). 90% rainfall is concentrated during the Bolivian winter, from December to March, while the austral winter does not show more than scarce and weak rainfall or snowfall that interrupt the intense dry period lasting the other part of the year. Daily rainfall rarely exceeds 40 mm, except for rainfalls observed in February 1977: 64 mm and 54 mm in Toconce; 47 mm and 50 mm in Linzor and 48.5 mm in Caspana. The rainy episode of the interval from February 13th to 26th, 1977, with exceptional pluviometric amounts: 239 in Toconce, 223 mm in Linzor and 160 mm in Caspana, are probably very scarce. The annual rainfall means, about 10 mm at the lowest point the basin (Loa River confluence), ranges from 50 mm up to almost 200 mm between the lower zone and the upper part of the upper basin.

Among the seven hydro-pluviometric existing stations, only three provide continuous information, being the upper basin control station, "Salado en sifón Ayquina", noticeable for the quality of the information. Its monthly and annual modules are shown in table 2. Floods are scarce, even none in some years, but peaks can reach great magnitudes, the highest being the one observed during the flood from February 16th to 25th, 1977: 288 m3/s, that is 367 l/s/km2. From 1975 to 1990, annual specific modules are fixed at about 2 l/s/km2, with monthly values between 25 l/s/km2 (February 1977) and 1.5 l/s/km2 (November 1979).

PRELIMINARY CONCLUSIONS: According to the methodology exposed in the chapter 3, table 3 presents the annual values of water input and output. In the upper basin, the 15-year-long observation series allows the following comments:

- the terms "deficit" and "overflow" have quite similar values during years with next-to-mean pluviometric frequency;- unless extremely abnormal events occur, such as those in 1976-77, the overflow layer shows little variation in time, with values of about 50 to 60 mm per year. In other words, it seems independent from rainfalls on the basin.

- during the dry years, the overflow layer is very high and may exceed the precipitation layer.

5 - PROBLEMATIC IDENTIFICATION AND INTERPRETATIVES HYPOTHESIS

The study of the results shows an imbalance between the input and the output of water to the hydrological system. According to the present knowledge, the whole physical-climatic characteristics cannot explain the strongly anomalous value of the overflow layers.

Besides the mentioned desequilibrium, it is observed that groundwater of remote origin is playing a predominant role in the flows of the drainage net. It would show that, nowadays, the natural overload is very scarce, almost nil and could also reflect a very long transit underground.

In some way, the runoff watched in the rivers and the resources that actually are developed could be considered as fossil waters.

The anomalous values and the difficulties for equilibrating the terms of the equation from water balance can have its origin in a missestimation of a true value of its components or in the unknown intervening processes. Between the various posible hypotheses, the most likely are the following:

- Hypothesis 1: Missestimation of actual rainfalls: Due to the absence of registering devices in the higher points, pluviometry is probably underestimated. Moreover, we cannot evaluate the contribution of solid precipitation and the snow cover, the infiltration of which could notably increase the value of the entry variable.

- Hypothesis 2: Existence of external contribution: The drainage net, constituted by the intense regional fault and the block-tectonic system, could drive waters from outside the studied area. Maybe from bolivian altiplano (although there is no evidence). It cannot be forgotten that H. Cusicanqui et al. (1975) and W. F. Giggenbach (1978), [4], proved that thermal waters of El Tatio geothermic field are derivated from precipitations fallen in the oriental sector, the same that, on its migration through west and passing below the volcanic chain, acquire their chemical composition and temperature. A contribution of youthful volcanic waters is also possible, though probably limitated.

- Hypothesis 3: Discharge of powerful aquifers: Between the formations of the volcanic complexes would be interstratificated powerful aquifer layers with very important reservoirs. Due to the acumulation of waters until higher altitudes, the ground waters would be found with hydraulic charge and would discharge to the drainage net a flow that may be limitated, but continous and independent from precipitations. This hypothesis, in association with hypothesis number 1, is most likely but needs imperative conditions. Some of them have been already partially investigated in the view of the present studies: need of palaeoclimatological conditions different from actual aridity and the existence of enough reservoir-rock volume.

In any case, these hypotheses are meant to be a referential baseline to define the topics, methodologies, techniques and instrumentation that should be set to contribute with the necessary elements to understand the mechanisms that rule the constitution and circulation of waters in the system.

6 - RESEARCH TO COME: METHODS AND TECHNIQUES

The future studies are based on the mentioned hypotheses. They refer to three different aspects and time scales of fluctuation of the water cycle in reponse to climatic changes:

- interannual: evolution of the snow cover at high altitudes;
- centennal: evolution of surface and groundwaters in systems of internal drainage;
- last 20,000 years: palaeohydrological and palaeoclimatological fluctuatios of groundwater and surface water systems.

6.1 - Evolution of the snow cover

PROBLEMS: It is here considered that snow is the main source of precipitation. The snow cover can undergo several processes of evolution: evaporation in the solid form, melting (evaporation in the liquid form, surface run-off and infiltration) and transformation into ice. The characterization and separation of these different processes will involve:

- an evaluation of the annual average of accumulation in relation with the origin of the atmospheric vapor and with the condensation processes;
- an attempt to correlate the identified processes of evolution to large scale meteorological and climatic parameter.

METHODS and TECHNIQUES: The studies will encompass the following techniques:

- satellite imagery to define the origin of the atmospheric condensing vapor;
- snowgauging for determination of the water equivalent, by using gamma-neutrons gauging;
- estimation of sublimation by measurement of the content in salt of meteoric origin (Cl-, Br-) and of the snow, with automatic sampler (ORSTOM technology);
- evaluation of the ablation out of the melting zone by rainfall and flow gauging; the data collection system will use autonomous and transmitting recorders:
* electronic and computer device for measurement of level and temperature water SPI III, connected to the CHLOE-D limnimetric recorder (ELSYDE/ORSTOM)
* self power recording and transmitting rainfall data system LOGGER 91 (ELSYDE)
* field data transmission to satellite with capability of telemetring: ARGOS system with SRDA direct readout station
- extrapolation of the process of ablation/accumulation at the local scale to the general scale by satellite imagery NOAA, SPOT, ERS1 (radar wavelength) and GOES;
- comparaison of the water balance obtained with the ice balance, through ORSTOM studies on glaciers in Perou and Equator;
- global comparaison with net radiation balance at the stations.

EXPECTED RESULTS: The studies would provide:

- estimates of the amount of snow and fraction of it which is available for the renewal of water resources;
- in Perou and Equator, evaluation of the relative importance of precipitation and temperature seasonal increase or decrease, to account for the general retreat of glaciers.

The simultaneous comparaison with satellite data will allow to generalize the mechanisms at the scale of a large part of Andean range. The corresponding time scale is of some years, in the case of snow, and of some decades, in the case of ice.

6.2 - Evolution of surface flows and groundwaters in systems of internal drainage

PROBLEMS: Water derived from snow melting can infiltrate or flow at the surface. A good knowledge of of the relations between these two parameters will constrain any plans of management of water resources because of the sensitivity of the water cycle to short term climatic fluctuation. Of special relevance with that respect are:

- the very high evaporation rates, with increase in salinity in zones of poor drainage;
- the large variations in permeability of the rocks which model the aquifer system;
- the contribution of perivolcanics emanations which can add various amounts of indesirable chemical components (As, Hg, B) in groundwaters.

This part of the study will investigate:

- the infiltration rate from snowmelt;
- the origin and age of waters in the surface flows and in the aquifers;
- the mechanisms of salinization by evaporation and by dissolution of soluble salts.

METHODS: Emphasis will be put on:

- soils profiles in the zones of snow melting to investigate the infiltration rate by tensiometry and neutron gauge measurements;
- measurements of stable isotopes (2H and 18O) and tritium concentrations of pore waters form the same soil profiles to determinate the fluctuation of the recharge rates;
- flow measurements in some selected sections of rios Zapaleri and upper Loa;
- chemical, stable isotope and radiocarbon (14C and 13C) measurements in surface waters to estimate the conditions of drainage in permanent rivers (selected sections of rio Loa);
- chemical and isotope studies, including volatile components, of El Tatio geothermal field.

EXPECTED RESULTS: This part of the studies will provide informations on hydrological variations at the scale of some decades to some centuries, and will allow to discuss the related influence of recent climatic changes and hydrological events.

6.3 - Palaeohydrological and palaeoclimatological fluctuations

PROBLEMS: This part of the studies aims to a synthesis of all the previous researches for a better understanding of environmental and climatic changes in the arid and semi-arid zones of the Andes.

Due to the relatively recent uplift of the Andean Mountain Range and to the low precipitation over the reliefs, most of the basins are of internal drainage. This implies that the terminal lakes of the surface network are preserving the record of environmental conditions during sedimentation. Most of these basins are presently evolving as saline depressions (salares) but during humid periods or when exceptional precipitation events occurred, groundwater drainage was reactivated and fresh or brackish lakes could develop. This part of the studies aims to identify theses fluctuations at various times scales and to relate them to natural and man-induced changes in hydrological and environmental conditions.

A clear separation between summer and winter precipitation and snow fall events is required for a proper palaeohydrological interpretation. Furthermore, it would be interesting to try to correlate the patterns of recent precipitation and atmospheric circulations to the El Niño events.

BASIN SELECTION AND TECHNIQUES: For its climatic, hydrological, geological and geomorphological conditions, the closed depression of the salar of Atacama has been selected. Detailed studies of high lake levels will be combined with piston coring up to 40 m of bottom soft sediments. The following techniques will be used:
- dating: 14C and 230Th/234U;
- geochemical studies: major and trace elements (Sr/Ca and Mg/Ca), stables isotopes;
- sedimentology/mineralogy/density;
- bioindicators (pollens, diatoms, ostracods, etc.);
- further analytical studies depending on field work conditions.

EXPECTED RESULTS: Assuming a reasonable sedimentation rate of 1 to 2 mm.a-1, a time period up to 20 ka could be explored. This range covers the Last Glacial Maximum, the deglaciation period and the climatic oscillations of the Holocene.

The first purpose is to compare the fluctuations recorded in sediments with those obtained in other sites of the region and with the general scheme of climatic evolution recorded from other parts of the world. At the global scale, the studies will contribute to the debate on the coupling, or decoupling, of Northern and Southern Hemisphere in their responses to orbital or non-orbital forcing during the Post Glacial period and the Holocene. At the regional scale, the studies are expected to contribute to a better definition of the occurrence of major El Niño events.

NOTE: All the studies will provide related contribution to several international Global Change Programmes (IGBP, PAGES, PEP, Palaeomonsoon, etc.). Under UCN and ORSTOM coordination, will participate various research institutes as Laboratoire d'Hydrologie et Géochimie Isotopique, Université de Orsay-Paris XI, Laboratoire de Glaciologie et Géophysique de l'Environnement, Université de Grenoble, France, Dipartimento di Scienze della Terra, Universita di Torino, Italy, Department of Physical Geography, University of

Berne, Swiss, Institut of Geographie, University of Erlangen-Nürnberg, Germany, Niedersächsischer Landsamt für Budenforschung 14C Laboratory Hannover, Germany, and Escuela de Ingeniería Civil de Madrid, España. The extended program will be presented for external financiation to CEE-DG12-"Environment" and ECOS of France.

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Table 1. Annual Rainfall of Rio Salado Basin (mm)(*)

STATION	74-75	75-76	76-77	77-78	78-79	79-80	80-81	81-82	82-83	83-84	84-85	85-86	86-87	87-88	88-89	89-90	PROM
CHIU-CHIU	6.9	10.0	3.5	1.1	9.5	0.0	1.5	1.0	9.5	14.0	2.5	12.0	24.0	2.0	3.0	0.0	6.3
AYQUINA	91.2	44.0	104.3	18.5	23.0	17.6	28.0	7.0	18.5	86.5	27.2	65.5	90.5	9.5	43.0	6.5	42.6
TURI	106.3	50.0	132.9	25.0	28.1	17.9	30.0	8.6	16.3	96.5	48.8	76.0	86.0	10.0	68.0	8.0	50.5
CUPO	183.0	90.1	159.4	18.6	47.0	3.2	33.6	5.6	17.2	201.4	112.0	79.8	100.7	27.5	77.0	7.5	72.7
CASPANA	168.0	113.7	180.5	38.5	46.5	12.0	74.5	11.1	34.1	178.0	47.8	79.1	125.0	15.5	95.0	24.0	77.7
SALADO EMB.	157.8	85.5	175.0	23.0	19.5	14.0	67.0	13.0	22.0	177.3	68.4	107.4	110.9	22.3	102.3	27.2	74.5
TOCONCE	278.3	125.0	264.0	25.0	44.7	24.1	68.6	22.0	36.4	245.6	137.5	118.3	164.0	40.5	123.3	42.7	110.0
EL TATIO	382.2	274.7	312.1	89.1	144.9	54.6	127.2	51.7	124.0	367.9	234.7	155.8	338.1	72.7	134.7	63.6	183.0
LINZOR	411.5	298.4	345.0	82.5	114.2	84.5	182.0	53.0	123.9	353.0	233.2	145.0	279.1	72.6	169.3	31.0	186.1
CONCHI EMB.	44.7	8.8	35.0	15.5	34.0	8.0	16.0	4.0	21.5	42.5	17.5	19.0	39.1	3.5	10.5	1.5	20.1
O. DE S.PEDRO	163.9	175.7	117.6	37.7	66.7	94.5	65.0	33.3	59.0	120.0	148.0	82.7	104.3	21.7	48.5	34.5	85.8
RIO GRANDE	222.5	133.0	168.0	35.9	68.2	30.1	114.9	9.1	59.3	234.5	95.5	104.3	159.9	9.0	89.6	25.5	97.5
AVERAGE	200.1	121.9	185.9	36.0	55.3	24.8	67.7	18.1	44.4	194.0	100.4	92.0	147.1	29.3	88.8	22.2	89.2
U. BAS. AVER.	316.0	214.9	276.8	63.8	94.1	48.2	119.7	37.4	87.3	294.1	172.5	129.3	239.3	53.4	131.6	40.5	144.9

(*) Hydrological year starts on November 1st.

Table 2. Average Flows in "Rio Salado - Sifón Ayquina" (m /s)

YEAR	NOV	DEC	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	MEAN YEAR
							MONTH						
75-76	-	-	2.730	1.613	1.476	1.266	1.276	1.433	1.435	1.411	1.428	1.326	1.539
76-77	1.333	1.310	1.481	19.660	2.080	1.555	1.461	1.476	1.409	1.365	1.338	1.274	2.979
77-78	1.263	1.264	1.365	1.395	1.407	1.407	1.416	1.460	1.460	1.471	1.338	1.340	1.382
78-79	1.268	1.328	1.545	1.245	1.277	1.218	1.254	1.268	1.253	1.259	1.226	1.208	1.279
79-80	1.195	1.229	1.238	1.274	1.488	1.409	1.403	1.444	1.480	1.430	1.510	1.425	1.377
80-81	1.406	1.400	1.467	2.037	1.476	1.469	1.442	1.426	1.436	1.479	1.471	1.358	1.489
81-82	1.316	1.307	1.358	1.403	1.413	1.358	1.337	1.312	1.298	1.332	1.392	1.335	1.347
82-83	1.307	1.372	1.401	1.397	1.429	1.307	1.393	1.381	1.378	1.354	1.490	1.314	1.377
83-84	1.266	1.273	2.880	1.827	1.453	1.338	1.368	1.413	1.440	1.417	1.396	1.351	1.535
84-85	1.327	1.302	1.379	1.911	2.440	1.361	1.349	1.390	1.392	1.335	1.350	1.344	1.490
85-86	1.326	1.339	1.537	1.663	1.351	1.348	1.342	1.337	1.331	1.398	1.326	1.316	1.385
86-87	1.346	1.458	2.448	1.503	1.759	1.325	1.344	1.323	1.303	1.341	1.280	1.280	1.476
87-88	1.253	1.276	1.378	1.389	1.459	1.335	1.351	1.321	1.289	1.291	1.269	1.250	1.323
88-89	1.272	1.294	1.360	2.490	1.425	1.368	1.417	1.411	1.365	1.349	1.371	1.338	1.455
89-90	1.302	1.305	1.347	1.369	1.330	1.305	1.332	1.341	1.341	1.305	1.297	1.254	1.319
AVERAG	1.299	1.318	1.661	2.812	1.552	1.358	1.366	1.382	1.374	1.369	1.365	1.314	1.517

Table 3. Preliminar Hydric Balance of Rio Salado Basin

PERIOD HYDR. YEAR	BASIN	PRECIPIT. mm	RUNOF mm	DEFICIT mm	COEF. OF RUNOFF (%)
75-90	UPPER	134	70	64	52,2
	TOTAL	82	23	59	28,0
75-76	UPPER	215	62	153	28,8
	TOTAL	122	23	99	18,9
76-77	UPPER	277	120	157	43,3
	TOTAL	186	43	143	23,1
77-78	UPPER	64	56	8	87,5
	TOTAL	36	21	15	58,3
78-79	UPPER	94	51	43	54,3
	TOTAL	55	20	35	36,4
79-80	UPPER	48	55	-7	114,6
	TOTAL	25	21	4	84,0
80-81	UPPER	120	60	60	50,0
	TOTAL	68	23	45	33,8
81-82	UPPER	37	54	-17	145,9
	TOTAL	18	21	-3	116,7
82-83	UPPER	87	55	32	63,2
	TOTAL	44	21	23	47,7
83-84	UPPER	294	62	232	21,1
	TOTAL	194	23	171	11,9
84-85	UPPER	172	60	112	34,9
	TOTAL	100	23	77	23,0
85-86	UPPER	129	55	74	42,6
	TOTAL	92	21	71	22,8
86-87	UPPER	239	58	181	24,3
	TOTAL	147	22	125	15,0
87-88	UPPER	53	53	0	100,0
	TOTAL	29	20	9	69,0
88-89	UPPER	132	59	73	44,7
	TOTAL	89	22	67	24,7
89-90	UPPER	41	54	-13	131,7
	TOTAL	22	21	1	195,5

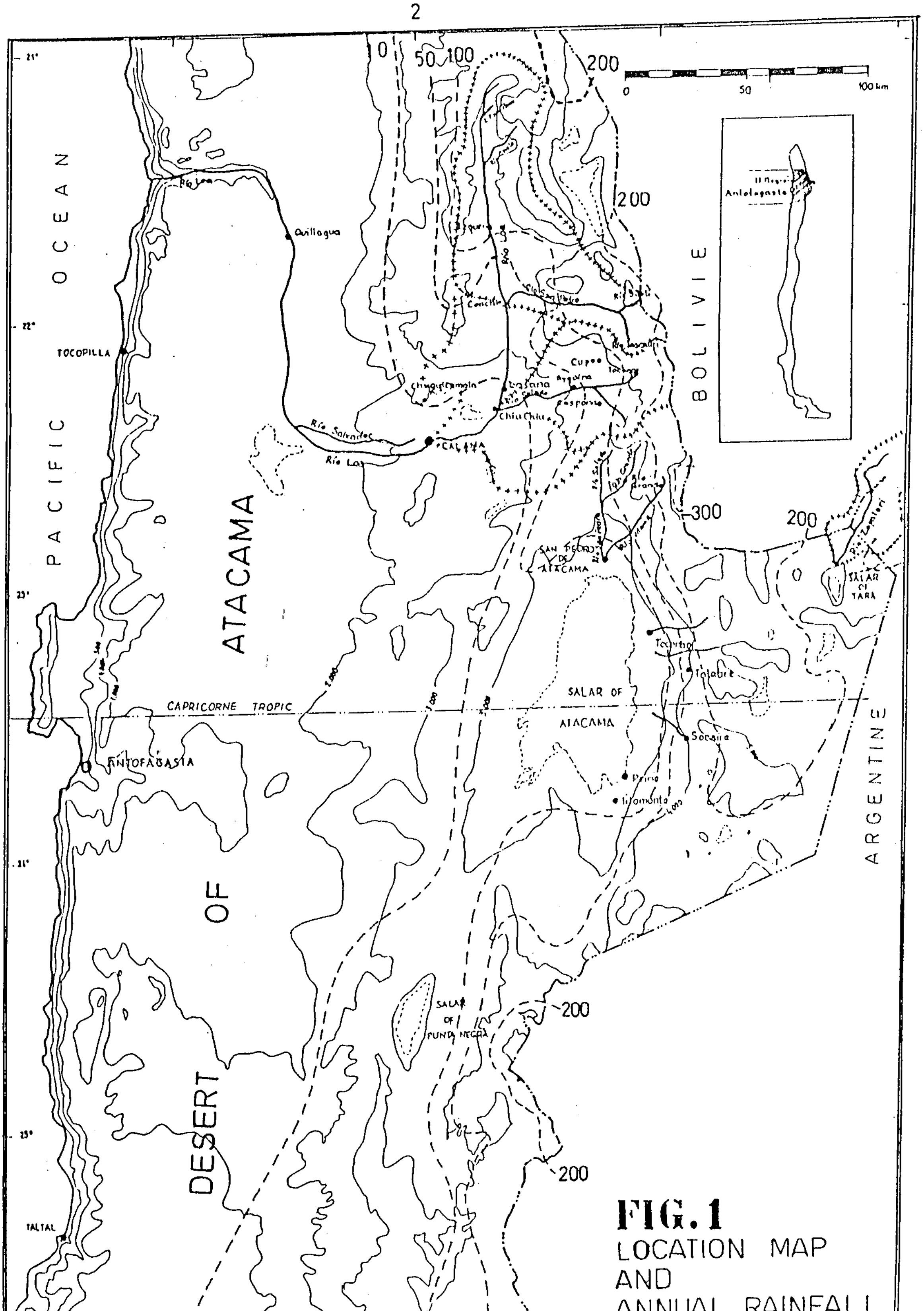

FIG. 1
LOCATION MAP
AND
ANNUAL RAINFALL
IN mm

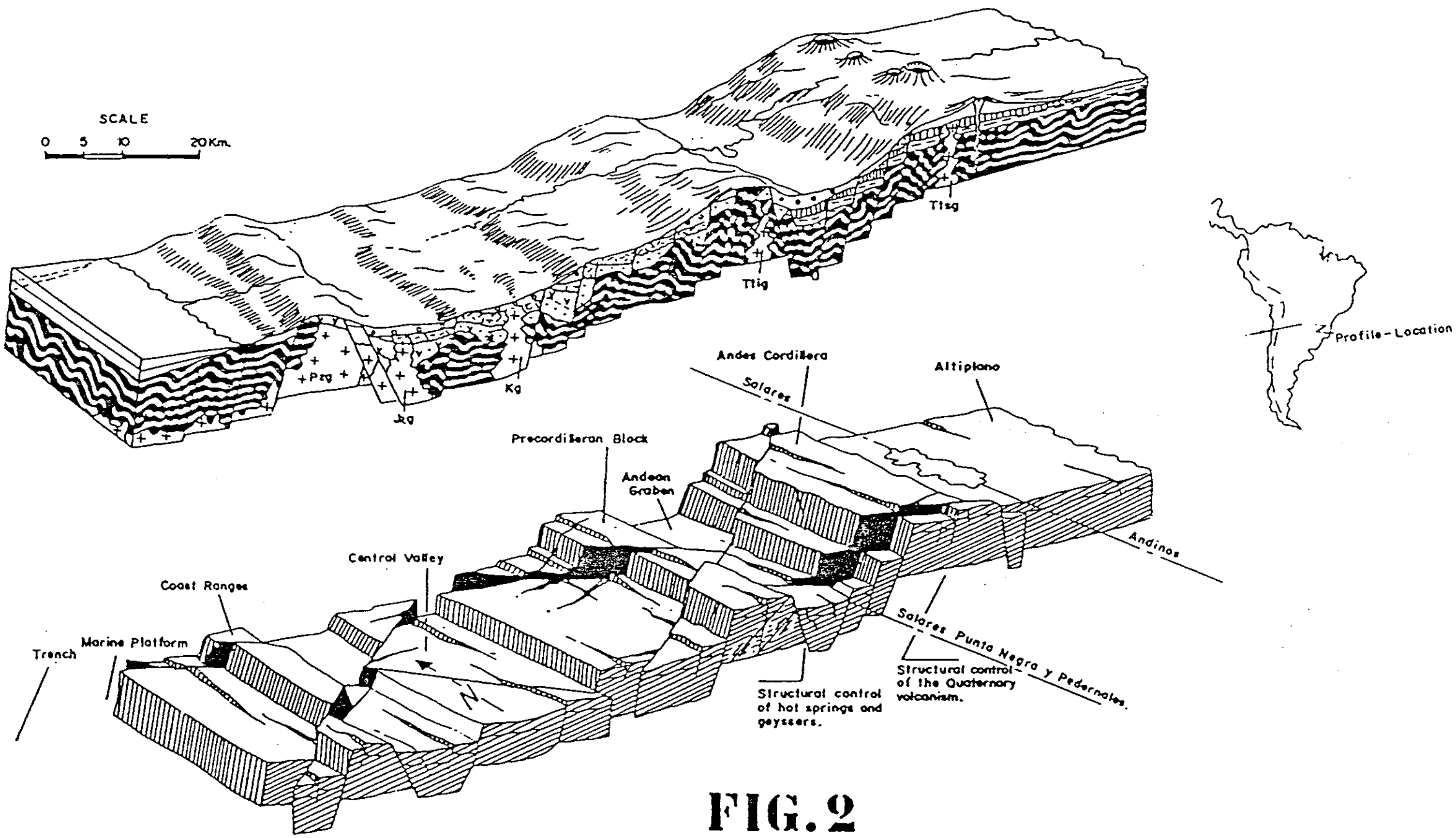

FIG. 2

GEOTECTONICAL DIAGRAM

SOURCE: FRUTOS, J., 1975
Annuaires Soc. Géol de Belgique

POURRUT-COVARRUBIAS

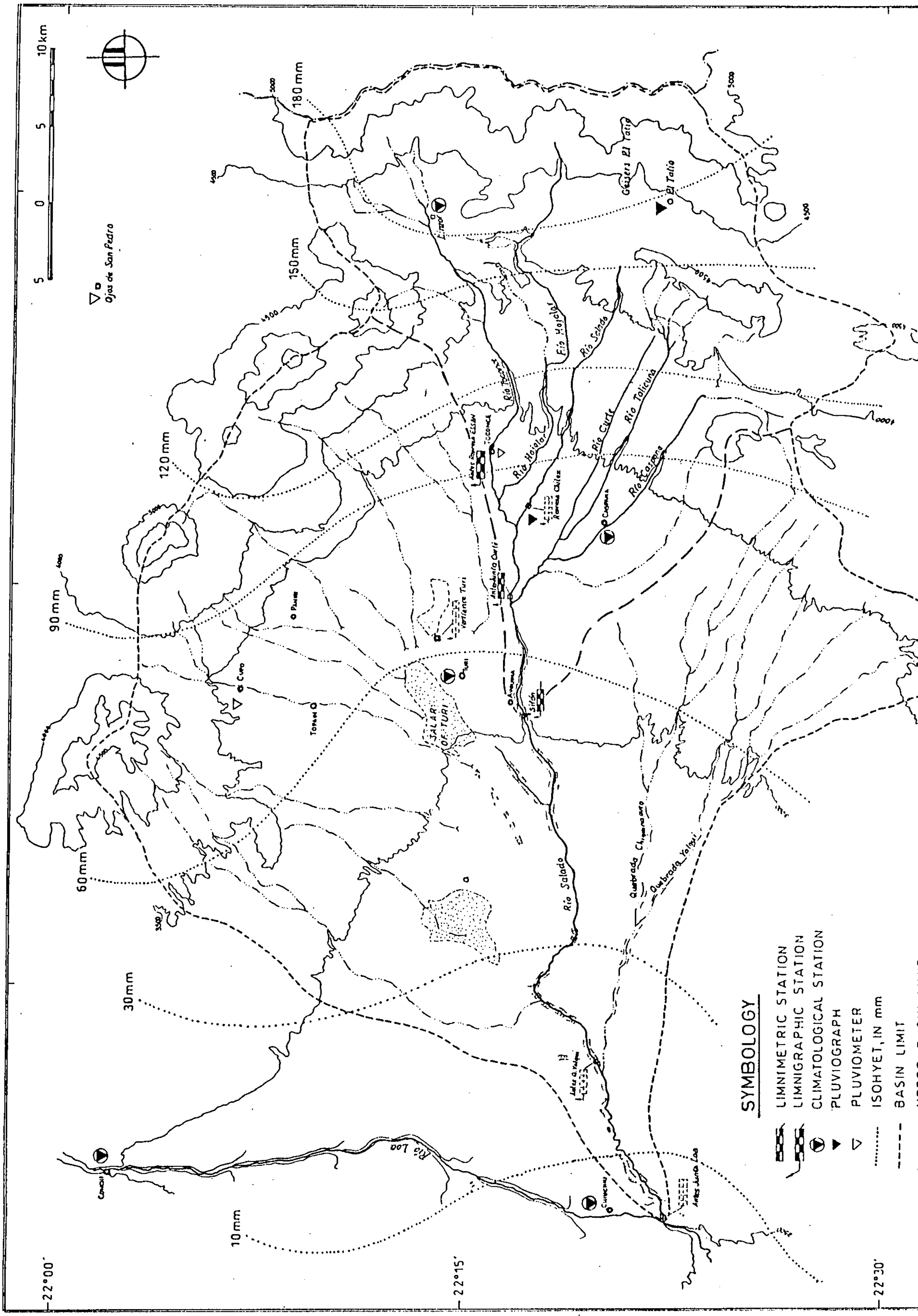

FIG.3
SALADO BASIN: HYDROCLIMATOLOGICAL NETWORK
AND ANNUAL RAINFALL

ACTA GEOLOGICA HISPANICA, v. 28 (1993), nº 2-3, p. 27-35

Variability of gully erosion in a small catchment in South-west Spain

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ABSTRACT

Research on gully erosion is carried out as part of an investigation on erosional and hydrological processes in a small catchment in Extremadura, Spain. The amount of gully erosion is estimated by repeated monitoring of transverse sections since summer 1990. Data obtained so far indicate high spatial and temporal variability. High amounts of erosion are related to a channel section with active headcut retreat, whereas net accumulation occurs where the gully is already deep and wide. Present erosion, apart from headcut retreat, is related to lateral undercutting of the channel bank and bank collapsing.

Temporal variability is related with rainfall characteristics and discharge production in the channel. Important sediment losses are produced by highly intensive storms. The most important gully process in the area is shown to be related with the erosive force of flowing surface waters. Furthermore it is discussed whether present active erosion is related to abandonment of cereal cultivation in parts of the catchment about 30 years ago. It is thought that reduced soil erosion on hillslopes due to the change in landuse, and hence lower sediment supply to the channel in the last decades increased gullying.

INTRODUCTION

Research of gully erosion is carried out as part of an investigation about hydrological and sedimentological processes in a small catchment (35 ha) in Extremadura, Spain (Fig. 1). Landuse of the study basin is the so-called dehesa system, which consists of openly spaced evergreen trees with silvo-pastoral exploitation. This landuse type constitutes about 50% of agriculturally used land in the provinces of south-west Spain. The interest of studying physical processes operating in this ecosystem lies in its economical as well as ecological importance (CAMPOS PALACIN & MARTIN BELLIDA, 1987).

The project initiated in 1990 includes studies on runoff and soil erosion at hillslopes, as well as erosion and discharge production of the main channel. The present paper focusses on gully erosion which is mainly observed in a channel section of about 300 m above the catchment outlet. The principal objetives are:

- to explain the spatial variability and the processes involved in gully erosion and the cause(s) for active gullying,

- to explain the reasons for temporal variability of erosion,

- and to estimate the average rate of erosion in the channel.

For this a survey of gully cross-sections is carried out, which give information on spatial as well as temporal variability. Estimations of net erosion or accumulation are related with discharge and rainfall characteristics. Rainfall during the observation period is compared with long-term data from a meteorological station nearby.

THE STUDY AREA

The study basin Guadalperalón is located in the Cáceres peniplain, an upper miocene erosion surface (GOMEZ AMELIA, 1985). After the formation of the erosion surface rivers have incised, so that the present landscape is characterized by gently undulating slopes in the upper parts and steeper slopes approaching the collectors. Soils formed in precambrian schists or granites are brown mediterranean soils, largely eroded to give place to Lithosols.

Climate is Mediterranean, with hot and dry summers and moderately cold and humid winters. Mean annual precipitation amounts to 511 mm. The annual rainfall distribution shows a dry season lasting from June to September and a wet season from October to March.
The bottom of valleys is formed by fluvio-colluvial sediments, which are thought to be related to accelerated soil erosion in historical time. Thickness of the accumulation is one to two metres and the material is composed of silty sand and gravels. This area possesses a dense herbaceous ground cover and trees are absent. Mainly in the lower part of the catchment fluvial erosion in the form of a gully takes place.

The land is grazed by sheep and pigs. Parts of the catchment were cultivated in the past and were abandoned about 30 years ago.

METHODS

The amount of gully erosion is determined with a repeated survey of topographic cross sections. For this a string is aligned horizontally between two fixed points, consisting of iron poles, at each side of the channel. It is important that the rope is tightly and horizontally fixed. The latter is controlled by a small spirit level which is hung at the rope at different points. For later measurements the string is fastened at the same height, determined as the height above the soil surface at the two iron rods. The distance between the line and the ground surface is measured at intervals of 10 cm with a metal ruler.

Location of fix points as well as headcuts were determined with an infra-red theodolite in order to establish a length profile of the channel and to determine the distance between each transect. Measurements are carried out at least once a year, usually after the rainy period,

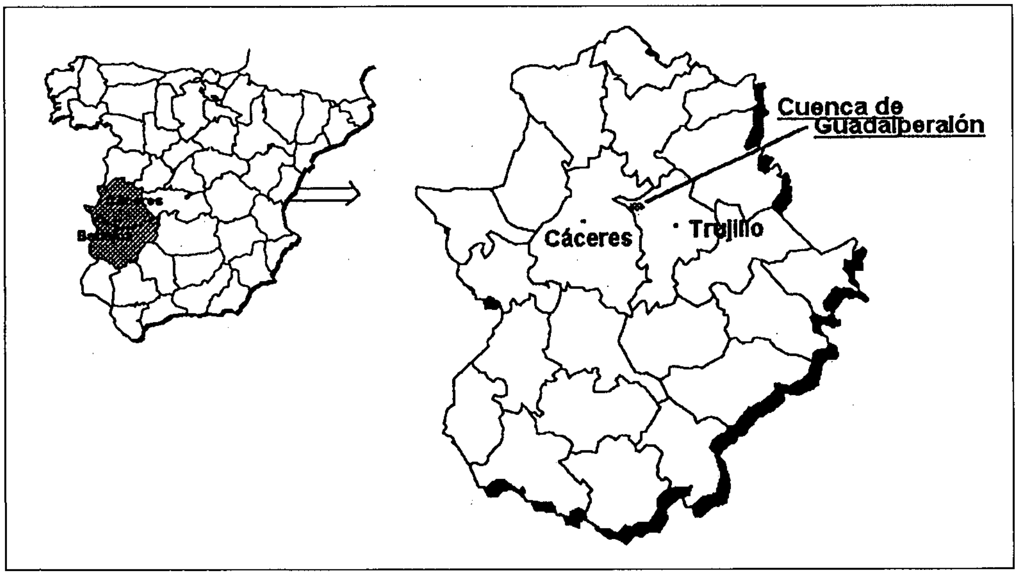

Figure 1. Localization of study catchment. Substrate in the Guadalperalón catchment is exclusively formed by schists. The silty soils are very shallow and have, with about 1,5%, a low content of organic matter. Areas with a tree cover of Quercus ilex alternate with areas lacking oaks. In the treeless zones, where rock outcropping is frequent and a soil is nearly absent, shrubs of the species Lavandula pedunculata are dominant and the herbaceous cover is very poor. At hillslopes where trees are growing herbs are dominant and soils have a depth of 5-20 cm. Slope gradients are in the order of 13% and 25% in the upper and lower parts of the catchment, respectively.

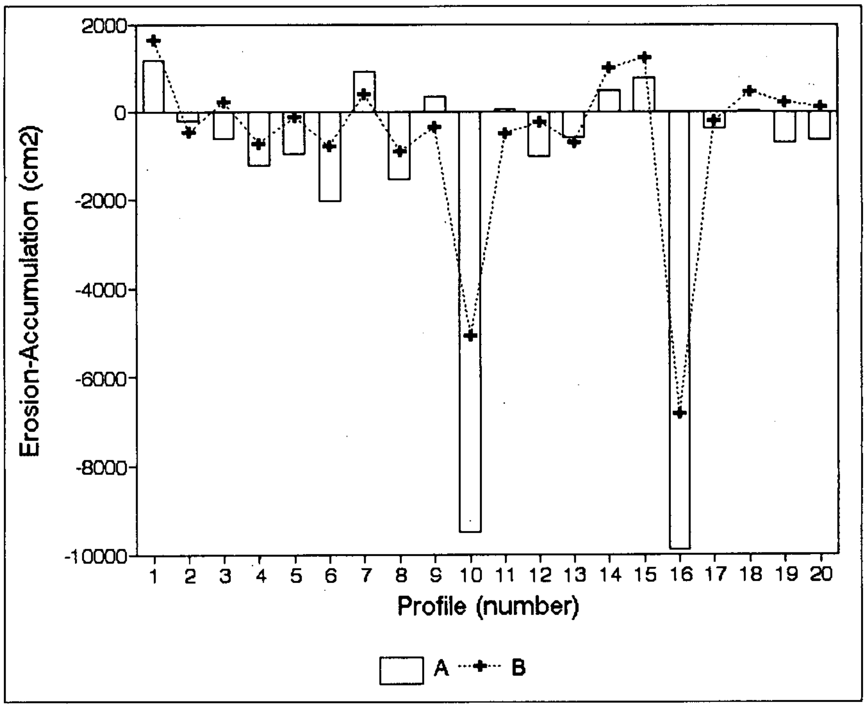

Figure 2. Erosion and accumulation of gully cross-sections: A- total from 1990 until 1993, B - caused by two rainstorms in August 1992.

that is during late spring or early summer. If exceptionally heavy rainstorms occur the survey is repeated.

Net erosion or accumulation of each cross-section is calculated (m^2), and the volume is determined as the mean erosion of two neighbouring profiles multiplied by their distance. Total erosion is the sum of the sections. The profiles are surveyed at intenvals of 15 m where possible. In one part of the channel, with active headcut retreat, the transects are more closely spaced.

Rainfall is registered automatically with a tipping-bucket device of 0,2 mm resolution in intervals of 5 minutes. The gauging station at the outlet of the catchment consists of a 3 feet H-flume with a time resolution of 5 minutes using the relationship between water depth and discharge. (U.S. DEPARTMENT OF AGRICULTURE, 1979).

In order to interpret monitored rainfall events with respect to their probability of occurrence, precipitation data from the meteorological station in Cáceres are analyzed. This station is located at about 30 km distance from the study basin. Maximum annual daily ranfall data are available since 1908 and data on maximum annual 30-minute intensities exist for a period of 37 years. Recurrence intervals and probabilities are calculated using the Gumbel Extreme Value Distribution EVI (SHAW, 1988). Furthermore, the annual distribution of 24-hours and 30-minute rainfall is analyzed. The former is based on data from 1908 onwards and the latter is only available for the last 13 years.

RESULTS

Spatial Variability

Erosion along the studied channel section is highly variable.

Figure 2 shows that the total amount of erosion varies between almost -10.000 cm2 and +1.185 cm2. The two peaks of erosion (Fig.2) correspond to cross-sections 10 and 16, which are located at 7,4 and 1,1 m below headcuts, respectively. In contrast, the area immediately above a headcut is only slightly eroding (profile 11 and 17).

Figures 3 and 4 show the development of cross-section 10. It ilustrates strong erosion due to active headcut retreat, and furthermore shows lateral incision, in one case of almost 40 cm, which caused subsequent bank collapse (Fig 4).

Extent of gullying is fairly smaller in the lower channel section with net accumulation above the catchment outlet. The gully in this area typically has a flat bed and almost vertical banks (Fig. 5). Also here, the dominant process seems to be lateral incision with following bank collapse.

No scouring of the channel bed is observed, except for the area where headcuts are present. Sediment accumu-

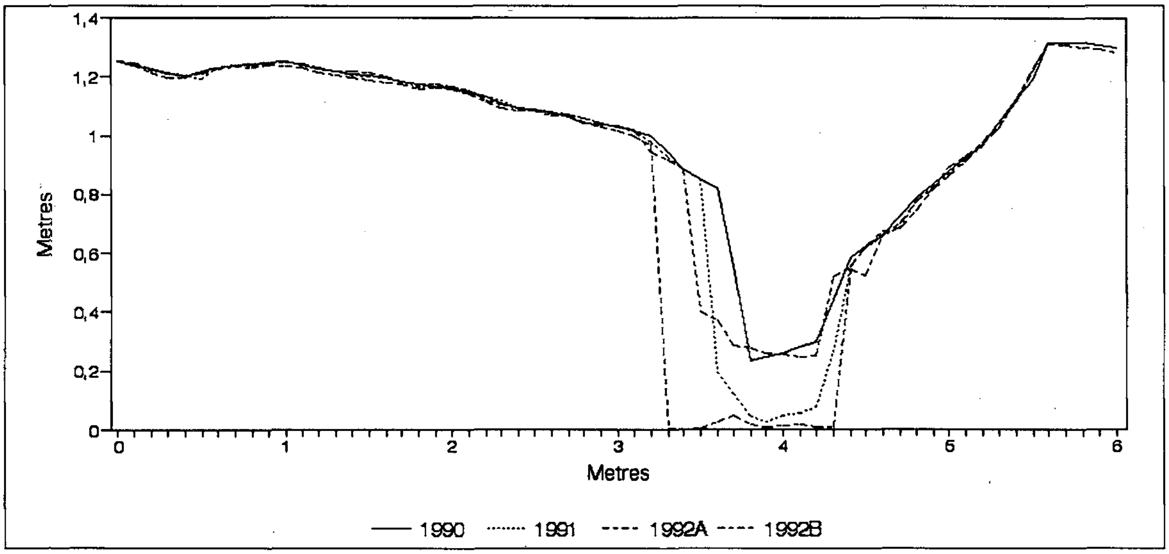

Figure 3. Profile 10, note that 1992A corresponds to June 1992 and 1992B - after two rainstorms in August.

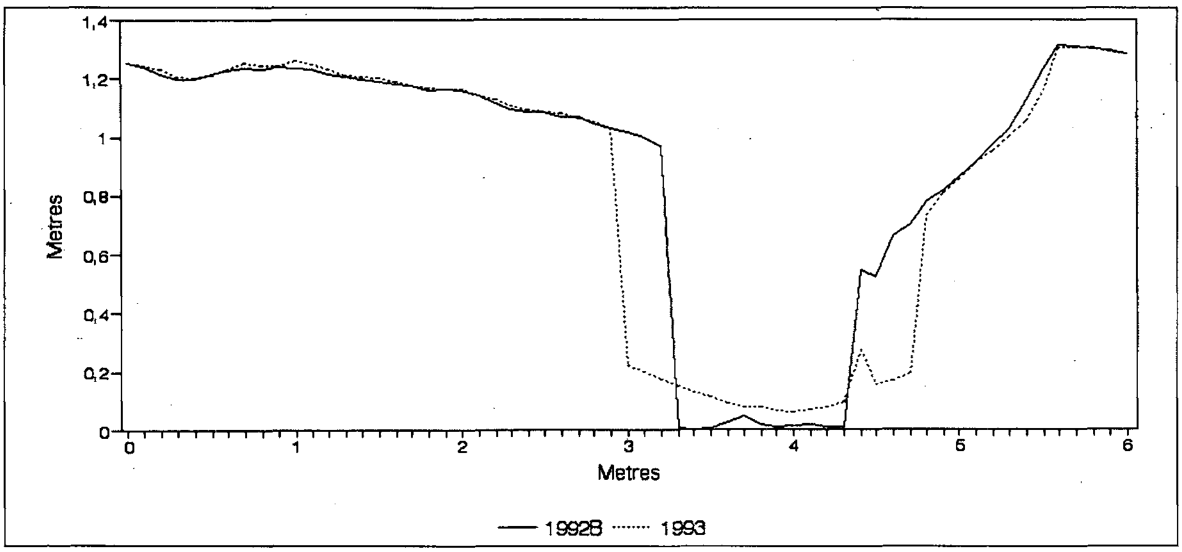

Figure 4. Profile 10, note lateral incision of 1992B causing bank collapse diring the following year.

lated in the channel bed is composed principally of sand and gravels. Figure 6 shows the estimated total volume of sediment eroded or accumulated along the gully, highlighting the described spatial variability.

Temporal Variability

There was high variability of erosion during the three years of observation. Figure 7 shows a maximum corresponding to erosion caused by two heavy rainstorms of August 1992. At some cross-sections 50 % or more of the total amount eroded, is related to these summer storms (Fig. 2). During the hydrological year 1991-92 (except August!) the channel section experienced net-accumulation. The questions related with this temporal variability are:

a) Is there a relationship with discharge and hence rainfall characteritics?
b) Do the results indicate that the dominant processes of gullying are related with the erosive force of flowing water and not, like reported by PIEST et al. (1975), due to moisture saturation causing collapse of banks and headwalls?
c)Is the erosion, which ocurred during August 1992 an extreme event and what does this mean for the estimation of an erosion rate?

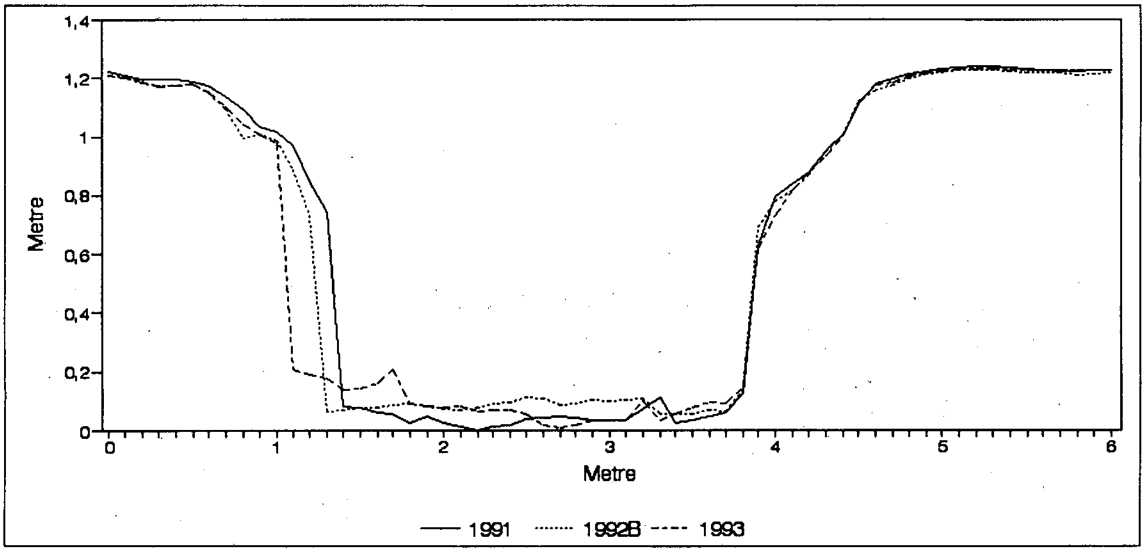

Figure 5. Profile 4, note accumulation during 1991-92, and lateral incision produced by two runoff events in August (1992B) with subsequent bank collapse.

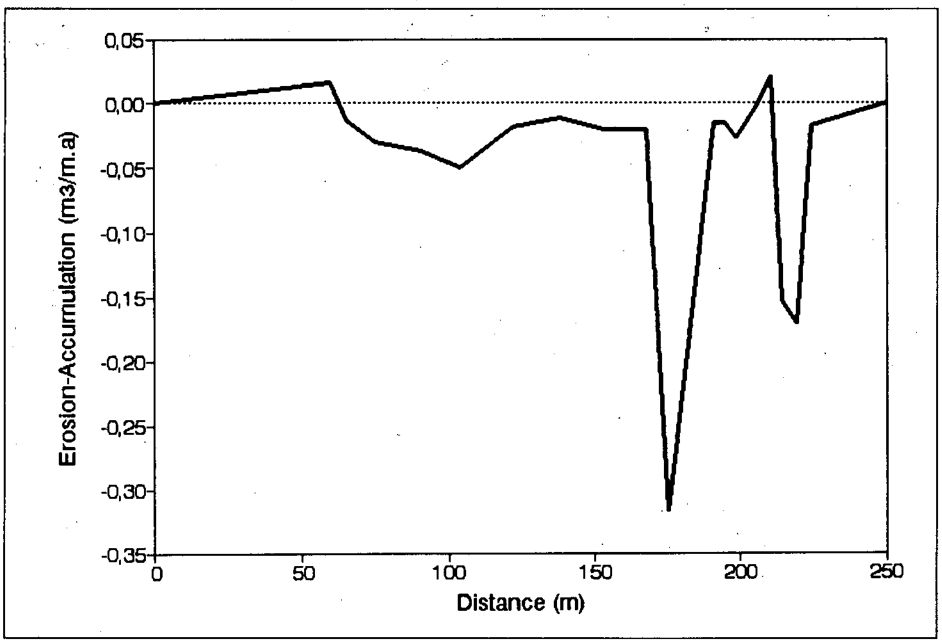

Figure 6. Estimated annual erosion or accumulation along the studied channel section.

No discharge data are available for the first years, and unfortunately also not for the heavy summer storms in 1992. Though, we know that the rainfall at 7/8/1992 of 21,6 mm and with a maximum 10-minute intensity of 60 mm/h caused overflow of the gauging station, which means that maximum discharge was at least 860 l/s. The 10-minute as well as the 30-minute maximum intesity (I-10 and I-30), which constitute the highest observed intensities during the study period, have a recurrence interval of about four years. The rainfall amount has an average annual frequency of five. This event can therefore not be considered a rare one.

Runoff in the cachment is rapidly produced due to the shallow soil cover resulting in rapid discharge production with peak flow occurring 5 to 10 minutes after the precipitation peak. Total discharge is low, with runoff coeficients in the order of 1 to 5% (Table 1). High water flow in the channel is of short duration, usually lasting just one hour or less. Baseflow is insignificant with respect to its erosive force.

Table 1 shows the few observed discharge events, which demonstrates that erosion by flowing water throughout a year is only possible during a small number of days.There is agreement between gully erosion (Fig. 7) and maximum discharge (Table 1). Only a few discharge events having low maximum flows were observed during 1991-92, when net-accumulation was determined. In contrast, highest erosion corresponds to the two rainstorms of summer 1992, and fairly high erosion occurred during 1992-93, when several discharge events of high peak flows were registered.

pecially if short-term data is used for estimating average gully erosion. There is, with two years, too few discharge data available. This is particularly important for the Mediterranean area with its high rainfall variability. Annual rainfall during the observation period was below average (Table 3). The area suffered a prolonged drought starting in April 1991 and lasting until March 1993. As a result the herbaceous ground cover was reduced, culminating during summer 1992 with an almost bare soil surface. The effect of the drought on gully erosion cannot be explained yet. On one hand, reduction of

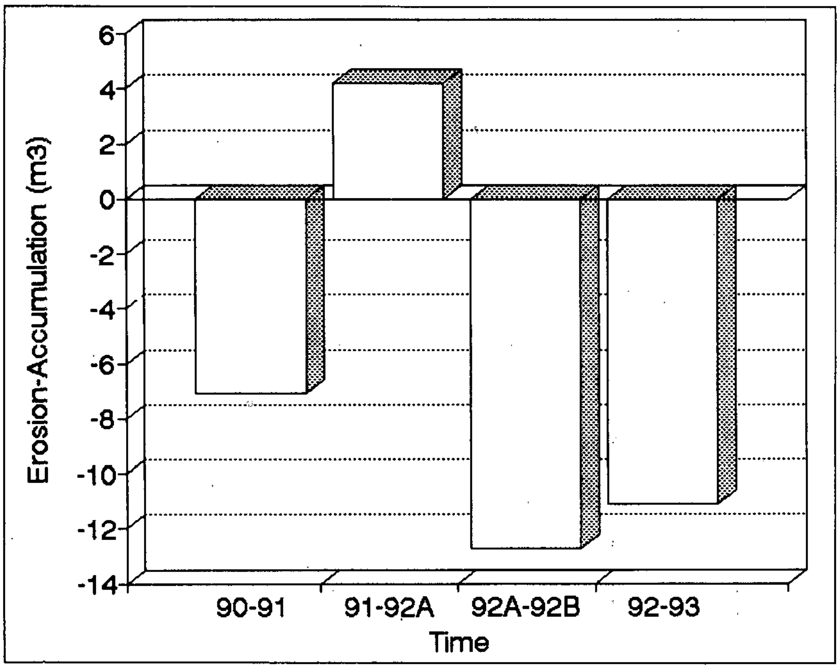

Figure 7. Variability of estimated gully erosion. Each column corresponds to the total of a hydrological year, except for 92A-92B, which is the result of two rainstorms in August.

Linear regression analysis between maximum discharge and rainfall shows best relationship with I-30 (R=0,5, significant at the 0,0000 probability level). Therefore, I-30 data during the study period (Table 2) are compared with average annual frequencies of the meteorological station in Cáceres. During 1990-91 and 1992-93 no I-30 > 20mm/h was observed, which has an a avevage frequency of 1,7. However, 1992-93 had a higher number of moderate intense storms. If the four rainfall events of August 1992 are not considered, the year 1991-92 had a below average number of intense rainfall.
The above described indicates a relationship between rainfall, discharge and gully erosion, and hence suggests that the dominant process of gullying is due to the erosive force of flowing water. In addition, bank collapses have been observed, which were due to heavy lateral incicion of the channel bank.
However several limitations of interpretation exist, es-

the vegetation cover on hillslopes may increase runoff production and therefore increase peak discharge. Also vegetation cover along the gully channel was reduced which may have increased erosion. Though the latter seems less important as high erosion occurs on channel banks, which are nearly void of vegetation. The heavy rainstorm

On the other hand, we did not experience a humid year with high precipitation totals during succesive events, nor any exceptionally high rainfall amounts (Table 2 and 4), which might produce even higher gully erosion than the observed one.

Average rate of erosion in the channel is 9,41 m^3 or 0,27 m^3/ha. On the basis of the cross-sections the total volume of sediment eroded in the channel was calculated. Taking 9,41 m^3 as the average rate of erosion, it is esti-

Table 1. Discharge and rainfall of Guadalperalón catchment during the hydrological years 1991-92 and 1992-93. PTOT - rainfall total of event, I-10 and I-30 - maximum 10-minute and 30-minute intensity, Q - discharge total, KOEF - runoff coefficient, Q-max - maximum discharge.

DATE	PTOT (mm)	I-10 (mm/h)	I-30 (mm/h)	Q (m^3)	KOEF (%)	Q-max (l/s)
30/03/92	15,6	25,2	12,4	88,6	1,6	49,5
01/12/91	30,2	9,6	7,6	228,6	2,2	41,9
02/04/92	10,1	8,4	7,9	49,1	1,4	17,0
19/02/92	19,6	4,8	4,8	33,5	0,5	6,3
15/06/92	15,4	15,6	12,4	13,6	0,3	3,0
02/04/92	20,2	8,4	4,4	18,6	0,3	2,5
07/08/92	21,6	60,0	32,8	884,6	11,7	860,0
24/04/93	20,4	24,0	14,2	280,7	3,9	295,0
29/10/92	15,8	28,8	14,4	347,2	6,3	226,3
26/09/92	26,4	19,2	12,8	215,9	2,3	86,9
14/04/93	7,8	33,6	14,4	145,4	5,3	74,2
19/10/92	19,0	9,6	8,0	198,7	3,0	73,5
16/10/92	12,2	16,8	9,2	79,7	1,9	69,3
11/10/92	9,2	15,6	5,4	149,4	4,6	62,2
15/12/92	10,8	10,8	8,8	42,4	1,1	28,4
03/05/93	9,2	20,4	16,4	59,3	1,8	24,1
26/05/93	9,4	21,6	10,4	49,1	1,5	17,0
04/12/92	9,4	22,8	12,4	38,7	1,2	17,0
19/12/92	10,2	8,4	7,2	34,9	1,0	14,7
30/04/93	13,8	4,8	4,8	27,1	0,6	2,7

mated that the present gully could have been formed during the last 70 years.

Monitoring of soil loss indicates a fairly low erosion rate, with about 300 g per metre hillslope. Parts of the catchment were cultivated in the past and abandoned about 30 years ago. It is thought that slope erosion has been decreasing with abandonment. Consequently the amount of sediment delivered to the channel decreased. On the other hand, overland flow production is probably equally high, because no increase of the infiltration capacity of the shallow soils can be expected. It is therefore possible that gully formation, or activation of erosion of an existing channel, occurred during the last decates in relation with land abandonment, because a decrease of sediment concentration produces an increase of the erosive force of flowing water. This assumptions seems possible as it can be assumed that the gully is of fairly young age.

CONCLUSIONS

Spatial variability along the studied channel section is high with maximum erosion occurring in an area with active headcut retreat.

The results indicate that the most important gully process in the area is related with the erosive force of flowing water produced by high magnitude and low frequency rainstorms. Gullying is therefore similar to the processes described by LEOPOLD et al. (1964).

At present time the channel is actively eroding. It is possible that gullying was initiated during this century, with an increase of channel erosion due to abandonment of cultivation.

Variation of total erosion with time can be related to variation in rainfall characteristics. However giving the high variability of precipitation in the area, the calculated average rate of erosion (0,24 m^3/ha) can only be considered a crude estimate.

Table 2. Frequency of rainfall amount and rainfall intensity in the study catchment. I-10 and I-30 is 10-minute and 30-minute maximum rainfall intensity, Day - daily rainfall.

	>50	40-50	30-40	20-30	10-20	5-10	0,2-5	Total	Year
I-10				1	6	19	67	93	90-91
(mm/h)	1	1	1	3	6	13	56	81	91-92
			1	5	12	10	57	85	92-93
I-30					6	9	78	93	90-91
(mm/h)			1	1	5	8	66	81	91-92
					9	12	64	85	92-93
Day		1		2	11	9	70	93	90-91
(mm)			1	2	10	13	56	81	91-92
				2	11	15	57	85	92-93

Table 3. Annual rainfall characteristics at Cáceres meteorological station (hydrological years 1907/08 - 1992/93), and annual rainfall during the study period.event of August 1992, producing heavy gullying, coincide with lowest herbaceous ground cover. Since it is the only event of such magnitude during the three year period, no conclusions can be made yet.

		Year	Rainfall (mm)
Mean	511		
Median	495		
Minimum	247		
Maximum	981	1990-9	477
Lower quartile	399	1991-92	407
Upper quartile	597	1992-93	380

Table 4. Daily rainfall and 30-minute intensity for different recurrence intervals.

Recurrence Interval (years)	Daily Rainfall (mm)	30-minute intensity (mm/h)
5	52,1	34,8
10	61,8	40,8
20	71,2	46,5
30	76,8	50,1
50	83,7	54,3
100	92,5	59,8

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Modelling of hydrological and hydrochemical variability under environmental change impact

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ABSTRACT

Climate change inducing significant variations in the seasonal dynamics of the water cycle and hydrological extremes and land use change as one of the anthropogenic effects are among the major driving forces of global environmental change. Large and meso-scale river basins represent an appropriate unit to study global change impacts on water quality and quantity at the regional scale and to estimate water, material, and biogeochemical budgets. The Elbe river basin is used as test region for the application of the Nested Watershed Approach based on the Two-Domains-Modelling-Concept to test, modify, and validate models, parametrization schemes, and up- and downscaling techniques. In order to achieve the objectives of the project a Hydrological Modular Simulation System for catchment basin studies is developed, based on a Geographical Information System (GIS) and the Modular Modelling System (MMS). Within this system a number of hydrological and hydrochemical simulation models and model components are tested and modified with special attention to regional approaches for the analysis and modelling of hydrological, biogeochemical, and hydrochemical processes.

INTRODUCTION

Processes of the hydrological cycle (runoff, evapotranspiration, infiltration, etc.) depend strongly on climate characteristics, their seasonal variation, and long-term changes. Climate change may lead to significant variations in seasonal dynamics and spatial distribution of water resources, resulting in water scarcity or oversupply and surface water deterioration. On the other hand, it is well established that land use changes are among the major driving forces of global environmental change. Changes in land use influence evapotranspiration and mineralization of soil organic matter and river runoff, which have clear feedback effect on the climate.

Watersheds are important as integrators of many forces, including climate. Still knowledge about the dynamics of dominant processes in watersheds is rather limited. Large and meso-scale river basins with their natural boundaries and hierarchical structure represent an appropriate unit for the study of climate change impact at the regional scale. Our approach is to develop a modelling system for analysis and modelling of climate change impact on hydrology and water quality for meso-scale river basins. The German part of the Elbe river and some of its tributaries have been chosen for case studies.

Previous efforts in watershed modelling were concentrated mainly on developing either continuous-time spatially lumped models or single event spatially distributed models. The availability of GIS tools and more powerful computing facilities make it possible to develop distributed continuous time models, based on available regional information.

Our recent and previous research on water quality affected by nonpoint source pollution (NPS) has been focused either on agriculture nutrient (N,P) pollution causing eutrophication problems, or areas exposed to acid rain (Krysanova et al, 1989; Müller et al., 1993). Consideration of potential climate impacts is a new challenge for research on changes of fresh water supply.

Objectives of the Study

The project is focused on investigations of hydrological and hydrochemical processes and structures in catchments that are parts either of semi-natural not intensively used ecosystems or are embedded in forestry and agriculture landscapes. The project includes four main closely interlinked tasks:

(1) Analysis and modelling of hydrological processes, and their interaction with other processes (ecological, meteorological, etc.) at different scales,
(2) Analysis and modelling of biogeochemical and hydrochemical processes, determining fresh water quality,
(3) Development and test of up- and downscaling methods in the context of a nested multiscale watershed approach,
(4) Providing of tools for coupling the terrestrial water cycle with atmospheric circulation processes, taking into account existing feedback mechanisms,
(5) Development of a Hydrological Modular Simulation System HYDROSIM) (Fig.1) for regional and catchment basin studies of climate change impact.

DESCRIPTION OF THE STUDY AREA

Regions and Scales of Application

Protection of water resources requires international efforts, since usually rivers are not limited to any administrative boundaries. In the frame of the Global Energy and Water Cycle Experiment (GEWEX) the region of the BALTIC SEA with the drainage basins of all inflowing rivers, as well as the Elbe and Weser river basin, have been selected as the primary study area in Europe. This study is called BALTIC SEA EXPERIMENT (BALTEX). A mesoscale atmospheric model is intended to be developed for the BALTEX model area at the Max-Planck-Institute of Hydrology in Hamburg, using a regular grid of 15 to 20 km mesh size. This model will be run in a semi-coupled manner with the Hamburg General Circulation Model (GCM) ECHAM to provide information on present and potential future climate scenarios at a regional scale. The output of the model shall be used as input to the model of the Elbe river as part of the BALTEX study area.

The second international programme the project is closely related to is the International Geosphere-Biosphere Programme, especially its core project "Biospheric Aspects of the Hydrological Cycle (BAHC)".

Early facing of problems affecting river systems led to the establishment of a number of international commissions. One of them is most important for our case study: The International Commission for the Protection of the Elbe / Labe river.

Furthermore, the project is related to the research priority programme of the Deutsche Forschungsgemeinschaft (DFG) on "Regionalization in Hydrology". Collaboration with this programme concerns a number of aspects including GIS tools and data exchange.

Within the Elbe river basin smaller subbasins are selected for finer scale studies. One such subbasin is the Stör river basin north of Hamburg, for which the Ecosystem Research Center of the Kiel University has provided the data required for modelling. It is also planned to collaborate with ZALF, Müncheberg in the North-East Study Area, which covers the north-eastern part of the Elbe river basin, i.e. the upper Havel basin.

The Elbe river drainage basin and environmental problems

The Elbe River is one of the longest rivers in Europe (1092 km), with a drainage basin of 148268 km^2 and 24,9 million inhabitants. About 2/3 of the drainage basin belong to Germany and 1/3 to the Czech Republic. Since large parts of the river systems are located in regions of not more than 1500 m elevation, the river discharge is characterized by winter and spring high water periods. The upper, mainly Czech part of the river is dominated by weirs and dams, whereas the middle part can be considered as a semi-natural river system. Its final 142 km are effected by tide processes. Potential erosion increases downstream in glacially formed landscapes of the north German lowland. The combination of these natural factors with river management actions caused a depletion of low level water tables in the lower parts of the river during the last decades (Simon, 1993).

Compared to the discharge behaviour of the Rhine valley it is obvious, that in the Elbe natural upstream buffer storages of glacial snow packages against both flood discharge and low flow are missing due to the lack of high mountain regions. Thus the span between monthly low flow and high flow in the Elbe is 1:21 as compared to 1:1,75 in the Rhine valley.

Agriculture areas that cover about 56 % of the total area of the drainage basin represent one of the most important sources of pollution (Simon, 1993). The Elbe and its tributaries are intensively used for fresh water supply (drinking water, irrigation and industrial process water).

Due to ineffective sewage water treatment from municipal and industrial sources, and lack of nonpoint source pollution control (especially in the Eastern part of Germany and the Czech Republic), fresh water supply is severely limited and requires extensive waste water treatment. Dominant pollutants are heavy metals, nutrients, chlorinated hydrocarbons, and other organic pollutants.

Though efforts to reduce pollutant loadings to the Elbe river during the last four years have been quite successful, the Elbe is still one of the most heavily contaminated water courses in Europe (Mobs, 1993; Reincke, 1993). On the other hand, the river system and its riparian areas include unique biotopes for various plant and animal species that are threatened of extinction. Many large Czech and German nature preserves belong to the Elbe drainage area. The UNESCO-biosphere preserve "Middle Elbe" has been established to protect one of the largest continuous alluvial forests in Central Europe.

Hydrological temporal variability and spatial changes

What are expected changes in hydrological characteristics in time and space for the Elbe basin? First of all, it is clear that critical situations can be caused by the increasing probabilities of floods and drought, and increase of water demand in a warmer climate. Climate change may influence hydrological extremes, affecting all their parameters - intensity, magnitude, frequency. For example, increased winter precipitation can cause winter floods, whereas increased evapotranspiration in summer can cause soil moisture deficits and negative consequences for agroecosystems.

Changes of air and soil temperature may lead to changes in both snow cover, evapotranspiration (ET), atmospheric circulation, and sea water levels. The direction of some of the consequences can not be estimated: e.g. an increase of ET may lead to an increase of cloudiness, subsequently causing an ET reduction. In addition, not all the quantitative aspects of the effects are yet known.

One should be aware of the possibilities that the spatial and temporal patterns of precipitation events may change, rather than just the mean values, and that stable periods of either heavy rain fall or drought may increase, leading to a rise of seasonal extremes of water availability. Further more, runoff amounts may stay on the same level as a result of a combined increase in precipitation and evapotranspiration.

Significant hydrological vulnerability in the Elbe region, especially in its north-eastern part, is due to intensive water use all along the river and comparatively low yearly amounts of precipitation (between 450 - 500 mm/yr.), causing severe problems both to agriculture (irrigation) and for stability of drinking water supply. These problems are particularly severe around Berlin (about 4 Mio. inhabitants) since this area is heavily depending on fresh water supply from Elbe tributaries. It has been stated several times, that the water table in the region of Havel and Spree may not fall below a certain threshold to insure about 60 % of the water supply as well as energy production for more than 4 Mio. people in and around the metropolis.

In addition, the area has been effected by open cut mining during the last decades leading to 3300 km2 of groundwater depletion. During the last 3 years mining activity has been reduced causing that less ground water from deep aquifers has been spilled into the water courses. It has been estimated, that groundwater deficits in this region amounts to about 13 billion m^3.

The period from 1989 to 1992 has been the second driest in this century (Simon, 1993) with respect to mean values of yearly discharge during the century, and 1991 is on the fourth place of dry single years in the Elbe during the last 90 years. During those years runoff was about 60 % of the mean yearly values at most of the international gauge stations. In spring 1994 the north eastern parts of Germany suffered from an extensive rainfall period leading to flooding events in various parts of the Elbe basin lowland, which caused huge damages to agriculture and housing. Assessment of this latest events may include satellite pictures taken by the environmental mission shuttle "Endeavor".

In addition to meteorological fluctuations leading to river runoff change, changes in land use and agriculture practices are of great importance for water quality. The nitrate load of the Elbe steadily increased during the period 1970 to 1990. Reduction of agricultural land use - after the German unification - caused by economic restrictions and restructuration in the former German Democratic Republic was primarily focused on the less productive soils. While the oxygen concentration in the Elbe has increased, inorganic nitrogen increased as well, due to the time delay between nitrogen application and release from soils to tributaries (Meissner and Rupp, 1993).

METHODOLOGY

Nested multiscale watershed approach

River networks represent a useful organizing concept in subdividing land surfaces into subunits, i.e. drainage basins of tributaries, where all water flows are integrated over the entire basin area. This is the key of the watershed approach, which can be applied in a nested way and used to test available models and parametrization schemes at different scales, as well as up- and down-scaling techniques for landscapes differing in their degree of heterogeneity.

The nested watershed approach takes into account the natural integrating behaviour of river basins for all surface and sub-surface water flows and related waterborne transports (sediments, nutrients, heavy metals, etc.). This approach will allow to test, modify, and validate hydrological models for land surface processes at different scales, for areas where data availability is clearly differing. It is especially important for nonpoint pollution modelling, as data availability for hydrochemical processes modelling is usually very scarce, and extrapolation from small representative sub-watersheds to the whole drainage basin is the only possibility to overcome these difficulties. This would allow to estimate water, material and biogeochemical budgets for watersheds of any size, where water and constituent discharge is measured at the closing river cross-section. These budgets can be used for basin related tests of the applied hydrological and hydrochemical models.

It is intended to model in sequence several subbasins of different size in the Elbe river basin with scale specific or scale-adapted hydrological models, starting from very small study areas, where detailed information on important processes and controlling geocharacteristics is available (catchment areas of a few km2) to ever larger watersheds, e.g.

- upper Stör and its tributaries,
- Spree and some of its tributaries,
- Elbe river basin downstream of Dresden,
- whole Elbe basin, including the Czech part.

Linkages to General Circulation Models and to regional atmospheric models will be provided.

Two domains modelling concept

The Nested Watershed Approach includes the application of the Two-Domains-Modelling-Concept (introduced by Becker and Nemec in 1987), which allows to decouple and model separately the "vertical fluxes domain" (all vertical water, energy, and other fluxes at the land surface - atmosphere interface) and the "lateral flows domain" (all lateral water flows and chemical matter transports across a reference area, e.g. a river drainage basin). This concept allows to apply different space resolutions and spatial discretization schemes in both domains and to directly couple the vertical fluxes domain model with atmospheric models. This means, while a high resolution scheme may be appropriate in modelling the vertical fluxes domain to assess land surface heterogeneity, much larger areal resolutions can be applied for modelling the lateral flows in river basins. In this context a major research subject will be to identify those hydrological model components (modules) and spatial discretization schemes which are best suited for regional impact studies.

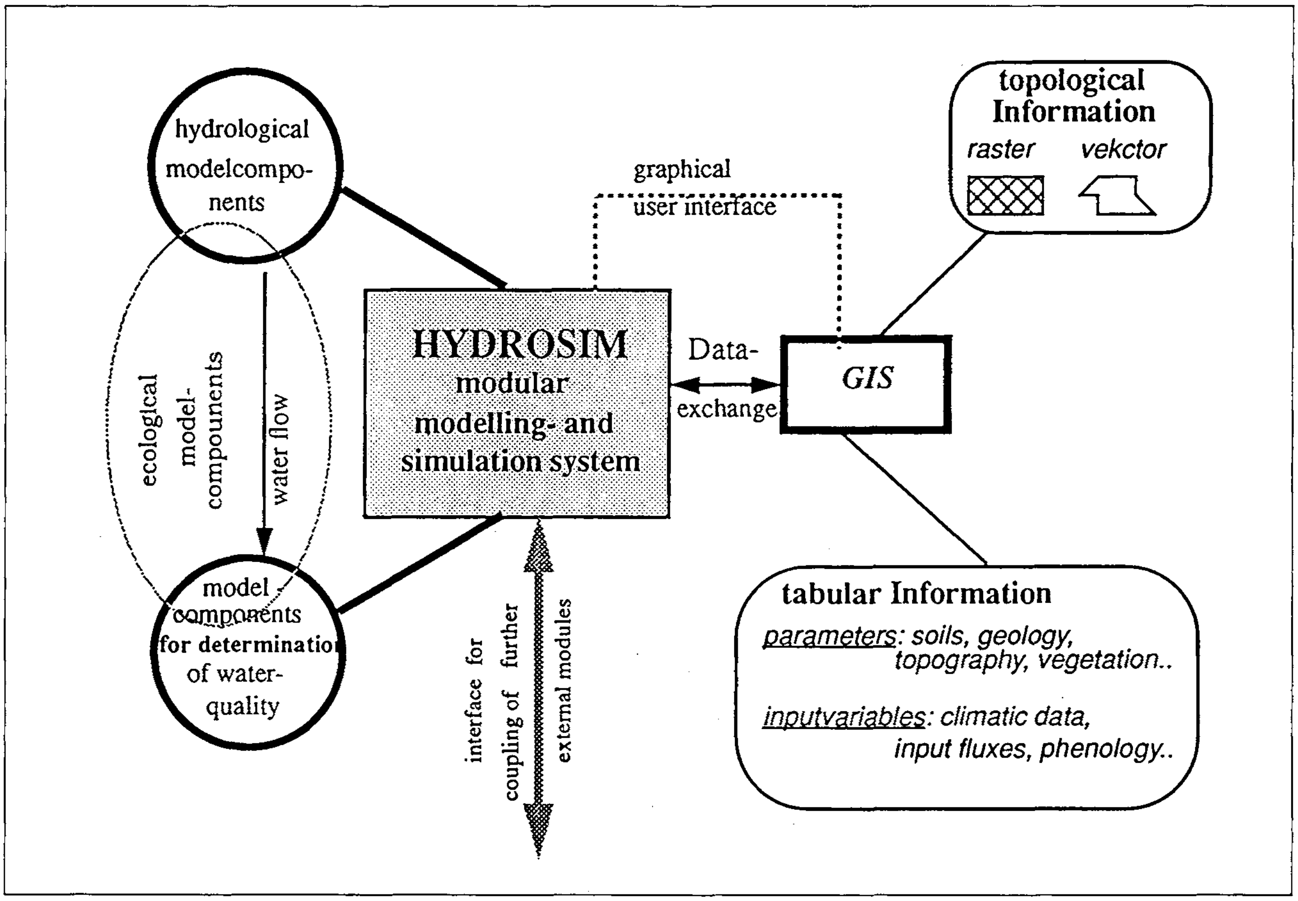
Downscaling from GCM

A classical scale of hydrological analysis refers to a moderately steep and homogeneous catchment in a certain climate. Downscaling from GCM models which is necessary to make use of GCMs for hydrological study, is a very serious computational problem, as the resolution of GCM models is too crude. Nevertheless, it is possible to couple them with hydrological models to use the output from GCM model as the input to watershed model. Some methods are being developed for this purpose.

In this context, the coupling of atmospheric and improved land-surface hydrological models is receiving increasing attention. Especially on the mesoscale level remarkable progress is expected from the coupling of atmospheric models with complex hydrological models for river basins. To contribute to this development is an important aim of our work.

Upscaling for estimation of NPS pollution

Nonpoint source pollution (NPS) is the most difficult problem in the water quality studies for large areas. Basing on our previous experience, we suggest to solve it through upscaling: physically-based balance simulation models at the micro-scale can give a possibility to assess Unit Area Loads (UAL) as dependent from a num-



ber of relevant driving parameters, like hydrological regime, soil structure, and agriculture practices. This can give a possibility to apply the UAL estimations at a higher watershed scale. In that sense, a regional approach provides an interface between the global scale with systems oriented goals and micro-scale with disciplinary oriented goals.

Models and Tools

In order to achieve the objectives of the project powerful tools and software systems are needed. The Geographical Information systems ARC/INFO and GRASS are widely used for spatial data processing, preparation of input data, analysis and graphical presentation of spatially distributed data (soil types, land use, vegetation cover, ground water level, etc.). In our study we intend to combine the existing tools and new submodels with the Modular Modelling System (MMS, Restrepo, 1993). For this purpose a number of hydrological and hydrochemical simulation models are tested and modified with special attention to regional approaches:

Hydrological Models

•Semi-distributed conceptual hydrological model EGMO ("Einzugsgebietsmodell - catchment modelling system" , Becker and Pfützner 1987)
•Distributed hydrological model MIKE - SHE ("System Hydrologique European", Danish Hydraulic Institute, 1993)
•Topography-based model "TOPMODEL" (Beven and Kirkby 1979)

Hydrochemical Models:

- Nonpoint source pollution model for field scale OPUS (Smith, 1993)
- Water quality model for agricultural mesoscale watersheds MATSALU (Krysanova et al., 1989; Krysanova, Luik, 1989)
- Integrated hydrology-water-quality models SWRRB (Arnold et al. 1990) and SWAT (Srinivasan and Arnold 1993)

These models and model components (resulting from our own research or received from other research groups) are tested and modified to choose the appropriate modules to be incorporated into comprehensive modelling systems for regions or river basins. In a first phase, only the hydrological and hydrochemical processes are studied. Later on, ecological submodels and socio-economic submodels (representing driving forces of land and water resources use change) will be developed and included into the simulation systems. On a long term basis, the development of a comprehensive simulation system is envisaged, which can be used for different impact studies in any region with different data availability.

CONCLUSIONS

The area of the German part of the Elbe drainage basin has been affected by both changes in discharge and hydrochemical loads during the last years, and nested watershed modelling can be useful for the prediction of hydrological variability and changes in hydrochemical flows. The development of a nested watershed modelling approach and a comprehensive simulation system for meso-scale watersheds can be useful for different regional impact studies, which are most appropriate for decision makers, as

- such a modelling system can help to bridge the gap between the large and small scale modelling,
- coupling GCMs with hydrological models can improve GCM predictions at the regional scale,
- focus on critical situations may improve estimation of probabilities of extreme hydrological events under global environmental change,
- the nested watershed approach could be the only possibility to evaluate water quality, namely nutrient and pollutant flows with surface and subsurface runoff.

An improvement in the description and modelling of the global environmental change may not be limited to climate change but has to take into consideration various anthropogenic effects, particularly in densely populated regions like Central Europe.

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Effect of the clearfelling on the water quality: Example of a spruce forest on a small catchment in France

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SUMMARY

This paper presents the variation of the hydrology and the water quality of a spruce catchment, located at Mont-Lozére (France), in a mediterranean mountain climate area , in relation to the forest status during 12 years (1981-1993). Four situations were successively examined : healthy forest (1981-84), declining stand with pest (1984-87), gradual clearfelling (1987-89) and reforestation (1989-93). An undisturbed beech catchment was used to provide reference values.

In the hydrological budgets, the P-Q value (as ETR) was slightly higher in the spruce catchment than in the beech one during the first period and decreased progressively in the following ones as a consequence of: (1) the declining stand of the forest and (2) the clearfelling.

No change was observed for cations, and NO_3 concentrations remained were very low during the whole period in the streamwater of the beech catchment , in relation to the steady state of that ecosystem. Iin the spruce catchment,the concentrations of cations and NO_3 were always higher, and increased slightly during the disease. During the clearfelling, NO_3 was strongly related to Ca and Mg. Six months after the reforestation, NO_3, Ca , Mg concentrations were respectively 11,9 , 2,6 and 3,6 higher than at the beginning of the clearfelling.They returned to previous values at the end of 1993.

The Input-Output budget of cations presented a continuous storage in the beech catchment and simultanously a permanent release in the spruce catchment . The mean loss, -expressed as the denudation cation rate, in keq.ha-1.year-1 was as follow: -0,41 (1981-84), -0,65 (1984-87), -1,60 (1987-89) and -0,82 (1989-93).The leaching was observed during more than 6 years after the clearfelling, resulting probably from the duration of the drought period , and from the mineralization of the remaining important organic matter compartment.

1. INTRODUCTION

In many european countries, large areas easing acidity and aluminium leaching (Likens et al.,1979, Stevens and Hornung,1987, Hornung et al.,1990), particularly with aluminium-rich soils. (Hultberg,1985, Adamson et al., 1987) .

In low air-polluted areas, such as in the south of Europe, where this process is of less importance, studies on the clearfelling effects have just been developped, in France (Didon-Lescot et al..,1992), or in Spain (Sabate and Gracia, 1993).

In the low air-polluted area of Mont-Lozère (Durand et al..,1991), three experimental catchments are monitored for biogeochemical functionning studies (Dupraz et al..,1984, Lelong et al..,1990).Two of them are forested catchments, one is covered by a beech coppice, the other by a spruce plantation. The spruce forest has progressively declined from 1984 to 1987. So, there was an opportunity to describe the evolution of the biogeochemical processes during the ecosystem change. The water chemistry of the spruce catchment was examined during the four successive situations of the forest, i.e, healthy forest (1981-1984), followed by 3 years of declining stand (1984-1987),then gradual clearfelling (1987-1989) and finally reforestation (1989-1994).

The aim of this paper is to present the results concerning the hydrological budgets and the evolution of cations and NO_3 concentrations in the streamwaters, as well as the Input-Output budgets in the two forested catchments. The hydrochemical and hydrological characteristics of the beech catchment are presented to provide reference values of an undisturbed ecosystem.

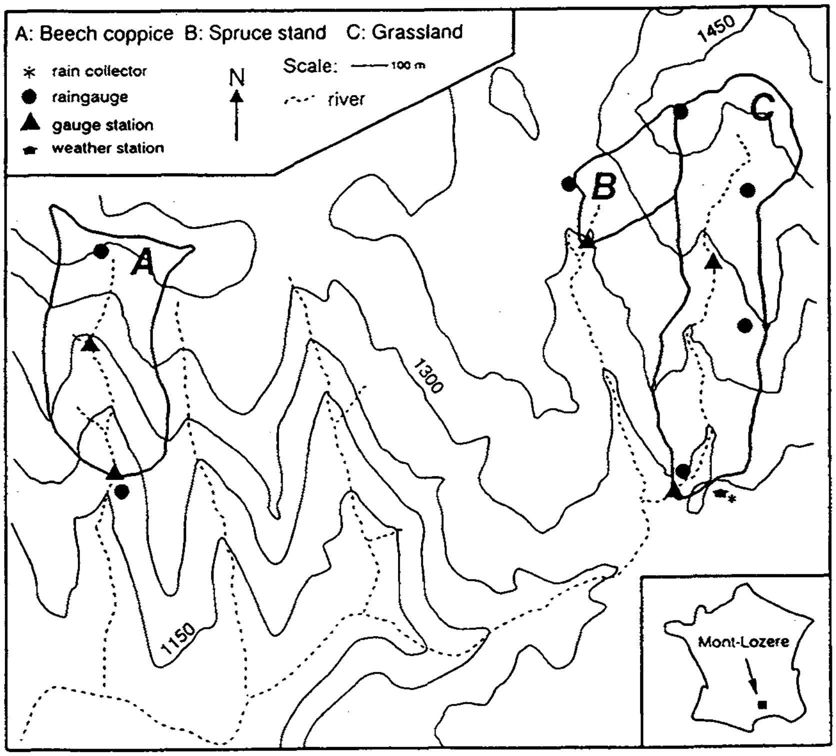

Figure 1. Site location and equipment

2. METHOD

2.1. The study area :

The Mont-Lozere research catchments (ERBFR110) are located in the National Park of Cevennes (south of Massif Central), in a M.A.B sanctuary with slight anthropic perturbation (Lelong et al..,1990) . They are 80 km distant from the Mediterranean Sea , and they range from 1150 to 1500 m of elevation.They have small surfaces, i.e. 0.20 km^2 for the spruce catchment and 0.54 km2 for the beech catchment (Figure 1).

The parent bedrock is a granite, and they have similar types of soil (dystrochrepts, i.e. acid and humiferous soils).The climate is mediterranean (rainstorms in automn and spring , possible long drought period in summer) with mountain characteristics (annual mean temperature of 6°C, and snowpack varying from 10 to 25 % of the total precipitations).

The beech, covering 80% of the surface of the beech catchment, is a coppice of low production unexploited since more than 60 years, characterized by a large number of trees (4270 boles/ha) (Hanchi, 1994). The spruce covered 85% of the second catchment, where there was 5 % of mountain pine. In the coniferous catchment, the spruce stand has been gradually attacked by a bark-beetle (Dendrochtonus sp.) probably after the drought of 1983. A gradual clearfelling was prescribed during the summers of 1987,1988,1989, and trunks of commercial value were extracted by logging mecanically. Finally young conifers were planted in automn 1989 after removing and gathering of the tree debris stored in strips parallel to the slope.The wooden area is presently restricted to the extreme-top of the catchment and to a small plot of declining trees maintained for studying the internal cycling of nutrients.

Table 1. Precipitacions (P), Runoff (Q) and P-Q (in mm). Mean for period 1 to 4 (bold); mean 1981-93 and standard deviation (italic)

	Beech				Spruce		
	P	Q	P-Q		P	Q	P-Q
81-82	1630,4	1162,9	467,5	period 1	1594,1	1210,7	383,4
82-83	2237,1	1431,7	805,4	(1/7/81	2475,0	1554,7	920,3
83-84	1440,0	877,8	562,2	30/6/84)	1506,1	928,6	577,5
	1769,2	**1157,5**	**611,7**		**1858,4**	**1231,3**	**627,1**
84-85	2137,6	1379,4	758,2	period 2	2189,2	1484	705,2
85-86	1374,7	968	406,7	(1/7/84	1469,2	1044,4	424,8
86-87	1906,2	1219,1	687,1	30/6/87)	1954,4	1297	657, 4
	1806,1	**1188,8**	**617,3**		**1870,9**	**1275,1**	**595,8**
87-88	2428,7	1639,3	789,4	period 3	2665,4	1985,5	679,9
88-89	1238,5	698,2	540,3	(1/7/87	1349,7	931,6	418,1
	1833,6	**1168,8**	**664,8**	30/6/89)	**2007,6**	**1458,6**	**549,0**
89-90	1339,6	680,4	659,2		1431,3	798,3	633,0
90-91	1394,8	938,3	456,5	period 4	1446,61071	375,6	
91-92	1431,4	688,4	743,0	(1/7/89	1490,9	798,9	692,0
92-93	2028,6	1225	803,6	30/6/93)	1954	1286,1	668,4
	1548,6	**883,0**	**665,6**		**1580,8**	**988,6**	**592,2**
Mean	**1715,6**	**1075,7**	**639,9**		**1793,9**	**1199,2**	**594,6**
S.d.	**409,5**	**317,0**	**146,9**		**447,3**	**350,2**	**164,8**

2.2. Method.

The equipment of the catchment is described in previous papers (Dupraz et al., 1984, Lelong et al., 1990).It includes , in the two forested catchments:

- 2 rain gauges with automatic recorder (CR2M-PLA),
- one bulk precipitation collector devoted to chemical analysis.
- one gauge station with a V-notch-weir and 2 stream-level recorders : one AOTT R10 and one with ultra-sonic and temperature probes connected to a data-logger (CR2M-LUS-I) for streamflow measurements.
- one water sampler .During the major hydrological events , especially during the study of the clearfelling , manual water samplings were performed.
- soil lysimeters and throughfall devices have completed the field equipment from 1986 (Vannier et al, 1993).

Chemical analyses have been performed as quickly as possible, alkalinity by Gran method, NO_3 and NH_4 by colorimetry, and cations by atomic absorption spectrometry. Bulk precipitations are collected on an event basis (about 30 samples in a year), and stream water before and after the peak flow and during the baseflow .

Fluxes of solutes were computed using the combinaison of linearized hydrograph and instantaneous concentration, giving a "chemograph" (Lelong et al.,1990). Relatives error values are +/- 10% for input and +/-5% for output . After 12 years of study, the way of calculation has been simplified and in this study, the annual input and output of elements have been calculated by multiplying the Volume Weigthed Concentration (VWC) by the annual mean value of the rainfall or streamflow average (Hanchi,1994).

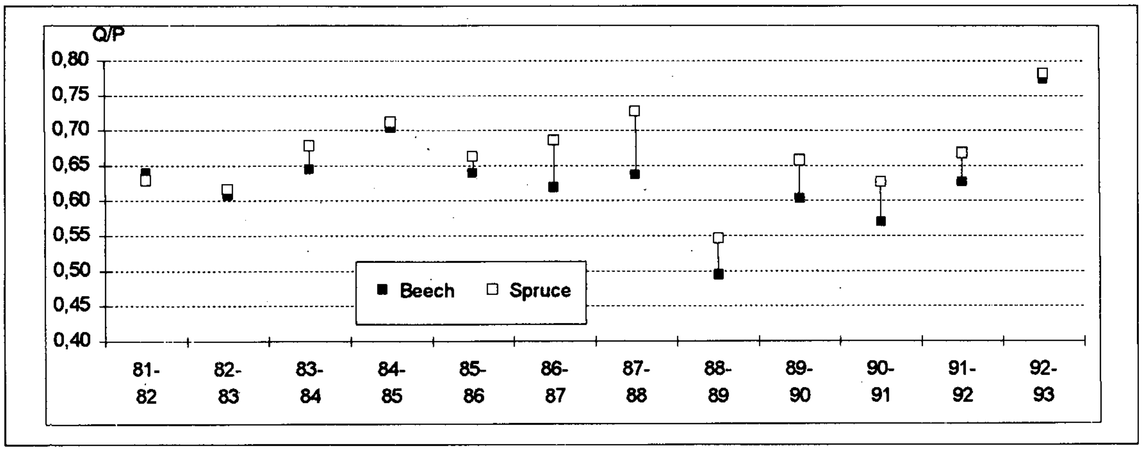

Figure 2. Variation of the early Q/P ratio for the two forested catchments. Period 01/07-03/06

3. RESULTS

3.1. Hydrological behaviour.

The hydrological budgets presented on Table1 are computed on the basis of an annual cycle (01 july to 30 june). During the 12-yrs period of study, the average of annual precipitations was lower in the beech catchment (1715,6 mm) than in the spruce one (1793,9 mm) .The greatest discrepancy between the two catchments was observed in 1982, and is related to a storm event (500 mm in 48 hours) that occurred particularly in the spruce catchment , which is more influenced by mediteranean storms (Dupraz et al..1984).

The interannual variations are important since precipitations have varied from 1094,0 mm in 1985 (civil year) which was probably the driest year of the century, to 2465,9 mm in 1984.In comparison with the long term average, established at the Villefort station by the National Meteorological Survey, the studied period may be considered as modal with contrasted, wet (1992-1993) or dry pluriannual periods (1989-1990-1991).

As it was noted before (Dupraz et al.., 1984, Lelong et al, 1990) the runoff was always higher in the spruce catchment. This appears as a consequence of more abundant precipitations and of different pathways of water circulation. A small peat bog area existing at the outlet of the spruce catchment could store more water, giving a small but permanent flood in summer. By contrast, the runoff was very low in the beech catchment and during the very dry summers, as in 1985 and 1989, the stream dried up.

The water budgets (Table 1) are calculated for a pluriannual scale. As the ground water storage is probably low in comparison with P and Q, the expression P-Q can be reasonably considered similar to ETR. In the beech catchment, the mean ETR was 674 mm for the whole 1981-1993 period, whereas in the spruce catchment , ETR was slightly higher (701 mm) during the first one (reference period). There was a progressive increase of runoff in the spruce catchment from the secund to the third period , that explains the increasing of the Q/P ratio difference observed between the two catchments (figure 2), and the maximum was found in 1987-88, the first year of the clearfelling.This could be explained not only by the reducing of the physiological activity of the trees but also by a change of the circulation speed of the soil water (Cosandey,1993) .

3.2. Water quality

3.2.1. Time variations of cation concentrations in the streamwater.

The variations with time of the element concentrations and the discharge in the streamwater are presented in the figure 3. The table 2 shows the Volume Weigthed Concentrations (VWC) for cations and NO_3.

In the beech catchment, the dominant cations are Na , Ca and Mg that represent 43%, 32% and 20% of the cationic charge respectively. The fact that Ca is dominated by Na, is probably related to the presence of Ca-depleted soils (Durand et al.,1991).

The relationship between Ca and Mg is better in the spruce catchment (g=0,609Ca+1,287; r2=0.982), (Mg=0,402Ca+9.240; r2=0,827) than in the beech catchment. Moreover this strong relation between the

Table 2. Semestrial Volume Weighted Concentrations (VWC, in µeq/1) and Runoff (Q, in mm) for cations and NO3 in the streamwater of the beech (left) and one of the spruce catchment (right). 1981-1993. 981-1=01/01/-30/06/1981; 81-2=01/07-31612/1981)

	Q mm	Ca	Mg	K	Na	NH4	Cat.	NO3	Q mm	Ca	Mg	K	Na	NH4	Cat.	NO3
81-2	513,3	37,3	24,3	4,9	48,4		114,9	3,8	522,8	63.2	40,2	5,6	52,7		161,7	1,6
82-1	649,6	35,5	22,9	3,9	44,8		107,1	1,8	687,9	60,6	38,3	5,5	50,3		154,7	11,0
82-2	683,8	34,3	23,7	7,3	46,0		11,3		791,4	61,2	39,5	7,9	49,9		158,4	
83-1	747,9	39,6	24,1	4,5	47,2		115,5		763,3	60,3	38,8	6,2	50,8		156,8	
83-2	132,5	40,7	26,6	6,4	56,8	0,5	131,0		140,3	68,3	44,1	6,9	56,3	0,3	176,0	
84-1	745,3	40,8	26,0	4,7	47,4	1,2	120,1		788,3	72,0	46,7	6,5	55,2	6,5	186,8	
84-2	921,5	39,7	24,5	7,9	48,6	0,1	120,7		1015,3	65,8	43,0	11,0	51,3	1,0	172,1	
85-1	457,9	35,7	24,7	4,1	47,9	0,0	112,4		468,7	61,7	43,4	4,4	54,6	0,2	164,2	
85-2	32,4	49,7	31,3	7,1	62,9	0,0	150,9		102,9	83,1	50,2	7,5	64,7	0,0	205,3	15,
86-1	935,6	52,9	29,4	5,2	57,0	0,3	144,7		941,5	78,2	47,4	7,4	63,1	0,0	196,2	11,0
86-2	601,9	48,0	28,9	7,3	61,9	2,3	148,5		687,3	77,1	44,8	10,8	67,9	4,6	205,3	15,8
87-1	617,2	43,5	25,3	4,9	62,1	2,7	138,4		609,7	77,7	42,9	6,7	63,1	2,0	192,4	28,4
87-2	635,1	40,8	26,1	5,1	58,1	0,7	130,8	0,7	846,9	95,5	60,1	12,0	66,7	2,1	236,4	94,5
88-1	1004,2	38,8	23,7	4,9	54,5	0,8	122,7	0,8	1138,6	102,5	61,7	8,5	71,1	0,8	244,7	157,6
88-2	253,8	41,2	26,0	5,4	59,0	1,7	133,3	2,0	524,8	114,6	72,6	12,2	74,6	2,5	276,5	170,8
89-1	444,4	53,7	29,7	4,9	53,6	0,6	142,6	2,1	406,8	160,5	102,7	11,7	80,5	1,6	357,0	238,9
89-2	250,6	46,6	29,4	10,1	58,4	1,6	146,1	1,9	366,9	194,1	111,9	19,7	84,1	3,5	414,0	304,0
90-1	429,8	45,8	28,5	5,6	58,4	2,0	140,3	1,8	431,4	203,8	132,6	13,1	91,0	1,5	442,0	337,7
90-2	294,4	40,9	23,8	4,3	61,0	1,0	131,0	0,6	402,3	193,5	117,7	11,2	91,4	0,7	414,5	309,3
91-1	643,9	46,3	25,1	4,7	52,4	1,4	129,8	2,3	668,7	191,0	101,9	14,1	76,7	1,2	384,8	292,4
91-2	184,3	53,8	30,3	5,9	65,3	1,0	156,3	0,8	272,5	178,5	107,2	11,1	77,6	3,3	377,7	243,4
92-1	504,1	52,1	26,2	4,1	59,9	59,9	1,2	1,1	526,4	145,9	77,4	7,9	69,4	0,6	301,1	192,5
92-2	626,2	40,3	26,3	3,0	44,8	1,3	115,6	0,5	738,5	104,3	57,6	10,0	53,7	1,7	227,2	109,7
93-1	598,8	39,7	25,4	2,3	54,5	1,2	123,1	1,2	547,6	86,9	52,3	3,8	60,4	2,5	205,8	82,0
93-2	695,6	37,9	23,4	4,0	55,3	1,2	121,8	0,4	924,0	70,4	44,8	14,0	50,6	2,1	181,9	42,6

two elements was not altered by the clearfelling.

The relative low K concentrations are subjected to seasonal variations,with a peak in automn, due to a rapid lixiviation of the decaying leaves and organic debris. NH_4 and NO_3 have comparable con-centrations, which remain low (0 to 2 µeq/l).

The highest values of the cationic charge (table 2) always occur after a long time of drought (1983 ,1985,1989,1991) .This can be explained by a strong mineralization of the forest floor during summer and automn when the soil moisture and temperature are still favorable to microbial activities. The leaching of the solutes occurs later at the beginning of winter once soils are water saturated (Durand et al., 1991). This process is especially obvious after the two driest automns of 1985 and 1989. By contrast, the rain storms that occur in automn and in spring, (figure 3) contribute to dilute the soil solution (exemples of spring 1988, and during the whole wet period of 1992-93).

The time variations of the cations are mainly due to the alternance of wet and dry periods, and are probably controlled by the importance of the precipitations. At a decennal scale, in the beech catchment, years of similar wetness as 1981 and 1992-93 are quite similar (Table 2).This could reflect the probable steady-state equilibrium of the forest (Lelong et al. 1990, Hanchi, 1994).

The fact that the cation concentration was higher in the spruce catchment (table 2), may be related to a greatest mobility of mineral anions. This could be partly a consequence of more important dry depositions collected

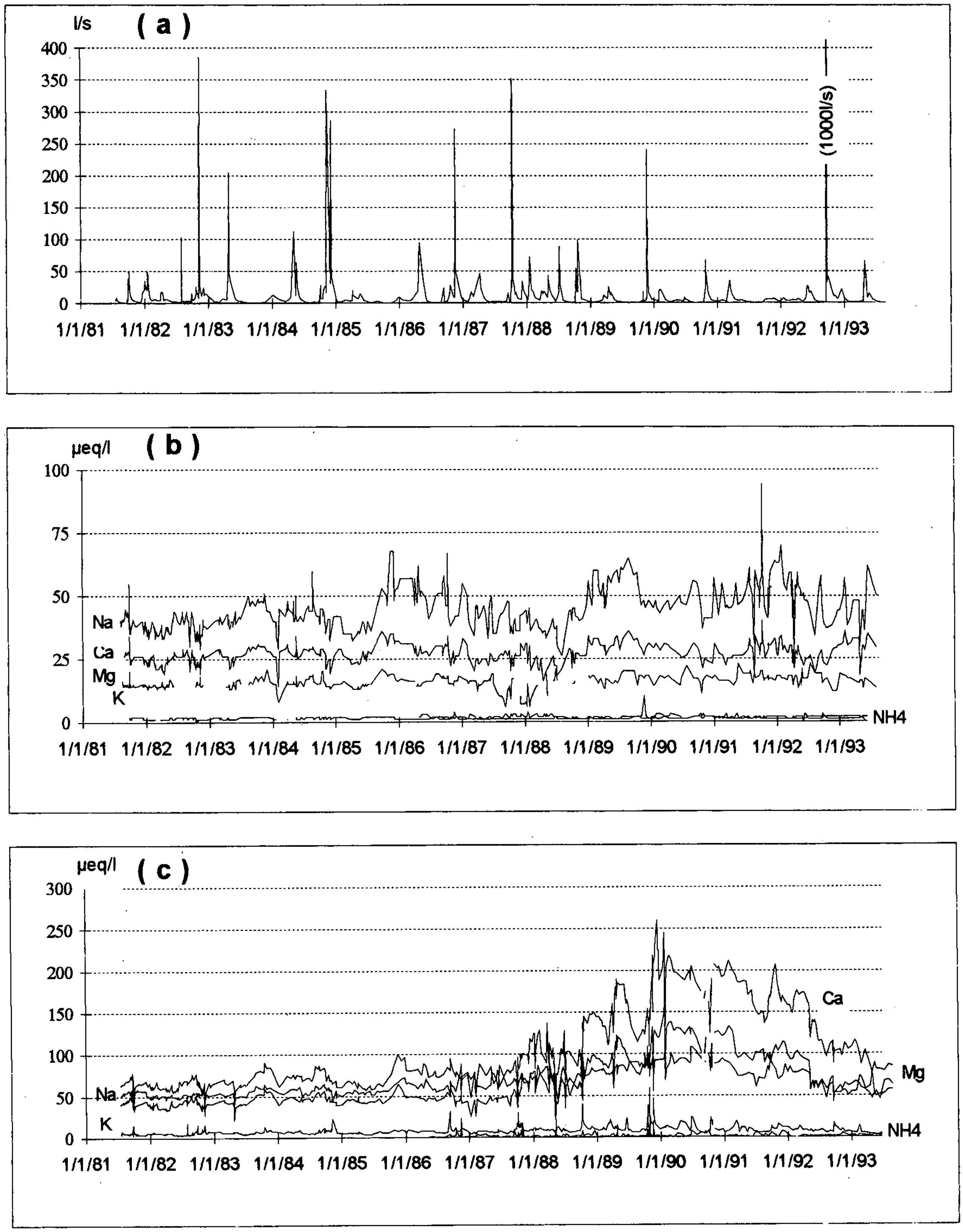

Figure 3. Time variation of discharge in the spruce catchment (a), cation concentrations in the streamwater of the beech catchment and the spruce catchment (c).

by the very efficient canopy of the conifers (Durand & al. 1991).

NO_3 increased slowly during the first and the second period and reached a VWC of 28,4 µe/l just before the beginning of the harvest (first semester of 1987, table 2).The increase was probably due to the declining stand of the trees, but it remained low and as NO_3 was not related to Ca (figure 4), we conclude that it had no effect on the leaching of this element at that moment (Didon-Lescot et al., 1992).

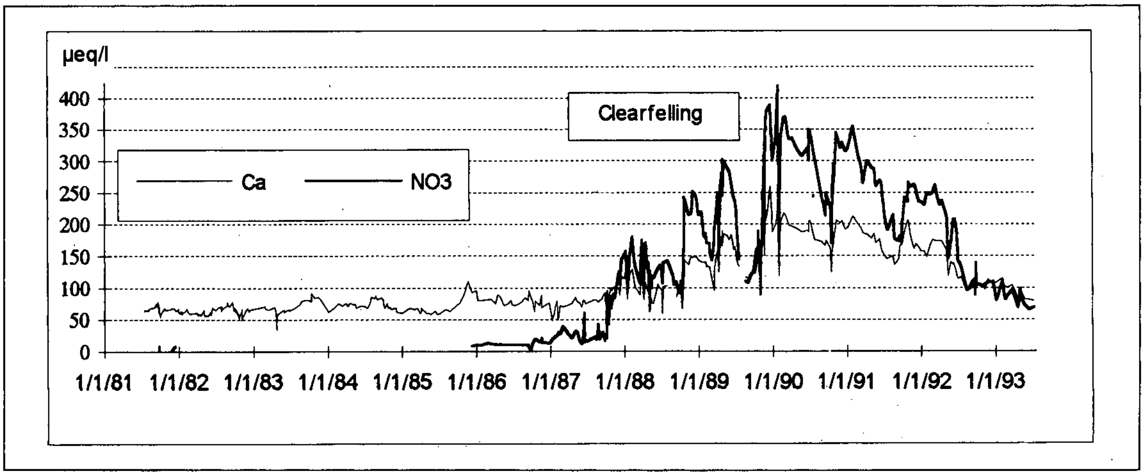

Figure 4. Time variations for Ca and NO3 in the spruce catchment. Period 1981-1993

3.2.2. Impact of the clearfelling of the spruce forest.

3.2.2.1. Time variations of concentrations from 1987 to 1993.

The clearfelling of the spruce was performed from summer 1987 (50% of trees have been cut) to summer 1989 (95%). Cations have increased gradually after the beginning of the disturbance (figure 3), strong seasonal variations appearing with lower concentrations in summer and higher in winter or spring.

The VWC (Table 2) for Ca and Mg have been multiplied respectively by 2,6 and 3,1 between summer 1987 and 1990. The maximum occurred in january 1990, 6 months after the end of the clearfelling and was of 245 µeq/l for Ca and 161 µeq/l for Mg. After early 1990, concentrations decreased regularly and reached the level before felling, 6,5 years after the beginning of the dis-tur-bance , at the end of 1993.

The Na increase was of 1,5 , with a maximum concentration in june 1990. At the end of 1993, concentrations were quite lower than in 1987. K showed high concentrations in automn such as during the important flood events of late november 1989, characterized by a peak at 41 µeq/l.

A higher value (59 µeq/l) was found at the peak flow of a very important flood, in september 1992 . NH4 which was in low concentration before the felling (0 to 5µeq/l) presented the same pattern as K. It appeared in the first flood events of automn 1989, and its maximum concentration was observed during september 1992.

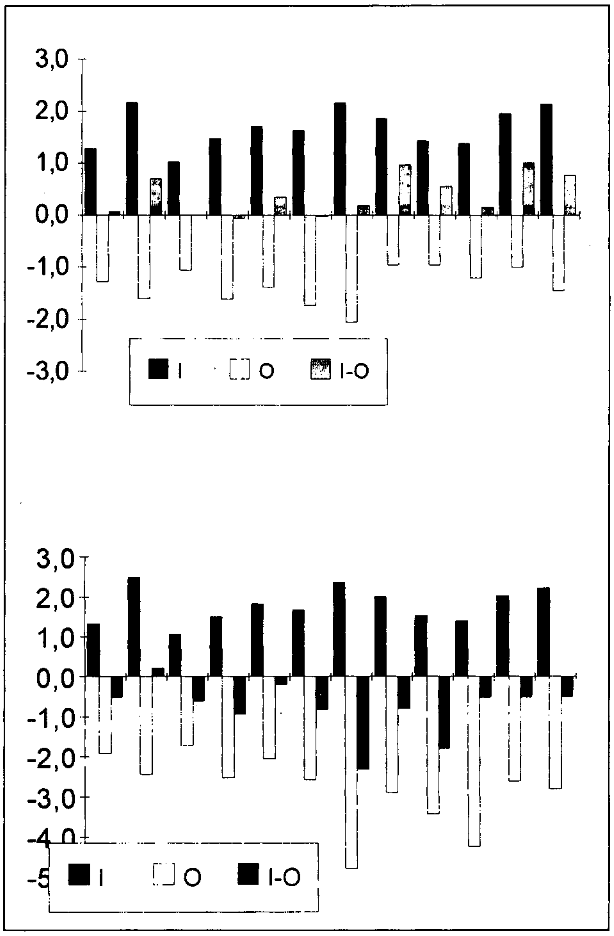

Figure 5. Input-Output budgets (in keq.ha^{-1}.yr^{-1}) for cations in the beech catchment (left), and in the spruce catchment (right) during the 1981-1993 period. (I=Input, O=Output.

The evolution of NO_3 seemed controlled by Ca (figure 4) and Mg leaching , as this can be deduced from their significative relation (NO_3 = 0,42 Ca + 61,86 , R2= 0,90 n=267, and NO_3 = 0,25 Mg+39,16, R2 = 0,85).

The maxima were observed at the same time (the VWC in 1990 was 11,9 higher than in 1987) and the concentrations returned to lower values at the end of 1993.

3.2.2.2. Input-Output budgets.

The figure 5 presents the cation (sum of Ca, Mg, K, Na, NH_4) budgets in the 2 catchments. As the clearfelling started in summer, ions budgets have been calculated from july to june.

In the beech catchment , the cationic denudation rate, expressed as the difference between the Input-Output (Hornung et al, 1990), was always positive (storage), (Dupraz,1984, Durand et al., 1991). As the cation budget is dominated by Ca, this is mainly due to the annual storage of Ca, the other elements being either quickly turned over (K) or stored in the biomass, either slightly leached (Na and Mg). The most efficient storage appeared in 1988-89 and 1990-91,due to the strong input of saharian dusts (Hanchi, 1994) .

In the spruce catchment, cations were always leached from the catchment (Dupraz, 1984, Durand et al.,1991) and Input-Output budgets were moderate (0,55 keq.ha-1.yr-1) during the period 1981-1987. With the clearfelling, the leaching of cations increased, and the most important losses occurred in the first year of treatment (Didon-Lescot et al, 1992). The important runoff (Q=1985,5mm), partly due to a large amount of precipitations, explains the level of the leaching . The magnitude of the cations losses was generally more moderate in the following years, depending of either low runoff in coincidence with relatively high input (drought period), or decreasing concentrations with much runoff corresponding to the wet years of the end of the experiment. Finally during the 4 periods of the spruce stand, the average denudation cation rates -expressed in keq.ha-1.y-1 - were as follow: -0,41 during the first period, -0,65 at the beginning and -1,60 at the end of the clearfelling, and -0,82 in summer 1993, 6 years after.

4- CONCLUSION

The Mont-Lozère clearfelling experiment corroborates the results of many other studies in North America (Bormann and Likens,1979, Feller and Kimmins,1984, Hornbeck et al., 1986, 1991) or in western Europe (Hultberg,1985, Ahtiainen,1988, Adamson and Hornung 1990).
The mecanisms involved in the different studies to produce NO_3 are the mineralisation of organic nitrogen from the ancient forest layer, the decomposition of fine roots and the drop of biological uptake. In the experiments ,the major cation losses are mainly due to the contribution of subsurface waters evacuating cations (Ca, Mg, K) that cannot be retained in ecosystems in reason of the decreasing cation uptake of the plant community (Hornung et al, 1990).

In these case studies, general attention has been allowed to the magnitude and duration of NO_3 losses that seem to be strongly exported during two or three years after the beginning of the disturbance, and return faster to previous values.

In the Mont-Lozère experiment, the period of leaching was important and lasted more than 6 years after the start of the clearfelling. This could be explained either by the effect of water stress that occured from 1989 to 1991 or by a progressive release of N and base cations associated to a compartment of organic matter which was probably sligtly mineralized.

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Spatial distribution and trend change of nitrate in Slovakia during 1968-1993

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ABSTRACT

The nitrate concentrations trends were evaluated on the basis of trend analysis in five sub-basins of the Ondava river basin for 25 - years time series 1968/69 - 1992/93. A rapid decrease of nitrate concentrations in surface waters was observed in four of them after 1989. The decrease resulted mainly from the lower intensity of agricultural production and fertilization in Slovakia due to economic changes. The application of nitrogen fertilizers decreased from 91 kg per ha of agricultural land (mean for Slovakia) in 1989 to 62 kg per ha in 1991. Therefore, the decrease of nitrate concentrations observed in research basins is expected to occur in other agricultural basins as well. Due to continuing changes in agriculture it is difficult to estimate the trends for next few years but probably the decreasing trend will turn again.

In the second part of the paper statistical analysis of measured daily nitrate concentrations was done. The coefficients of theoretical log-normal curves of non-exceedance of the nitrate concentrations were derived in the studied sub-catchments. Both mean annual and characteristic values of nitrate concentrations derived from daily samples were compared to those estimated from regular monthly samplings of hydrometeorological network. Important differences in the estimates of the characteristic values were found between both approaches.

INTRODUCTION

The deterioration of surface water quality in the last decades is typical for most of the industrial regions. Also the high intensity of agricultural production results in lower water quality. The tendency for self-suffiencyc Kendall test for the statistical analysis of water pollution trends from 403 stations and 27 parameters for 1978-87 in the Continental United States. They found, that the common ions and nutrients have mostly upward trends. The transport processes in streams related to land use were studied by Bowden et al. (1991), Ferguson (1986), and Hakamata et al. (1991). Vrba and Skořepa (1986) estimated the nitrate contents in groundwater since 1981, and Pelikán studied the nitrate concentrations in Moravian rivers in 1963-88. Nesměrák (1986a, 1986b) modelled nitrate washout as related to the runoff depths. He also elaborated the state standard ČSN 75 7221 "Classification of surface water quality" which is valid also for Slovakia since 1990.

In Slovakia the unfavourable influence of agricultural production on water quality was studied by Lichvár (1986). Hanzlíková and Büchlerová (1981) estimated the development of water quality in Velká Domaša reservoir basin and they processed the mean water quality parameters for last 30 years. This area is also corresponding to our investigations.

In our institute Babiaková and Palkovič (1992) were dealing with the accummulation of sulfates and nitrates in snow cover and Rončák and Koníček (1992) studied the nitrates in selected research basins of central Slovakia. Mendel and Pekárová (1993), Mendel and Halmová (1993), Pekárová and Pekár (1993), and Pekárová and Halmová (1994) were dealing with the possibilities of nitrate concentration modelling related to discharges

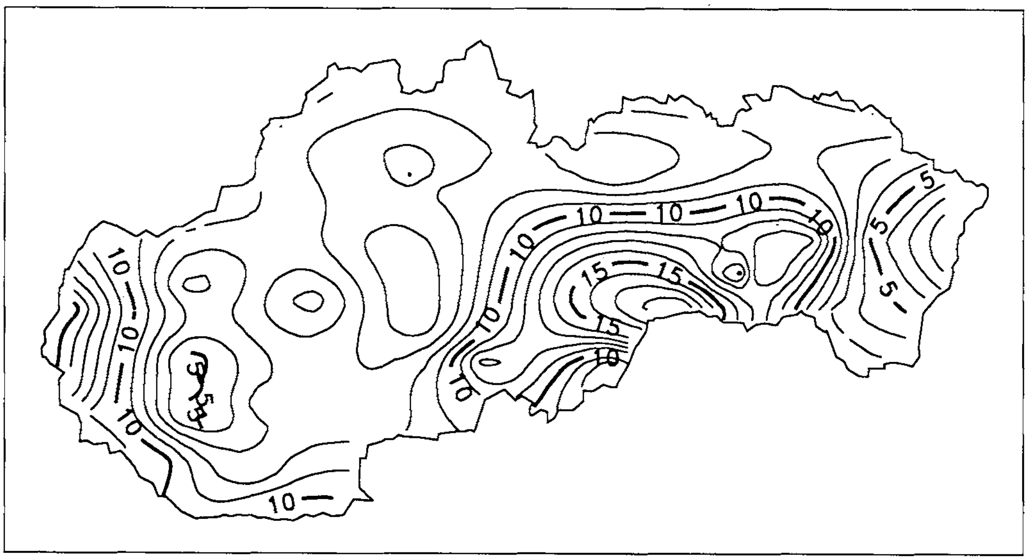


Figure 1. Spatial distribution of nitrates (mg l^{-1}) in surface water in Slovakia in 1989

and the forecasting of the concentrations.

The monitoring of the precipitation, surface and ground water quality is organized by the Slovak Hydrometeorological Institute. According to the data from selected 51 profiles in Slovakia the mean annual nitrate concentration in stream water was 9.7 mgl-1 in 1989 (SHMI, 1989). The maximum value occurred was 19.0 mgl-1 and the minimum one was 4.3 mgl-1.

For the analysis of long-term trends of water pollution the longer time series are needed. Moreover, to study the dynamics of water pollution changes in dependency on other hydrometeorological elements (discharge, temperature, precipitation) we need to know these changes even in time intervals less than 1 day. To study the spatial changes it is also necessary to sample with higher spatial density.

SAMPLING AREA

In order to study the changes of nitrate concentrations in surface flows the regular daily water quality sampling was organized in 5 sub-catchments of the Ondava river basin. The experiment lasted since 1986/87 to 1992/93. Also the measured monthly nitrate concentration series were available in this basin since 1968.

The Ondava river basin is situated in Eastern Slovakia (49o N lat., 22o E long.) in the Ondava hills. The area of the basin in Stropkov profile is 576.5 km2, and the stream length is 55.4 km. The mean elevation of the basin is 450 m a.s.l. The basin is situated in the flysh zone of the Carpathians and the permeability of the geological substrate is characterised as very low. The brown forest soil prevails with beech - oak cover. The climate of the area is characterised as continental with hot summer and cold winter. The mean annual air temperature is 7.5oC,the mean annual precipitation is 752 mm and the runoff coefficient is 0.45. Concerning the surface water pollution by nitrate the basin belongs to relatively clean regions of Slovakia (Fig. 1).

There are no important sources of industrial water pollution in the basin. The only point sources of pollution are the municipal sewerages of Svidník and Stropkov (both less than 15.000 inhabitants). One half of the area is forested, the other one is agricultural area (17% is arable land). On Ondava river 9 km below Stropkov the multi - purpose water reservoir Velká Domaša is in operation since 1970 (total capacity 187.5 mil.m^3, area 14.9 km^2). The possibility of its use for drinking water supply was considered by water authorities. Due to low soil fertility the mean annual doses of NO_3- -N fertilizer were up to 140 kg N ha-1a-1 in 1989.

For the study of spatial and temporal changes of the nitrate concentrations within the Ondava basin 5 sub-catchments were chosen. Their basic characteristics are as follows (Table 1). The land use of these catchments is different. The agricultural land use (pasture, arable land) can be expressed as a complement to the forest cover-to-drainage area which is 95%, 60%, 50%, 40% and 9%, respectively.

METHODS

The measured annual series of daily nitrate concentrations were put together in the form of frequency tables. Before selecting the most suitable type of probability distribution curve of nitrate concentrations, several types of probability distribution were tested, mostly used in hydrology: Pearson's of the third type, normal, log-normal, Gumbel's, and Weibull's. The choice of the "best one" for a given series belongs to the classical problems of hydrology and it is not our aim to solve it. For using some of the distribution types it was necessary to take into account a condition that the lower distribution boundary might be a non-negative.
As the studied series have high coefficients of asymmetry, and hold the inequality (1):

$$C_s \geq 3\, C_v$$

where: C_s - coefficient of asymmetry
C_v - coefficient of variation,

the most successful was the use of the three-parametric logarithmic-normal distribution in the form:

$$f(\log(x\text{-}a)) = \frac{1}{\sigma\sqrt{2\pi}} \cdot \frac{1}{x-a} \cdot e^{-\frac{(\log(x\text{-}a)-\mu)^2}{2\sigma^2}}$$

where: x - nitrate concentrations
μ - mean
σ - standard deviation

RESULTS

1. Spatial changes

In order to ascertain the spatial variability of nitrate concentrations and to identify the areas with higher pollution the water quality samples were taken in 72 checking profiles on flows in Ondava basin during 1990-92. The sampling was done in different seasons and different hydrological situations. The checking profiles were selected on both main stream and the tributaries above the Velká Domaša reservoir. The measured data were used for drawing nitrate concentration maps of Ondava basin in different seasons. Using these data also the areas with higher pollution of surface water by nitrates were identified within the basin (Fig. 2).

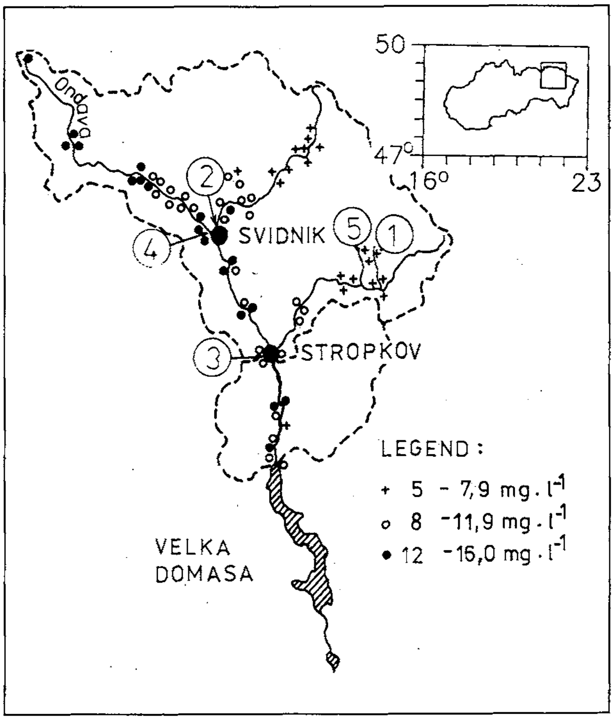


Figure 2. Mean nitrate concentrations in surface water in Ondava basin in 1990-1992. Network stations: 1- Manelo, 2- Ladomírka - svidník, 3- Ondava - Stropkov, 4- Ondava - Svidník, 5- Babie

2. Temporal changes

The water quality samples were taken daily at 6 a.m. during the hydrological years 1986/87 - 1992/93. The mean annual nitrate concentrations in the subcatchments are in Table 2 and the daily nitrate concentrations in cross section 3 are drawn on Fig 3 for seven hydrological years 1986/87 - 1992/93.

3. Trend analysis - smoothing of the monthly nitrate concentration time series

The 25 - years time series of monthly nitrate concentrations in profiles 2, 3 and 4 (see Fig. 2) were processed using trend analysis. On Fig. 4 there we can follow the increase of mean monthly nitrate concentrations in profile 3 Stropkov - Ondava in 1968/69 - 1989/90. The mean annual concentrations increased during 20 years from 2.9 mgl-1 in 1968/69 to 12.8 mgl-1 in 1988/89. After 1988/89 the mean annual nitrate concentration decreased to 5.3 mgl-1 in hydrological year 1992/93.
The simple moving averages method was used for smoothing the monthly nitrates concentration time series. The length of the moving averages was chosen as m = 24. On Fig. 5 there are both original and smoothed data

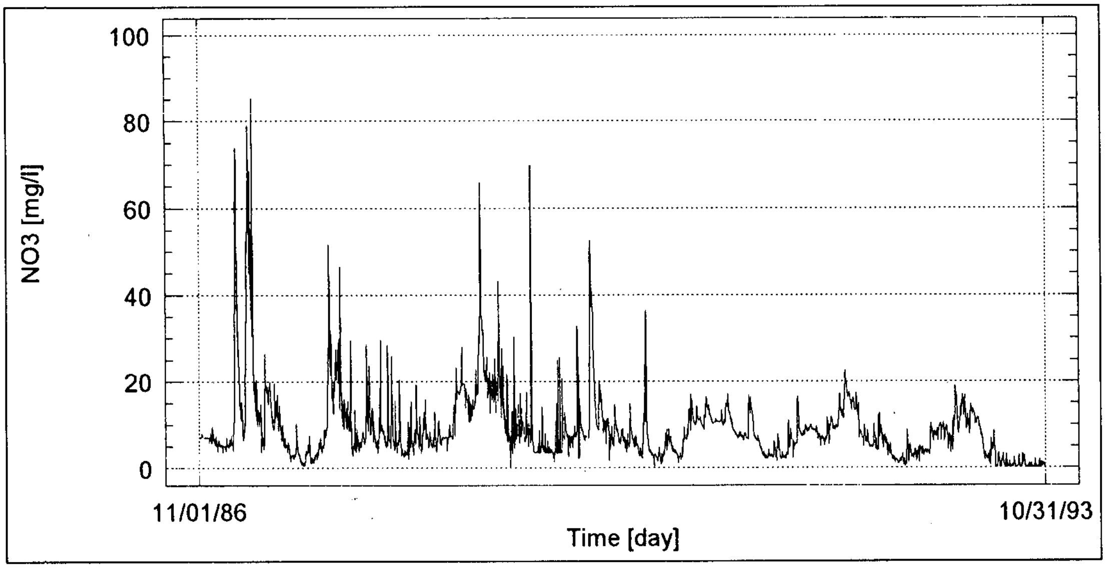


Figure 3. Observed daily nitrate concentrations at Stropkov during seven hydrological years 1986/87 - 1992/93

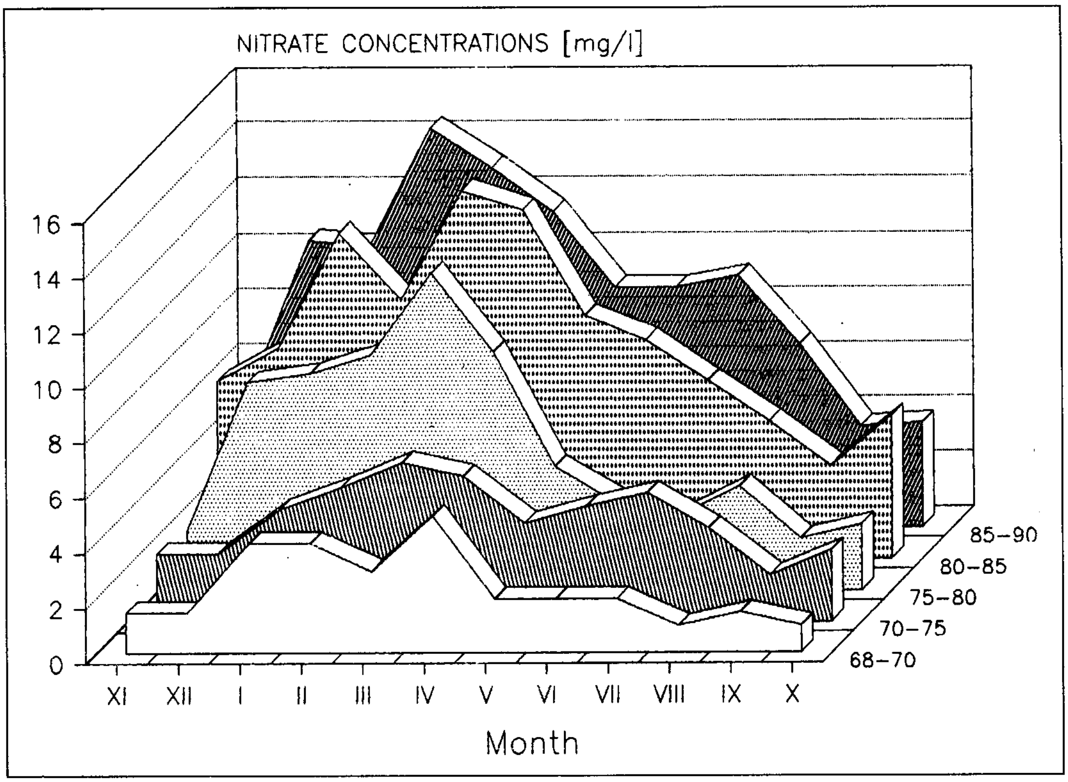


Figure 4. The increase of mean monthly nitrate concentrations at Stropkov in 1968/69 - 1989/90

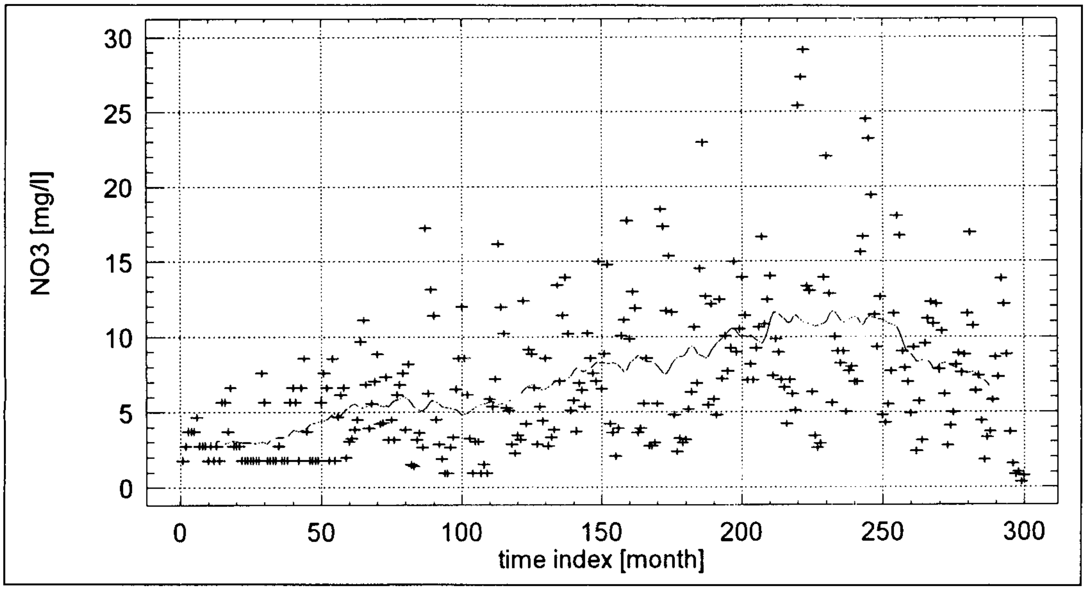


Figure 5. Trend of monthly nitrate concentrations. Measured and smoothed dara from profile 3 Ondava - Stropkov since November 1968 - October 1993

from profile 3. The development of nitrate concentrations in the studied area is also evident from Fig. 5. The concentrations were increasing during 1968/69 - 1986/87, the next four years were stabilized on the highest level and from 1989 onwards concentrations decreased.

4.Testing of daily nitrate concentrations time series - the choice of theoretical curves of the non - exceedance probability of the nitrate concentrations in the streams

In Table 3 there are the basic statistical characteristics of the measured series of daily nitrate concentrations for the six years.

In Figs. 6a - 6f can be visually evaluated an agreement of empirical values and fitted logarithmic-normal curve of distribution for the profile 3 Stropkov - Ondava. For mathematical evaluation of agreement between empirical and theoretical curve of distribution there were used: (2-test, Kolmogorov-Smirnov test and the Romanovsky criterion.

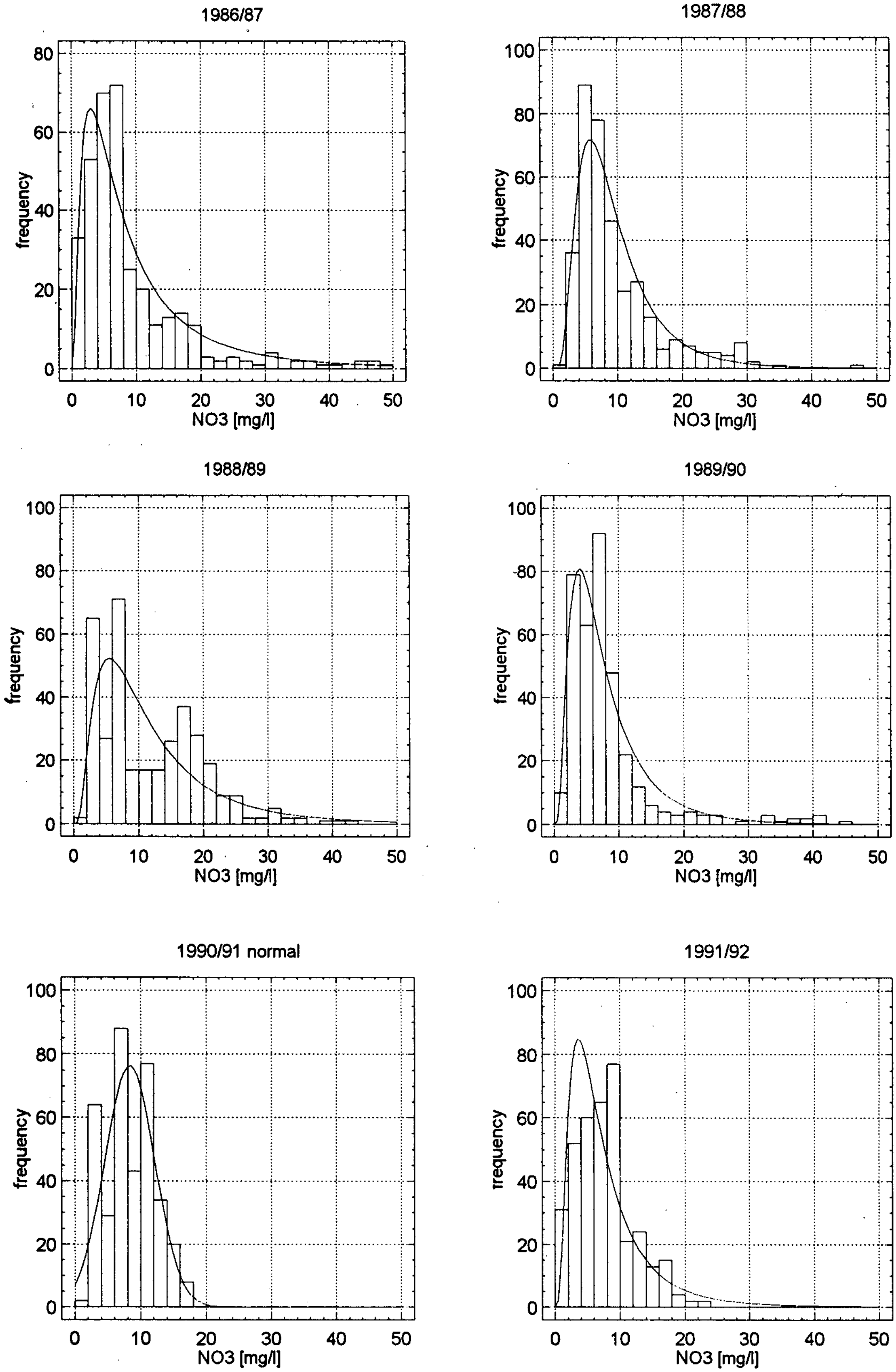
On Fig. 7 there are the empirical curves of non-exceedance of the daily nitrate concentrations in three basins: 1. - forested basin, 3. - mixed land use, 5.- agricultural basin.

The figure shows that most frequently are the higher values of nitrate concentrations exceeded in subbasin 5 Babie - Olšavka. This result is influenced by the fact that the forest-cover-to drainage is the lowest in this basin - 9%, i.e. 91% of the soil is agriculturally exploited.

On the basis of fitted curves of nitrate concentrations exceeedance Table 4 was produced, giving the critical values of nitrate concentrations. There are also the critical values derived from the observed values. The table was elaborated for all five profiles for six years. From Table 4 follows that the 6 - years fitted curves are not suitable for determination of the characteristic values (concentration with probability of non-exceedance 90%) because of the big differencies between corresponding values. The theoretical curves should be estimated for shorter periods which are not influenced by the trend.

By means of non-exceedance curves it is possible to evaluate the water quality in the Ondava river, with regard to nitrates, according to state standard ČSN 75 7221, valid for classification of the quality of surface waters. As a characteristic value is considered the value of concentration with probability of non-exceedance 90% which is calculated from at least 24 values. In the streams intended for drinking purposes (both for inhabitants and animal production), for breeding of salmon-like fish and at special protection of the basin, the 95% of probability non-exceedance is used.

Further, the state standard ČSN 75 7221 establishes the permissible nitrate concentration of up to 15 mg.l-1 for the II. class (pure water) and 50 mg.l-1 for the IV. class.



Figures 6a. 6f. Histograms and theoretical logarithmic-normal (normal) distribution curves for the profile 3 stropkov - Ondava in different hydrological years

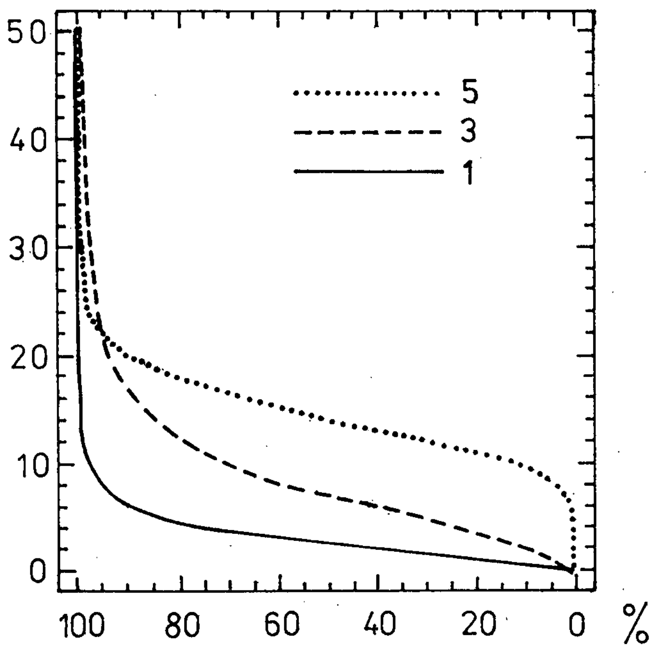


Figure 7. The empirycal curves of non exceedance of the daily nitrate concentrations in different basins (1.- forested, 3.- mixed land use, 5.- Agricultural)

The value of 15 mg.l-1 was, with 95% of probability, not exceeded only in the year 1987/88 in the profile 4 Svidnìk-Ladomìrka. In all other cases it was exceeded. In hydrological year 1988/89 in the profile 4 Svidnìk-Ondava there was exceeded the value of 50 mg.l-1, too.

CONCLUSIONS

The mean annual and characteristic values (values with the probability of non - exceedance 90 % according to the state standard ČSN 75 7221) derived from daily sampling in cross - sections 2, 3 and 4 in 1989 were compared to those derived by Slovak Hydrometeorological Institute in the same cross - sections from monthly sampling (Table 5)

It is obvious that the variability of the daily samples is higher than that of monthly samples. The higher frequency of sampling can discover the extreme values more probably. This assumption is confirmed in Table 5. According to monthly sampling is the upper Ondava river classified into the class II, while according to our data it should be classified as class III..

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Table 1. General characteristics of the five sub-basins

Flow	Profile	Area km²	Forest %	Elevation m a.s.l min	max	mean	Prec mm	Runoff mm	coeff.
1. Manelo	Gribov	0.195	95	305	485	383	674	297	0.44
2. Ladomírka	Svidník	185.8	60	225	–	-752	789	404	0.52
3. Ondava	Stropkov	574.0	50	183	543	752	765	345	0.45
4. Ondava	Svidník	167.5	40	225	543	752	747	347	0.46
5. Babie	Olšavka	0.345	9	292	520	406	767	408	0.53

Table 2. Arithmetic mean nitrate concentrations in 5 Ondava sub-basins estimated from measured daily concentrations in mgl-1

Flow	Profile	86/87	87/88	88/89	89/90	90/91	91/92	92/93
Manelo	Gribov	-	3.8	3.9	3.6	3.1	2.6	1.8
Ladomírka	Svidník	14.5	5.6	12.6	9.6	8.1	8.4	6.3
Ondava	Stropkov	11.7	9.9	12.8	8.5	8.3	7.6	5.3
Ondava	Svidník	15.4	7.2	15.3	11.9	9.9	8.6	7.5
Babie	Olšavka	-	16.3	13.7	13.6	14.3	16.4	13.3

Table 3. The basic statistical characteristics of the measured daily nitrate concentrations in 5 sub-catchments for the six hydrological years 1987/88 - 1992/93

sub-catchment number	1.	2.	3.	4.	5.
sample size	2192	2192	2192	2192	2192
mean concentration	3.11	8.44	8.76	10.60	14.60
maximum	24.00	87.40	69.90	135.2	88.00
median	2.40	7.30	7.28	8.72	13.88
modus	0.90	2.00	6.50	1.50	19.00
variation	6.66	46.00	50.72	88.34	28.86
standard deviation	2.58	6.78	7.12	9.40	5.37
standard error	0.06	0.14	0.15	0.22	0.11
coefficient of asymmetry	2.72	3.68	2.68	3.31	4.38
coefficient of variation	0.83	0.83	0.81	0.88	0.36

Table 4. Critical values of nitrate concentrations (mgl-1)

				From fitted curves				From measured values			
Nº	catchment	m	s	95%	90%	50%	10%	95%	90%	50%	10%
1	Gribov	3.14	2.75	8.1	6.2	2.3	0.9	7.9	6.3	2.4	0.9
2	Svidník	9.54	9.53	26.5	19.6	6.7	2.3	21.0	16.0	7.4	2.4
3	Stropkov	10.6	14.4	33.8	23.0	6.2	1.6	22.3	17.2	7.1	2.6
4	Svidník	11.9	14.3	35.9	25.5	7.6	2.2	27.8	22.3	8.8	1.95
5	Babie	14.7	5.4	24.8	21.1	13.8	8.8	21.9	19.7	13.8	9.8

Table 5. Mean annual and characteristic values c90 of nitrate concentrations (mgl-1) in cross-sections 2, 3 and 4 from monthly network samples (subscript n) and from daily experimental samples - measured (subscript m) and fitted curves (subscript t)

Cross-section	C_n	C_m	C_t	C_{90n}	class	C_{90m}	C_{90t}	class
2-Svidník - Ladomírka.	9.19	12.6	12.6	11.3	II	22.0	24.4	III
3-Stropkov - Ondava	11.2	12.7	13.0	13.7	II	22.4	25.8	III
4-Svidník- Ondava	11.6	15.3	16.7	14.0	II	26.3	27.0	III

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Nitrate concentrations in the Gulp catchment: some spatial and temporal considerations

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ABSTRACT

In a chalk catchment in the Belgium-Dutch boundary region the median NO3- concentration was 30 mg/l in 1991. Groundwater in wells, which are mostly located close to the villages, is more polluted than spring water of surface water from the Gulp brook. Median NO3- concentrations were 39, 22 and 17 mg/l, respectively. Since 1980 time-series of NO3- from two gauging stations in the Gulp brook have showed a distinct seasonal pattern. In wet periods (winter), when the discharge is higher, the nitrate concentration also is higher (30-35 mg NO3-/l), whereas in dry periods the opposite occurs; i.e. lower discharge and nitrate concentrations (10-15 mg NO3-/l). The positive correlation between the discharge and the NO3- concentrations cannot be explained by the contribution of overland flow and interflow (quick flow components) probably having higher NO3- contents. Saturated groundwater modelling shows that water following different flow paths might be an explanation. In periods with high groundwater recharge, young groundwater in the upper chalk layer can bypass medium-aged and old groundwater in the lower chalk and greensands. Young groundwater is likely to have higher NO3- concentrations than old groundwater. So, in periods with high groundwater recharge the stream flow of the Gulp brook might consist of more young groundwater with higher NO_3- concentrations, whereas in periods with low recharge mainly old groundwater with lower NO_3- concentrations feeds the Gulp.

INTRODUCTION

In the southeast of the Netherlands the groundwater quality has been deteriorated by an increase of nitrate. In this region 20.106 m^3 groundwater per annum (about 20% of the groundwater supplied by the Drinking Water Company of the Province of Limburg) is extracted from unconfined chalk aquifers. In one well field, i.e. IJzeren Kuilen, the NO_3- concentration has increased from 15 mg/l in 1955 to about 40 mg/l in the late 1980's. The NO3- concentration is expected to exceed the maximum allowable concentration of 50 mg/l in the early 2000's in spite of possible nitrogen reduction measures (Juhász-Holterman et al., 1989). The NO_3- concentration of extracted groundwater does not continuously increase, but is related to groundwater recharge. The NO_3- concentration significantly increases when wet years with a high groundwater recharge occur. In dry years the concentration drops again. Leaching of NO_3- from the thick unsaturated zone, inclusion of a nitrate-rich unsaturated zone in the saturated zone due to a rise of the water table, and preferential saturated flow in the upper part of the chalk causes the positive correlation between NO_3- concentrations of extracted groundwater and groundwater recharge in wet years (Juhász-Holterman, 1991).

The increased NO^3- concentrations in the groundwater system also affect the surface water quality. In some chalk streams, e.g. the Mechelderbeek, the NO_3- concentration has doubled (rise from 16 to 33 mg/l) in the period 1973-1984 (Schouten et al., 1986). Nota et al. (1988) show that a typical chalk brook like the Gulp has a distinct seasonal pattern. NO_3- concentrations are positively correlated with discharge. In the winter period both discharge and NO_3- concentrations are higher than during the summer. They explain these seasonal NO_3-

differences by assuming that during the winter hillslope subsurface water flushes NO_3- from the top layers towards the valley. Although they suggest this flow path, no specific investigations are carried out yet to obtain firm evidence. The objective of this paper is: (1) to investigate spatial differences of NO_3- concentration of groundwater and spring water in distinct parts of the Gulp catchment, (2) to investigate if the observed seasonal NO_3- patterns by Nota et al. (1988) still exist in the early 1990's, and (3) to explore possible groundwater flow paths using groundwater modelling, which might explain the seasonal variability of NO_3- concentrations of surface water in the Gulp brook. We assume that groundwater that follows a shallow flow route through the upper part of the chalk has a higher probability on increased NO_3- concentrations than the groundwater following a deeper route. Therefore the proportional distribution of groundwater following these different flow paths for various groundwater recharge regimes was investigated.

METHODS AND MATERIALS

Gulp catchment

The Gulp brook is a stream that drains the dissected chalk plateau in the southeast of the Netherlands and adjacent Belgium areas (Fig. 1). The total area of the basin is about 4600 ha, the length of the valley is 18 km, while the maximum width is 4 km. The Gulp brook rises at 285 m a.m.s.l. in Henri-Chapelle (Belgium), and joins the Geul, which is a tributary of the Meuse, in the village of Gulpen at 88 m a.m.s.l.. Near the village of Slenaken the Gulp brook enters Dutch territory. The cross profile of the Gulp catchment is asymmetrical (Fig. 1); the steepness of the eastern slope amounts to 17%, whereas the western slope averages 5%. Permanent grassland covers more than 90% of the catchment. Forest predominantly occurs on the steep, eastern slopes. The annual precipitation averages some 800 mm in Hombourg, and the average stream flow at the outlet is about 500 l/sec.

The geology of the Gulp catchment and some hydrogeological features are given in Fig. 1 and Table 1. Tilting and faulting because of the uneven Cenozoic uplifting has resulted in a fault zone in the western part of the basin.

The Gulp brook is deeply incised into the chalk plateau, which implies that surface water levels are far below the plateau surface. Outside the valleys the unsaturated zone is thick (up to 60 m). The earlier-mentioned Upper-Cretaceous sediments form a multiple-aquifer system. The chalk and the sands of the Aken Formation are aquifers, which permit horizontal groundwater flow. The permeability of the chalk varies with depth; the upper layer has a significantly higher permeability than the lower layers because of weathering processes (Juhász-Holterman et al., 1989; Rooijen & Amkreutz, 1982). From the lower chalk layers the bottom layer (bottom chalk layer) has a permeability of about 1 m/day, whereas the overlying layer (intermediate chalk layer) has a somewhat higher permeability, e.g. between 3 and 5 m/day. The silty layers of the Vaals Formation do not allow substantial horizontal groundwater flow. The interbedded fractured sandstone layers (thickness 0.1-0.2 m), however, act as horizontal groundwater drains. On the plateaus and the slopes the overburden allows readily infiltration of precipitation into the soil. Experimental research confirmed that surface runoff and interflow hardly occur (e.g. Jansen & Verhagen, 1991). Interflow was defined as temporal saturated flow in the unsaturated zone of the hillslopes. Hydrograph analysis showed that at least 70% of the stream flow consists of groundwater discharge (baseflow). This means that outside the villages (including the roads) and the narrow wet valley, the excess precipitation recharges groundwater. The deep water tables show a delayed and smoothed response on excess precipitation. Water reaching the unconfined Upper-Cretaceous aquifer flows through the chalk and fractured sandstone layers in the Vaals Formation to springs in the valley. Probably minor leakage prevails through the clayey silts of the Vaals Formation towards the deep aquifer in the Aken Formation.

Groundwater and surface water monitoring

In the period 1975-1985 groundwater and surface water was intensively monitored in the Gulp catchment. Groundwater heads were observed in several dug wells and observation wells. On some of these locations every second month groundwater was sampled for the analysis of the chemical composition. Discharge of some selected springs and tributaries of the Gulp brook was measured and the chemical composition was analyzed. Stream flow of the Gulp brook was continuously measured at five locations along the longitudinal profile of the Gulp brook, e.g. F6 and F9 (Fig. 1). Once a month surface water was sampled for chemical analysis. Nota & Van de Weerd (1978; 1980), Nota & Bakker (1983),

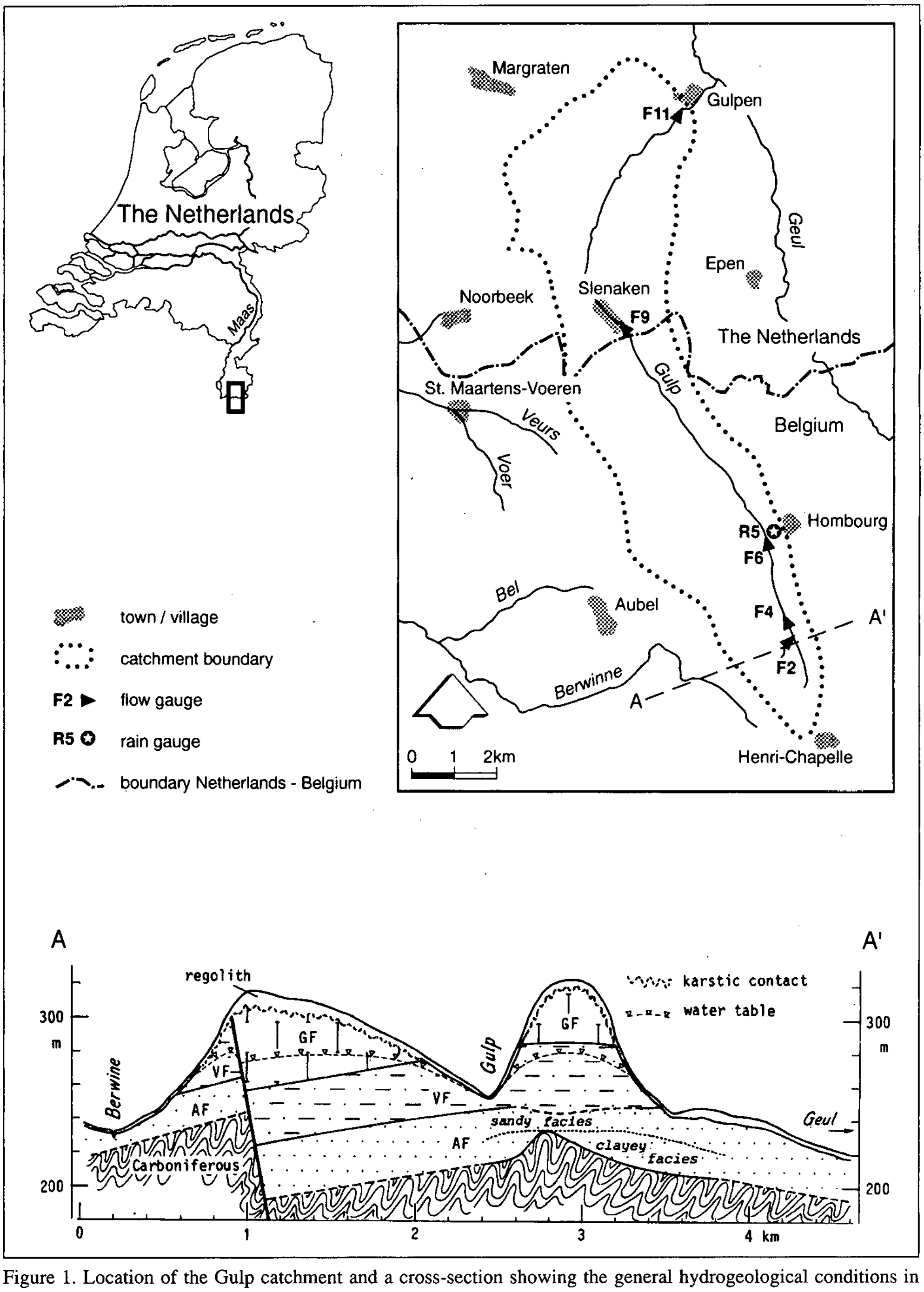

Figure 1. Location of the Gulp catchment and a cross-section showing the general hydrogeological conditions in the south of the catchment (derived from Nota et al., 1988).

and Nota et al. (1988) provide all relevant information. Since 1991 monitoring has started again on some strategic locations to investigate possible trends. In this recent period groundwater heads and surface water discharge including the chemical composition were occasionally measured for specific purposes on nearly all locations of the 1975-1985' network (e.g. Van Duinen, 1992).

The NO_3- concentration of groundwater and surface water was measured in the period 1980-1984 and since 1991 again. In the laboratory ion chromatography was used as an analytical procedure to determine NO_3- concentration of the samples. For the incidental measurements NO3- was analyzed in a field laboratory using an ion-selective electrode. Comparison of both methods showed a good agreement (Van Duinen, 1992).

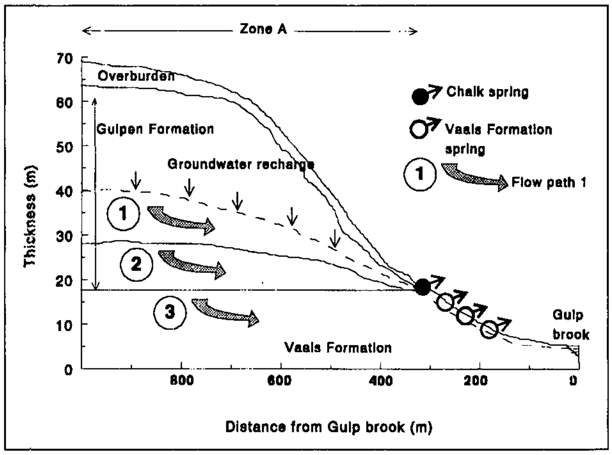

Figure 2. Schematic flow paths as modelled with MODFLOW for periods with a high groundwater recharge

Groundwater flow simulation

In the major part of the Gulp catchment excess NO3- can only reach the Gulp brook through various flow paths in the saturated groundwater system. Except from the villages and roads, no overland flow occurs. Groundwater flow modelling was applied to explore that flow route groundwater follows in the multi-layered aquifer system of the Gulp catchment. The amount of groundwater (specific discharge) following each of these flow paths was simulated. Although NO_3- is a reactive ion no transformations, e.g. denitrification, were simulated in this phase of the study; NO_3- was supposed to behave like a conservative ion. Saturated groundwater flow was simulated with a finite-difference groundwater flow model using the computer code MODFLOW (McDonald & Harbaugh, 1988). MODFLOW enables simulation of the groundwater head distribution and the water balance (e.g. groundwater flow towards a stream). In this reconnaissance stage modelling was restricted to the simulation of steady-state water flow in a vertical cross-section of the valley perpendicular to the Gulp brook. Such cross-sectional or profile models have been used frequently in an interpretive sense to study patterns in regional flow systems before designing a full three-dimensional model (e.g. Anderson & Woessner, 1992). The cross-section was subdivided in cells. In the horizontal direction from the watershed to the Gulp brook 100 cells were distinguished with a length of 10 m. In the vertical direction three cells, i.e. representing different hydro-stratigraphic layers, were selected. In the area where chalk is overlying the Vaals Formation (Fig.1) the upper layer represents the permeable intermediate chalk layer; the less-permeable bottom chalk layer and the Vaals Formation are represented by layer 2 and 3. In the area next to the Gulp brook, where the chalk is eroded, the three vertical cells represent different layers of the Vaals Formation. No leakage to, or seepage from the underlying Aken Formation was assumed to occur. Flow in the thick unsaturated zone was not considered in the model. Groundwater recharge can follow three different flow routes in the model (Fig. 2): (1) horizontal flow through the permeable intermediate chalk layer towards a chalk spring at the margin of the chalk area (represented as a drain with MODFLOW), (2) vertical flow through the permeable intermediate chalk layer to the less-permeable bottom chalk layer and from there in a horizontal direction towards the earlier-mentioned chalk spring (see 1), and (3) vertical flow through both chalk layers to the Vaals Formation and from there in a horizontal direction towards springs or seepage areas at locations where the Vaals Formation outcrops (represented as drains at different elevations with MODFLOW). Part of the groundwater in the Vaals Formation seeps directly away into the Gulp brook (diffuse groundwater drainage). In reality the chalk springs are not found exactly at the permeability break between the chalk and the Vaals Formation, but at slightly lower elevations.

The results of MODFLOW permit computation of the water balance of different parts of the defined cross-section. For the area where the chalk is found (western part of the catchment, zone A in Fig. 2) we calculated the specific discharge of groundwater following the earlier-defined flow routes. This implies that the proportional distribution of the groundwater recharge over the three distinguished flow routes was computed.

The simulations were carried out for four different groundwater recharge regimes ranging from dry to wet under Dutch conditions, i.e. 0.0005, 0.0007, 0.001 and 0.002 m/day (Van Lanen, 1979). In the dry situation part or the entire permeable chalk layer (intermediate layer) might become unsaturated. Under these conditions no groundwater follows flow path 1 (Fig. 2). The schematic conditions in Fig. 2 apply to wet conditions, where part of the groundwater flows through the permeable intermediate layer (flow path 1). MODFLOW can handle a variable saturated thickness of the upper layer and allocates groundwater recharge to the first saturated layer. Heijnen & De Jong (1994) provide more details about the modelling in the Gulp catchment.

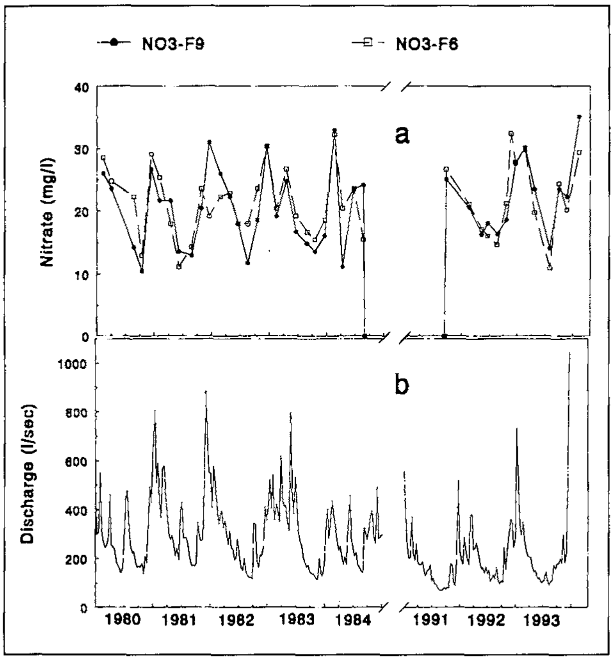

Figure 3. NO_3- and stream flow rate of the Gulp brook: (a): variation of NO_3- of surface water at two gauging stations along the Gulp brook (F6 and F9), and (b): 10-day' average discharge at Slenaken (F9).

RESULTS

Spatial distribution of nitrate observed in the Gulp catchment

The water analyses from the period 1981-1984 showed that NO_3- pollution of groundwater was concentrated around the hamlets, where most of the groundwater observation wells are found. In some areas the NO_3- concentration exceeded the 100 mg/l. Surface waters had significantly lower NO_3- contents (Nota et al., 1988).

In June 1991 the NO_3- concentration was determined of 53 groundwater wells and 50 springs. Surface water of the Gulp brook also was sampled on a few locations (Van Duinen, 1992). The results are summarized in Table 2.

Groundwater sampled from the wells contains higher NO_3- concentrations than groundwater from the springs (median: 39 and 22 mg/l, respectively). This supports the conclusion of Nota et al. (1988) that groundwater taken from the wells reflects more local features (point observations) than spring water or surface water that shows more areal features. On the western slope groundwater taken from half the wells has NO_3- contents exceeding the maximum allowable drinking water limit of 50 mg NO_3- per litre. The occurrence of forests and the lower agricultural nitrate load on the steeper eastern slopes is clearly reflected in the lower NO_3- concentrations of the eastern wells and springs. The groundwater discharge from the eastern part of the basin is substantially lower than from the western part. Stream flow of the Gulp mainly consists of groundwater from the western part of the catchment. Therefore nitrate concentrations in this part have a higher impact on the nitrate concentrations of the Gulp than the eastern part. Groundwater is more polluted than the surface water of the Gulp brook; the median NO_3- concentration is about twice as high (30 and 17 mg/l). This implies that the Gulp brook is also fed by other nitrate-poor groundwater than the groundwater taken from the selected wells and springs. The sampled wells are not fully representative for all groundwater. No clear zonation in NO_3- concentrations of the groundwater from the wells and springs could be observed in a cross-section from the plateau to the valley, which would identify discharge areas of groundwater following different flow paths (Fig. 2).

Temporal distribution of nitrate observed in the Gulp brook

In the years 1980-1984 the monthly determined NO_3- concentrations of the Gulp brook had a positive correlation with the 10-day' average stream flow (Nota et al., 1988). This positive correlation also applies to the period 1992-1993 (Fig. 3). More frequent investigations showed that during short periods, i.e. 22 May - 4 July 1991, when the stream flow was somewhat constant, NO_3- concentrations did not show much variation (Van Duinen, 1992).

The NO_3- time-series of the surface water sampled at the two gauging stations (for locations see Fig. 1) are similar. The observation period is rather short to investigate a possible trend in NO_3- concentrations-. The NO_3- concentrations of surface water of the Gulp brook, which consists predominantly of groundwater, has a remarkably seasonal variability. The proportional distribution of groundwater discharge following different flow paths in dry and wet periods might be an explanation for this phenomenon as will be shown below.

Simulated proportional distribution of groundwater flow following different flow paths

The proportional distribution of groundwater recharge in the chalk area over the three flow routes is calculated

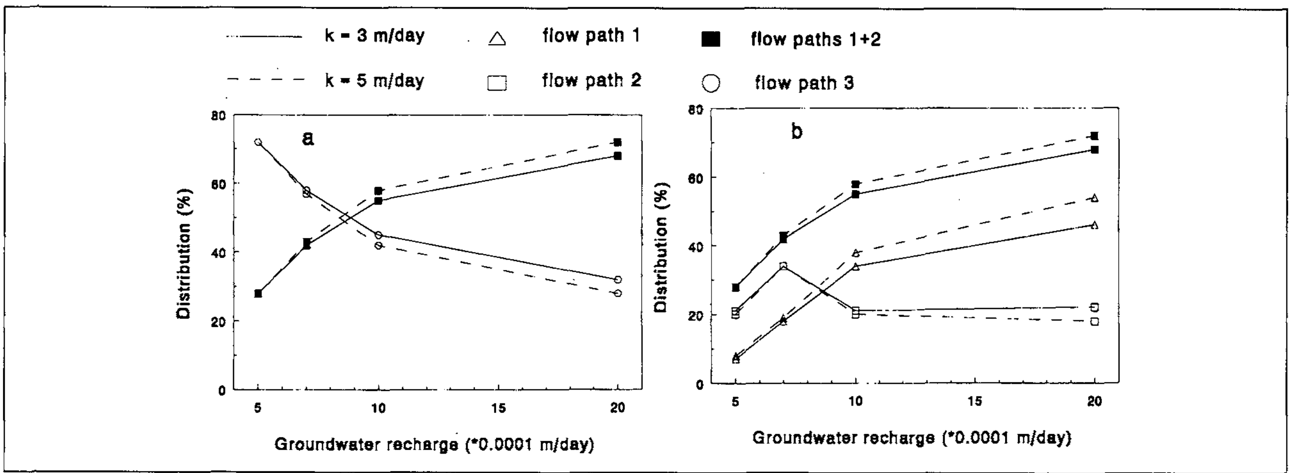

Figure 4. Simulated proportional distribution of the groundwater flow following different flow paths for various groundwater recharge regimes and two permeabilities of the intermediate chalk layer, (a) distribution of groundwater flowing through the chalk (flow paths 1+2) and groundwater flowing through the Vaals Formation (flow path 3), and (b) distribution of groundwater flowing through the permeable intermediate chalk layer (flow paths 1) and groundwater flowing through the less-permeable bottom chalk layer (flow path 2).

using MODFLOW for four different groundwater recharge regimes, i.e. varying from 0.0005 to 0.002 m/day. The simulations were carried out for two different permeabilities of the permeable intermediate chalk layer (layer 1), namely 3 and 5 m/day. The distribution of the recharge over the groundwater flow to the chalk spring (flow paths 1 and 2, see Fig. 2) and groundwater flow to the Formation of Vaals (flow path 3) is given in Fig. 4a.

In dry years (recharge 0.0005 m/day) about 30% of the groundwater recharge flows towards the chalk springs and the remaining part flows towards and through the Vaals Formation. In extremely wet years (groundwater recharge 0.002 m/day), when the intermediate chalk layer is saturated, the opposite occurs, about 70% of the recharge flows towards the chalk springs and only 30% through the Vaals Formation. Under more average conditions (recharge 0.0007-0.001 m/day) the distribution is close to 50%.

The distribution of the groundwater recharge over the two distinguished flow paths in the chalk, i.e. through the permeable intermediate chalk layer (flow path 1) and through the less-permeable bottom chalk layer (flow path 2) is given in Fig. 4b. In dry years the major part of the chalk spring flow (less than 30% of total recharge) comes from the less-permeable bottom chalk layer. This situation changes if recharge is slightly above average (recharge 0.001 m/day). In wet years chalk springs are predominantly fed by groundwater flowing through the more permeable intermediate chalk layer (about two-thirds of spring flow). A nearly doubling of the permeability of the intermediate chalk layer has a minor influence on the proportional distribution (Fig. 4a and 4b).

The simulated residence times of groundwater following flow path 1, 2 or 3 might be in the ratio of 1:10:100, although no firm evidence (e.g. tracer tests) is available. This implies that the surface water of the Gulp brook might comprise young groundwater from the intermediate chalk layer (flow path 1), medium-aged groundwater from the bottom chalk layer (flow path 2) and old groundwater from the Vaals Formation (flow path 3). The distribution of residence times of the water particles in the Gulp brook depends on groundwater recharge. In dry years the Gulp brook consists of more old water that followed flow path 3 through the Vaals Formation, whereas in wet years young groundwater that flowed through the intermediate chalk layer dominates. For example in wet years about half the surface water of the Gulp brook consists of young groundwater. In dry years the percentage young water drops to less than 10%. Under these conditions a substantial part of the groundwater recharge can relatively quickly flow towards the chalk springs. This shallow young groundwater bypasses deep old groundwater.

CONCLUSIONS AND DISCUSSION

Groundwater simulation showed that the proportional distribution of groundwater flow to the Gulp brook following different groundwater flow paths in the Upper-Cretaceous multiple aquifer system is substantially affected by groundwater recharge. In wet years the

Table1. Hydrogeological characteristics of the Gulp catchment

Time period	Formation	Description
Quaternary		unconsolidated regolith or overburden (1-10 m); mixture of clay with flints and loess; unsaturated; permeability up to 5 m/day
Upper-Cretaceous	Gulpen (GF)	slightly consolidated, light coloured, fine-grained chalk with fissures (max. 40 m); upper part unsaturated; permeability: 1-25 m/day
	Vaals (VF)	saturated; glauconite-containing, unconsolidated clayey silts with thin fractured sandstone layers (max. 40 m); permeability of silts and sandstone layers: 0.02-0.2 m/day and 100-500 m/day, respectively
	Aken (AF)	starts with clayey sediments and proceeds with slightly cemented fine-grained sands (max. 35 m); permeability of sands: 6-8 m/day
Upper-Carboniferous		consolidated shales and sandstones impermeable base

permeable intermediate chalk layer is saturated and a significant part of the groundwater recharge (40-50%, Fig. 4) can follow a short flow path through this chalk layer and quickly feed the Gulp brook. Even under average recharge conditions, when part of the intermediate chalk layer is unsaturated, half the groundwater recharge flows relatively fast through the less-permeable bottom chalk layer (Fig. 4), which bypasses groundwater flowing through the Vaals Formation. So, the Gulp brook is fed by groundwater with different ages, which vary dependent on the groundwater recharge. This might be an explanation for the distinct seasonal patterns of NO_3- in the surface water of the Gulp. The short groundwater flow routes (paths 1 and 2) are likely to have higher NO_3- concentrations than the slower one through the Vaals Formation. Groundwater following the short flow paths contains recently leached water from the unsaturated zone that generally is rich in NO_3-. Currently, in South-Limburg NO_3- concentrations of soil water above 100 mg/l are not exceptional under agricultural land (Juhász-Holterman et al., 1989; Bosch & Pijpers, 1991). In the well-aerated soils, which are poor in organic matter, denitrification of NO_3- is low (Schouten et. al., 1986). Deep groundwater that follows the long flow path contains less NO_3- because it might be infiltrated before the increase of nitrate load by agriculture started some decades ago. Moreover in the deeper layers denitrification is likely to occur because the oxygen content decreases and some pyrite occurs in the Vaals Formation (Nota et al., 1988). Denitrification also will occur in the wet valley itself, where organic matter causes transformation of the NO_3-, which could decrease the NO_3- contents of the deep groundwater from the Vaals Formation that feeds the Gulp brook there.

The assumption that shallow young groundwater is nitrate-rich is supported by high NO_3- concentrations of groundwater in most of the observation wells. The wells only penetrate the saturated zone over a few meters, thereby only young groundwater (leakage areas) or a mixture of shallow young grondwater and upcoming deep old groundwater (seepage areas) can be sampled.

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Table 2. Statistics of the NO_3 concentration (mg/l) in the Gulp catchment in 1991 (derived from Van Duinen, 1992).

Type	n	Median	Percentile	
			25th	75th
Groundwater				
All wells and springs	103	30	13	54
All wells	53	39	22	66
• western1) wells	41	50	29	74
• eastern1) wells	12	11	4	46
All springs	50	22	9	34
• western springs	32	30	13	37
• eastern springs	18	16	7	23
Surface water				
Gulp brook	4	17	17[2)]	19[2)]

1) western and eastern: west and east from Gulp brook, respectively
2) minimum and maximum, insufficient data to determine percentiles

on an earlier version improved the paper. The discharge data were kindly provided by the Waterschap Roer en Overmaas.

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A study of soil moisture controls on streamflow behaviour: results for the OCK basin, United Kingdom

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ABSTRACT

In humid temperate areas ground wetness plays a key role in storm runoff generation, but until recently there have been no instruments capable of providing continuous, reliable, records of changing soil moisture conditions in the field. A new instrument, the IH capacitance probe, can provide continuous measurements of soil water contents. Together with rainfall records these data have been used to study the variations in river flow response of a medium sized (234 km2) rural catchment. Daily flows were simulated, firstly using a standard rainfall runoff model (IHACRES) with conventional hydrological and climate data and, secondly, by replacing the net rainfall calculation by a simple functional relationship to the measured soil moisture contents.

It was found that incorporating soil moisture measurements in the runoff model:

a)Reduced the length of record required for model calibration,
b) Improved the simulation of streamflow.

INTRODUCTION

Hydrologists require information about rainfall and soil conditions for flood warnings and for design flood estimation. Whilst the measurement of rainfall has received much attention, especially since the implementation of weather radars, there is generally little information on the changing status of ground wetness. Yet this is a key factor affecting the response of humid temperate zone catchments to rainfall; very different flood responses are observed for similar rainfall inputs onto different initial ground wetness conditions. At present, measurements of soil water content using manually read neutron probes are normally available only on a weekly or monthly basis. For changes over shorter periods an index based on weather data (rainfall and potential evaporation) is commonly used, as for example in the Flood Studies Report (NERC, 1975) which is the standard method for engineering flood design in the UK.

Early research work on flood flows was based on observations of widespread overland flow in the semi arid south western USA (Horton, 1933). However, in well vegetated humid temperate areas, overland flow is rarely seen. Infiltration capacity is generally far greater than most rainfall intensities, so that except on disturbed ground infiltration excess runoff will not normally occur. Some studies show stormflow results mainly from direct channel precipitation and saturation excess overland flow generated by rainfall onto saturated areas close to streams (Cappus, 1960; Dunne and Black, 1970). Other research has indicated the importance of subsurface stormflow from close to stream channels (Hewlett and Hibbert, 1967). Evidence of the importance of subsurface flows also comes from observations of saturated soil layers above an impeding horizon (Weyman, 1970) and from macropores (Beven and Germann, 1982). It is now recognised that even within a single catchment a range of runoff generation processes will be operating. These studies have, however, emphasised the general importance of near surface soil water conditions on stormflow, whether through its control on the extent of saturation excess overland flow, or through its control on subsurface flows. Isotopic evidence also indicates the importance of subsurface flows (Pearce et al, 1986).

Spatial pattern of soil moisture

Once it became recognised that only the saturated parts of a catchment could contribute to quick flow, much work has been conducted to predict the location and extent of these zones. Dunne et al (1975) describe field survey methods for the recognition and mapping of the saturated zones in small catchments. In addition to contiguous channel-side areas, Kirkby & Chorley (1967) identified areas of subsurface flow convergence likely to lead to soil saturation. These comprise: concavities in plan (contour curvature) and in slope profile, and also areas of thin soils. To these situations may also be added soils in which porosity and permeability decrease with depth (especially, but not exclusively in layered soils) resulting in the building up of saturated layers, above any regional groundwater table.

Beven and Kirkby (1979) describe a semi distributed hydrological model, TOPMODEL, which uses a topographic index (Kirkby, 1975) to describe the propensity of any point in a catchment to develop saturated conditions. Calculations of the index for a number of catchments show it to have a skewed frequency distribution; some parts of a basin are subject to saturation more frequently, and for longer periods, than other areas. Some zones may rarely - if ever - become saturated except in the most exceptional conditions.

Problems of predicting storm runoff volumes

Much progress has been made in understanding the complexity of stormflow processes, but practical methods for estimating storm losses and runoff have yet to be developed (Pilgrim and Cordery, 1993). Tests of the US Soil Conservation Service method found large differences between observed and predicted peaks, with poorer results for dense vegetation cover than for bare soil or sparse vegetation; it was concluded that the assumed antecedent moisture condition had a major effect (Pilgrim and Cordery, 1993). The weakness of the estimation of storm runoff coefficients in UK flood design is also recognised (IH, 1987).

Whilst the critical importance of catchment wetness for streamflow generation is now widely appreciated, soil moisture measurements have been limited by the instrumentation available. Until recently soil moisture measurements had to be made by manual methods - gravimetric sampling or by neutron probe - requiring a site visit for each set of readings.

SOIL MOISTURE MEASUREMENTS

Recent instrument developments based on the measurement of bulk soil dielectric constant by Time Domain Reflectometry (Topp et al, 1980) or capacitance probe (Dean et al, 1987) are capable of measuring changes in soil moisture continuously. This paper describes initial results of a project to test the feasibility of such an approach using the Institute of Hydrology (IH) designed Capacitance Probe. Manually operated variants have been used in the field over a number of years, to monitor changing ground wetness (eg Dean et al, 1987; Robinson and Dean, 1993). This is the first application of a continuous logging multichannel system, using capacitance probes buried at different depths to record soil profile water content changes for flood studies. Due to its simpler electronic circuitry the capacitance probe is considerably cheaper than the TDR for individual applications, whilst the ability to multiplex the TDR makes it more appropriate where measurements are required at a number of points. The two techniques have been reviewed (eg Gardner et al, 1991).

Capacitance probes, in contrast to other instruments such as pressure transducer tensiometers, record water content directly and function over the whole range of soil water content encountered in the field, and only require attention once per year for battery replacement. The application of the capacitance probe to soil moisture studies capitalises on the very large difference between the dielectric constant of water (approx 80) and that of air (unity) and soil (approx 2 to 4 depending on the material). This makes the dielectric constant of the bulk soil (ie soil, air and water) very sensitive to changes in water content. The capacitance probe measures the dielectric constant by inserting two stainless steel electrodes into the soil. The electrode rods and the soil between them form a capacitor, and together with the probe body which contains a battery powered oscillator they form part of the oscillator circuit. The oscillator operates at about 150 MHz in air, and the frequency of the whole circuit (electronic components, rods and bulk soil) varies from about 90 MHz in wet soil to 140 MHz in dry soil.

Field sites

The river Ock catchment (234 km2) is a rural catchment in S England, some 30 km south west of Oxford (Figure 1). It has natural flows which are measured by a weir at Abingdon. The catchment comprises permeable chalk uplands (which sustain summer baseflow) and a central valley covered with relatively impermeable clay soils, which are the source areas of storm runoff and result in the catchment having a long history of flooding. The long term average annual precipitation is approximately 650 mm and the streamflow averages 200 mm. Daily catchment average areal rainfall was calculated from a network of five long term gauges, using isohyets. The Penman potential short grass evaporation is about 500 mm with a pronounced summer peak in June to August and low values from October to March.

Soil water monitoring has been conducted at two sites; each has a raingauge, three capacitance probes buried in the soil at 5, 15 and 45 cm depths, and three purgeable pressure transducer tensiometers at 15, 25 and 50 cm depths. A dedicated interface controls and links the sensors to a solid state logger. Since electronic signals from the probe would affect its electrical field, and hence the dielectric constant, each probe is connected to the logger by fibre optic cables. The oscillation frequency of the probe is converted to an optical signal which is transmitted from the probes and reconverted to an electrical signal which is received by the logger.

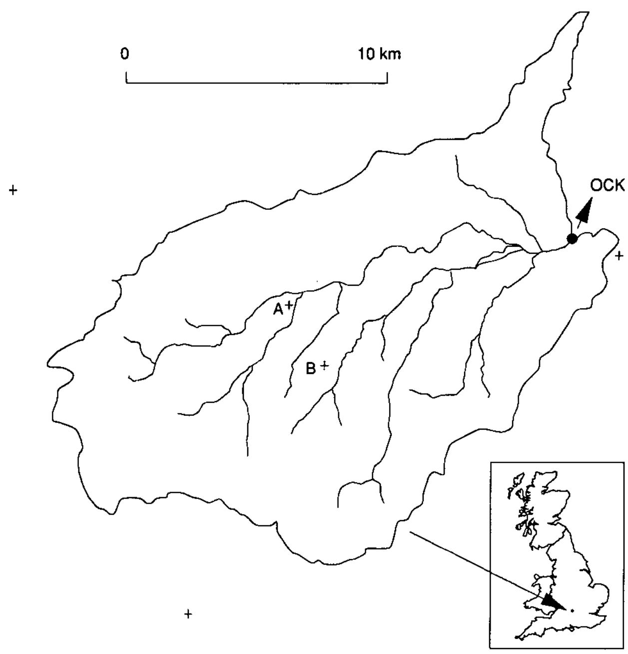

Fig 1

Figure 1. Catchment map showing the location of the measurement sites, including the streamflow gauge (•), raingauges (+) and soil water stations (A, Stanford Park; B, Challow Hill).

The soil water stations were sited on surface water gley soils in the central clay soil part of the Ock catchment, which provides the bulk of the storm flow response. The instruments at both sites are on permanent pasture, away from localised effects of stream channels, ditches and field drains, and the ground is fairly flat. The Stanford Park Farm site, lies in the valley bottom and is on Kimmeridge Clay (Jurassic period clays and clay shales). It has permanent grass, cut once per year for hay in summer and then used for grazing. The second field site, about 2 km to the south, is situated on higher ground at Challow Hill Farm, and is on Gault Clay (Cretaceous period, mainly clayey).

Due to the spatial variability of soil water, the global soil water storage of a catchment is unknown. However, manual readings at a network of sites in the catchment over a year (Hasnip, 1993), indicated that there is considerable similarity in the time varying seasonal patterns of near surface soil moisture across the catchment. Water content changes were well correlated and this finding indicates that each soil water measurement station can provide an index of the changing state of ground wetness over the catchment.

Calibration of the capacitance probes

Field calibration of all the capacitance probes (frequency to water content) was carried out. The sites are visited twice each month to download the loggers and check the instruments. The tensiometers are purged if required, and independent measurements made of soil water content (by neutron probe and gravimetric analysis). For the 5 and 15 cm probes, soil samples were taken to determine the volumetric water content by the thermogravimetric technique. For the deepest probes (45 cm) correlations have been made with water content measurements using a neutron probe. To avoid site destruction during the gravimetric sampling, the surface soil water contents at the location of the buried probes were measured using a hand held Surface Capacitance Insertion Probe (SCIP) (Robinson and Dean, 1993). Further SCIP readings were then taken at 5 to 10 metres distance until a similar value was obtained, indicating a location where the soil water content was the same. Here the soil sample was taken.

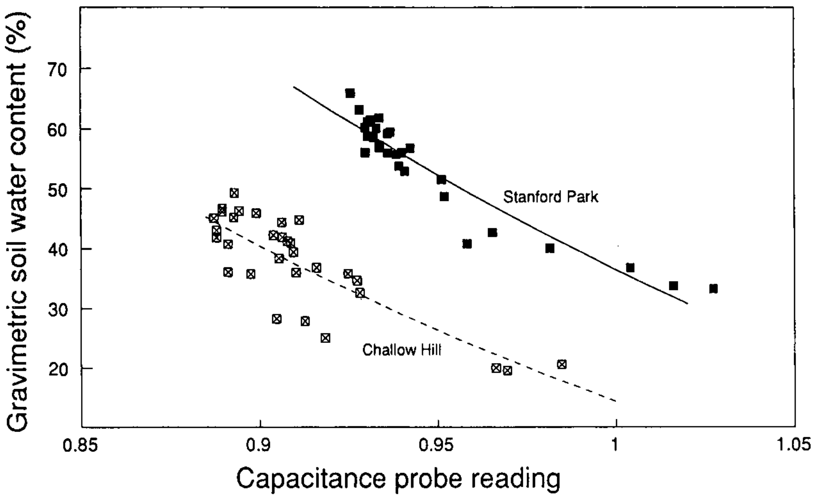

Figure 2. Field calibration curves for the 15 cm depth capacitance probes.
Upper curve - Stanford Park R2=90%
Lower curve - Challow Hill Farm R2=68%

Separate calibrations were obtained for each site and for each depth. The calibration curves show a high correlation despite the necessary transposition in space between the buried probes and the location of the gravimetric sample. For 5 cm depth and 15 cm depth probes 70 to 80% of the variance in water content was explained. The fit was poorer for 45 cm, where the very limited range in observed water contents (5%) was of similar magnitude to the errors of the neutron scattering method. Calibration curves (between the capacitance probe reading and the gravimetric soil water content) are given in Figure 2 for the two capacitance probes at 15 cm depth, since that depth proved to be the most useful in the streamflow modelling described later. The difference between the two sites is due to factors including soil density, texture, and organic matter content which influence the dielectric constant (eg Jacobsen and Schjønning, 1993). The soil at Stanford Park has a higher clay content than that at Challow Hill (50 vs 35%) and a lower bulk density (0.8 vs 1.2), factors which act to reduce the bulk soil dielectric constant and so increase the reading at a given water content.

APPLICATION OF SOIL WATER DATA TO STREAMFLOW PREDICTIONS

The benefit of using measured soil moisture data for streamflow prediction was investigated using a lumped black box time series rainfall runoff model, IHACRES (Jakeman et al, 1991; Littlewood and Jakeman, 1994). The model structure comprises a non-linear loss module which generates rainfall excess from areal rainfall and air temperature, followed by a linear rainfall excess streamflow module (Figure 3). This separate treatment of net rainfall calculation and subsequent routing makes it relatively easy to introduce a new net rainfall module based on soil water content. In normal application of the model the loss module calculates catchment wetness s, as a function of rainfall r, and temperature t. To calibrate this component of the model the optimum values of two parameters were determined. These comprise w, a time constant describing an exponential decrease in s due to evapotranspiration in the absence of rainfall, and f, a temperature modulation factor which quantifies how w changes per degree Celsius change of temperature. Net rainfall u, at time k, is then calculated by:

$$u_k=r_k*0.5(s_k+s_k-1) \qquad (1)$$

The computed net rainfall is then input to a system of linear storages in any series and/or parallel configuration to estimate streamflow. The standard configuration of the rainfall excess-streamflow module was used, which comprises two parallel flow components (see Figure 3). The model calibration ensures that the computed rainfall excess equals the volume of observed flow.

Model calibration without measured soil moisture data

Daily rainfall, streamflow and temperature data over the 10 year period (September 1982 to August 1992) were used for calibration. However, this proved to be

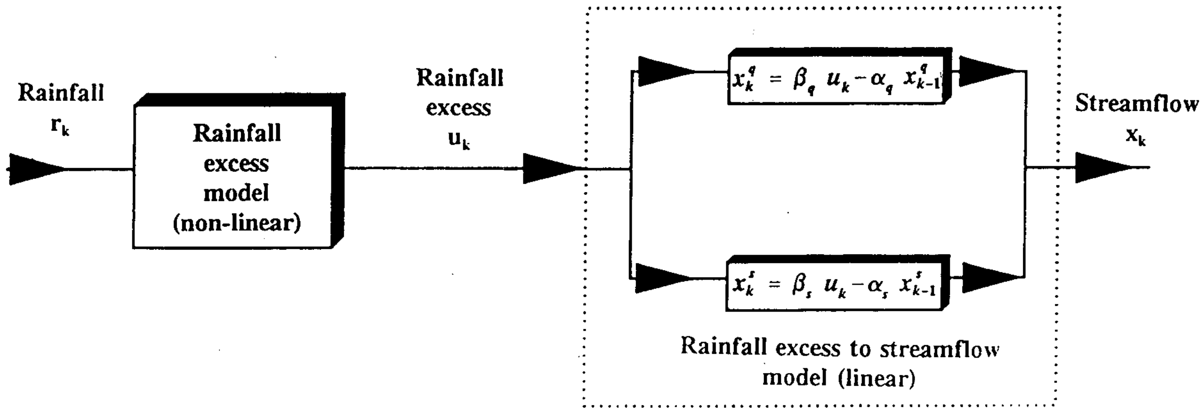

Figure 3. The IHACRES model structure using a parallel configuration of storages for quick and slow flow in the rainfall excess - streamflow module.
xq is quick flow
xs is slow flow
x is total streamflow
and ß are parameters for quick and slow flow which are determined by model calibration.

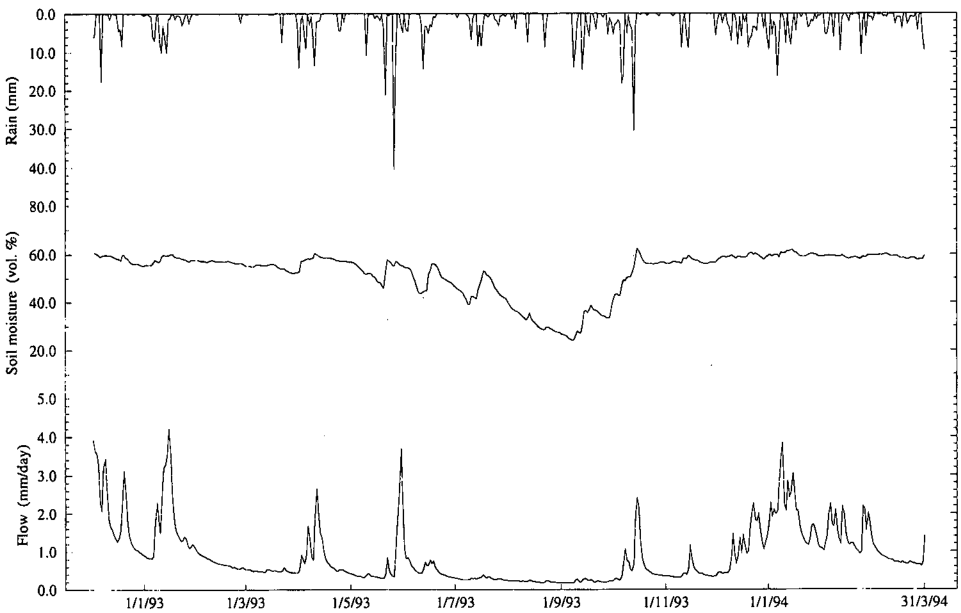

Figure 4. Daily rainfall, streamflow and measured soil water content (Stanford Park) 1992-1994.

too long a period for model calibration. Accordingly, the record was divided in four overlapping subperiods, each of three years length, which were independently calibrated. Experience with IHACRES has shown that three years is a suitable length for a satisfactory calibration of this model - it generally contains a sufficient variety in weather and flow conditions, but not too many data to prevent a model fit (Littlewood & Jakeman, 1994). Using subperiods enabled the stability of the model parameter values to be assessed. Due to the model `volume forcing' the predicted flows to those observed the calibration period should be chosen to begin and end with similar stream discharge (it being assumed this would ensure similar storage); and it is generally preferable to start the model during a time of low flow. The subperiods were chosen to start from 1st September and end on 31st August since the end of August was usually a time of low streamflow. Since some years did not have suitably dry summers to start or end the model calibration, there is an overlap between some of these subperiods.

Table 1. Calibration fits for each subperiod, showing the influence of flow conditions (mean and coefficient of variation) on net rainfall model parameters values, f and w .

Subperiod	f	τ_w hrs	R^2 (%)	ARPE (%)	Mean flow m^3s^{-1}	Coeff Var.
1) 1982-85	3.4	1.9	75.8	0.083	1.556	0.974
2) 1984-87	3.5	1.6	74.6	0.084	1.628	0.957
3) 1986-89	3.4	2.4	78.0	0.085	1.543	1.071
4) 1989-92	4.0	3.8	86.4	0.055	0.855	1.383

Table 2. Simulation fits for each subperiod and for the whole period September 1982 to August 1992, using the model parameter values derived for the four subperiods.

Subperiod	1982-1985		1984-1987		1986-1989		1989-1992		1982-1992	
	R^2	Bias	R^2	Bias	R^2	Bias	R^2	Bias	R^2	Bias
1	xxxx	xxxx	73.1	.00	74.1	.21	49.6	-.35	72.1	-.04
2	76.8	.00	xxxx	xxxx	74.2	.22	44.0	-.35	71.8	-.03
3	67.0	-.26	64.9	-.25	xxxx	xxxx	77.7	.00	63.3-	.25
4	57.6	.32	53.1	.37	54.7	.51	xxxx	xxxx	63.0	.29

Table 3. Goodness of fits for flows simulated March 1993- March 1994, with net rainfall model parameter values optimised using: a) Rainfall and runoff data in the 4 subperiods (Eq.1), b) Measured soil water data 1993-94 (Eq.4)

Period:	1982-85	1984-87	1986-89	1989-92	Soil water data
R^2(%)	80.3	81.6	81.5	61.9	87.1
Bias (m^3s^{-1})	0.23	0.23	-0.03	0.65	0.01

Table 1 shows the optimum combination of net rainfall parameters, for each subperiod and the goodness of flow fit determined by two measures, the coefficient of determination,R2 and the ARPE (Average Relative Parameter Error).

The optimum parameter values for subperiods 1 and 2 are very similar, while those for subperiod 4 are quite different, and those for subperiod 3 differ to a lesser extent. As in most models, the optimised parameter values are affected by the weather conditions during the calibration period (the number and magnitude of storms influencing the efficiency of fit of high flows, for example). The most recent subperiod (1989 to 1992) was particularly dry, which may account for the difference in the parameter values obtained compared with those for the first two subperiods.

Subsequently, these four sets of calibrated parameter values were used in simulation mode for the other three subperiods and for the whole period 1982-1992 (mean flow 1.353 m3s-1, coefficient of variation 1.203). This was to establish how well the values represent the long term catchment behaviour (Table 2). The fit is measured in terms of R2(%) and the volume bias (the difference between the observed and modelled mean flow in m3s-1).

Not surprisingly, the very similar parameter values from subperiods 1 and 2 gave similar fits when interchanged, and also when applied to the whole period, 1982-92.

The fit was poorer when the parameter values from the later subperiods were used, particularly subperiod 4.

These parameter values were then applied to the period for which soil water measurements were available (December 1992-March 1994). The initial period December 1992-February 1993 had to be omitted, however, since the model had problems with the `wet start' in December 1992. The period was reduced to 1st March 1993-31st March 1994, providing a period starting and ending with similar streamflow and, it would be assumed, catchment storage. This was a period of high flows (mean 2.13 m^3s^{-1}, and CV of 0.864), and it was fitted best using the parameter values derived from subperiod 3 (Table 3). Unfortunately it was not possible to obtain a calibration of the model on this year alone, since it was too short a period of time.

Applying measured water contents

The structure of the IHACRES model meant that it was relatively easy to replace the standard rainfall loss module based on rainfall and temperature to estimate catchment wetness with a net rainfall filter based on the measured water contents. Several functional relationships of varying degrees of complexity were investigated. In the first instance, the net rainfall at time k was taken to be the product of gross rainfall and the volumetric moisture content, w, in an analogous form to equation (1):

$$u_k=r_k*w_{k-1}$$

Since there was so little variation in moisture content for the 45 cm deep capacitance probes, this model was only applied using the 5 and 15 cm depth probes on each site. Previously, it had not been possible to calibrate IHACRES in its standard form on only one year of data, but by using this simple modification to the net rainfall calculation it was possible to obtain calibrations in three of the four cases (the exception being the shallowest probe at Hill Farm). This ability to achieve a model calibration over a much shorter length of data was the first practical benefit of soil moisture data for streamflow prediction to arise from the project.

Not surprisingly, given the simple nature of the filter the calibration fits were modest (with R^2, values of 0.68 and 0.62 for Stanford Park 15 cm and 5 cm deep probes respectively, and 0.56 using Hill Farm 15 cm; the ARPE varied between 0.37 to 0.64). To improve the calibration fit further, a more realistic function is required: it may be hypothesised that the relationship between storm runoff volume and soil wetness will not be linear, but rather of a sigmoid (s-curve) form. This may be justified as follows. However dry the catchment soil becomes, there will always be a small runoff response due to direct channel precipitation. As the catchment becomes wetter streamflow will slowly increase from riparian areas. Progressively more areas will contribute, and streamflow will increase at an increasing rate. Eventually the rate of increase will lessen and ultimately there will be an upper limit as some parts of the catchment (hill tops, very permeable soil, etc) will never contribute to stormflow. This picture is quite compatible with the catchment studies of stormflow generation and contributing areas, described earlier, and with the modelling work based on topographic form such as the approaches of Kirkby (1975) and O'Loughlin (1986).

There are a number of exponential and logarithmic functions which incorporate a sigmoid shape for part of the range. Several forms were examined, and the one selected here is:

$$y= b \exp(-(ax)^4)$$

For a, b >0 this function monotonically increases from $x=-\infty$ to $x=0$. The non-linear filter for the proportion of gross rainfall becoming excess rainfall, sk, has been thus been established in the form:

$$s_k= b \exp(-(a [w_k - 0.7])^4) + c$$

where a,b,c>0, and b+c=<1 ensure the excess rainfall never exceeds the gross rainfall. This filter has been applied to the 15 cm depth water contents at Stanford Park, since this probe had the strongest correlation to the streamflow. The maximum recorded moisture content was 0.7, so this value is subtracted from x (ie the measured moisture content) to ensure that the net rainfall proportion rises to a maximum value at this water content. There are three parameters to be optimised, of which two of them are semi-physical and their approximate values can be determined. Parameter b, is the maximum proportion of net rainfall when the water content is at its maximum, and parameter c represents the underlying baseflow component, and is necessary to make allowance for the groundwater contribution to total streamflow from the chalk areas of the basin.

The best calibration fit using soil water data for the period March 1993 - March 1994 had an R^2 of 87.1% and an ARPE of 0.117. This is much better than using the

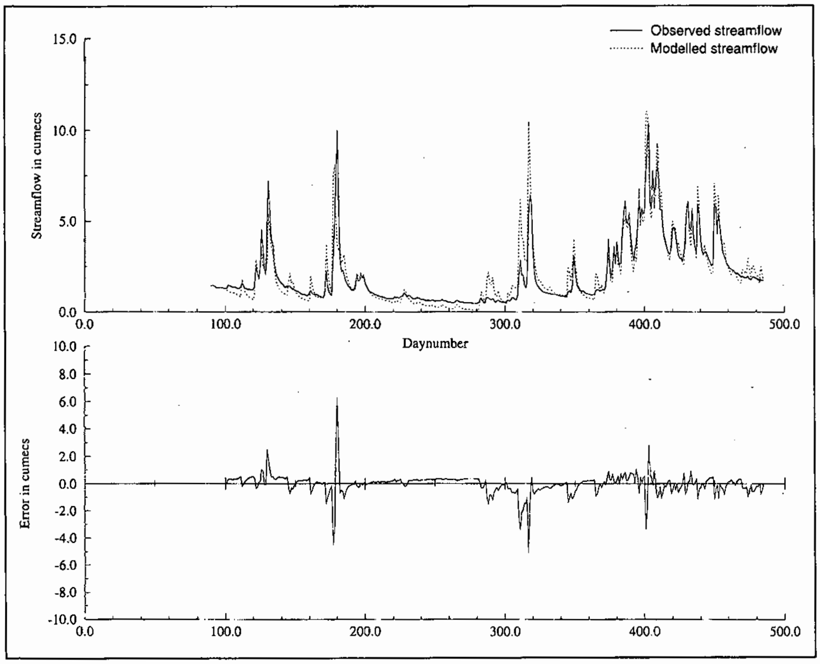

Figure 5a. Simulation for March 1993-March 1994, using IHACRES with the optimum parameters of subperiod 3.

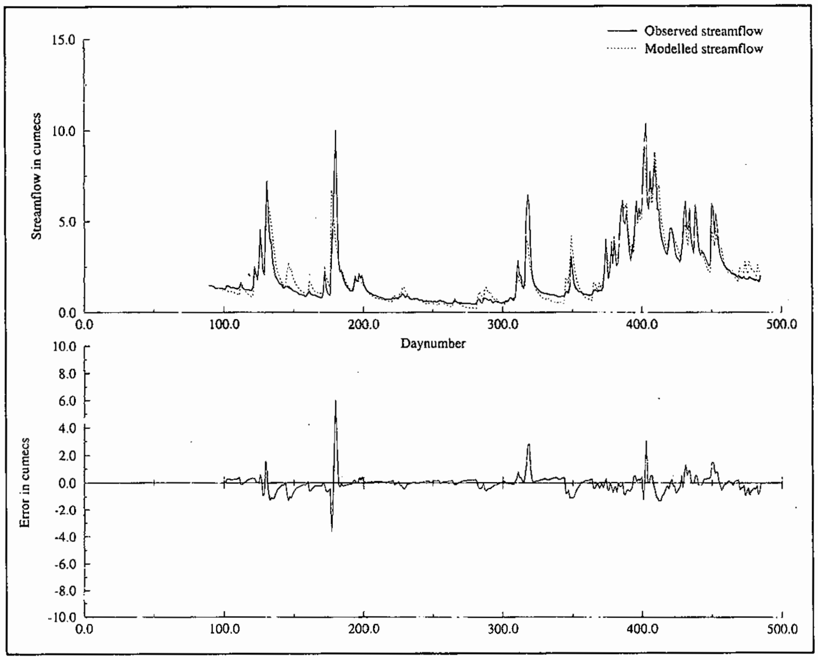

Figure 5b. Calibration for March 1993-March 1994, using the net rainfall filter based on measured soil water contents.

standard model with the best set of parameter values derived from the three year subperiods (Table 3). Figure 5 shows the fit between observed and modelled streamflow for a) standard model with the best parameter values obtained from subperiod 3, and b) calibration using measured soil moisture. It is obviously desirable to have a longer period of soil water data in order to be able to compare the predicted and observed flows for an independent period, and this remains the long term goal of this project. Additionally, it is hoped that advances with remote sensing of soil moisture (eg Cognard et al, 1995) may one day enable extrapolation of the point measurements of ground wetness across a whole catchment.

CONCLUSIONS

This study has demonstrated that the recent advances in instrumentation make it possible to collect continuous soil moisture data, and they may be used in rainfall runoff models to improve rainfall-runoff model performance in several ways:

Firstly, because of the extra constraints put onto the model behaviour it is possible to calibrate a streamflow model on a shorter period of observation.

Secondly, the extra information can improve the model calibration fit.

The time series of monitored soil water data currently available are too short to be used in an independent test of the model in simulation mode, but it is intended that the data collection will be continued long enough for this goal to be achieved.

The results presented here use a very simple non linear s-curve relation between the net rainfall proportion and the water content at a single depth in the soil profile. More sophisticated analyses using the depth varying water content profile may further improve streamflow predictions.

ACKNOWLEDGEMENTS

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Monitoring soil water content variability in the Cal Parisa basin (Alt Llobregat) with TDR. Experimental design and first results

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ABSTRACT

A small catchment on abandoned fields in the Pyrenees was monitored with the TDR method to evaluate the temporal and spatial variations of soil water content in relation with the hydrological response. The first results obtained over one year show a series of three periods of increasing wetness and three periods of decreasing wetness. Frequently saturated zones show high moisture values and stability over several months, mesophile grassland zones show a wide and variable moisture range, and forested zones show the lowest soil moisture contents.

Key words: TDR, experimental catchment, abandoned fields, Pyrenees, soil water content.

1. INTRODUCTION

Studies in the Cal Parisa basin (Alt Llobregat) showed the great importance of the antecedent soil water content conditions in its hydrological functioning (Llorens, 1991). On the other hand, the abandoned agricultural terraces play an important role in the spatial distribution of saturated areas (Llorens et al. 1992, Gallart et al. 1994). These results showed the need for an accurate evaluation of temporal and spatial variations of soil water content with special emphasis on the role of the topographic structure and abandoned terraces, together with the effect of the forest expansion.

The soil water content monitoring design needed to fit the following requisites: 1) to measure a wide range of soil water content; 2) to be repeatable and non destructive, allowing different measurements at the same point; 3) a sufficient number of measurement points to obtain a good assessment of spatial soil water content variation; 4) to measure surface soil water content; and 5) economical, easy to install and quick to measure sensors.

The TDR method has been considered the optimum, because it does not involve environmental problems, allows to install a large number of probes at low price without soil calibration problems.

2. MATERIAL AND METHODS

2.1. The study area

The Cal Parisa catchment is located in the eastern Pyrenees, in the headwaters of the Llobregat river valley at 1400 m altitude (Fig. 1). This small catchment has an area of 36 ha. The climate is Mediterranean mountainous with a mean annual precipitation of 850 mm. The mean temperature is about 9°C (Llorens et al. 1992). The catchment shows four main geo-ecological units (Llorens, 1991): a) Abandoned terraces (40 % of surface); b) Bare limestones (6 %); c) Slope zone (17 %) and d) Old mudflow (36 %).

The abandoned terraces were built to supply cereals to

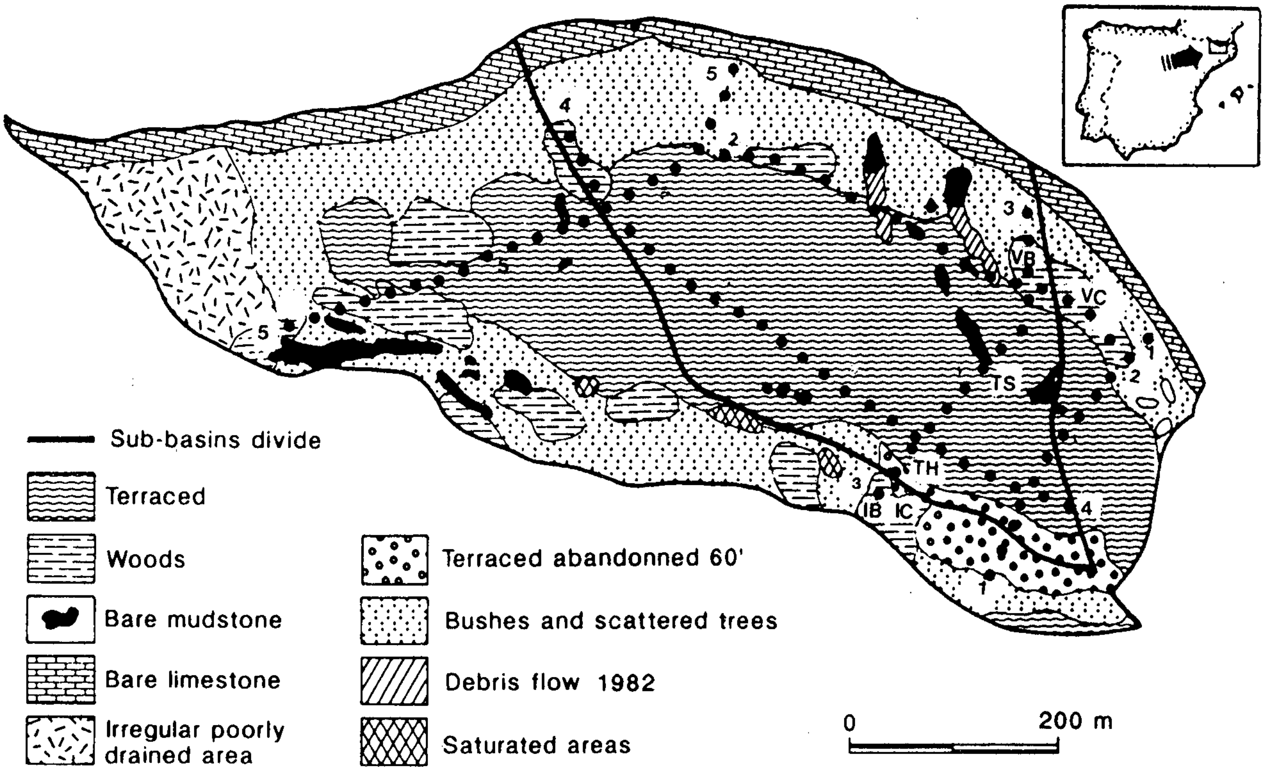

Figure 1. Main geoecological units of Cal Parisa experimental catchment (Llorens, 1991) with location of TDR profile stations: TH, wet terrace; TS, saturated terrace; IC, divide clearing; IB, divide forest; VC, hillslope clearing and VB, hillslope forest. The pointed lines show the five transects with TDR probes.

feed the increasing population during 19 th. Century (Gallart et al. 1994). The construction of terraces in such a clayey area needed the opening of artificial channels to drain water from saturated areas and to impede runoff across the terraces. Nowadays these terraces are abandoned and the area is used for cattle browsing.

The second geoecological unit (bare limestone) represents the northern and eastern divides of the catchment and belongs to an anticlinal flank.

The third unit (slope zone) is located near the bare limestone unit and shows a steep gradient towards the South (30 to 40 %) with some old small terraces.

The fourth unit (old mudflow) is a small range which separates two subcatchments (Fig. 1). Every catchment shows a different drainage system. The left (eastern) basin is drained by an artificial net of channels, whereas the right (western) exhibits a less modified drainage net.

The forest (Pinus sylvestris) overgrows in this area with gradual abandonment, the forest surface increased from 5% to 25% of the whole area during the period between 1967 and 1988.

The Cal Parisa soils are clayey with a significant content of expansive clays (Smectite). The high content of organic matter of these soils, together with the saturation by calcium involve a great structural stability in their shallow horizons (Solé et al. 1992, Pérez 1991). Precipitation, weather and runoff data are recorded in the Cal Parisa catchment since june 1989 (Llorens, 1991).

2.2. Method

The Time Domain Reflectometry (TDR) was designed to find breaks in buried cables. The TDR cable tester measures the double travel time needed by a high frequency electromagnetic pulse to go and return along a cable. The speed of this wave depends on the relative dielectric constant which is characteristic of the insulating material. If wave guides are inserted into the soil, this acts as insulating medium, and its dielectric constant depends on the influence of each soil phase (water, air and mineral solids). The dielectric constants of these phases are: 80.36 (at 20 C°), 1 and 3 to 5 respectively. This great contrast between the dielectric constant of water and the rest of the soil phases produces a good relation between soil dielectric constant and moisture content, the higher is the soil water content, the longer time the electromagnetic wave takes. This idea was developed by Topp et al. (1980).

To measure soil water content with TDR, a probe made of steel rods is inserted into the soil. This probe is connected to the TDR by a coaxial cable. There are different models of probes depending on the number of rods (2, 3, ...), distance between rods, and length of the rods. We use a probe of three stainless steel rods which have the same length and are parallel on a same plane (Zegelin et al., 1989). The central rod is connected to the central conductor of the coaxial cable and the lateral rods to the shield. In this way, the probe behaves like a coaxial cable into the soil, without significant impedance break. The relative dielectric constant is obtained from the TDR signal by:

$$K = (Va / Vs)^2 = (ts / ta)^2$$

K= soil relative dielectric constant.
Va= wave speed in the vacuum.
Vs= wave speed in the soil.
ta = travel time in the vacuum.
ts = travel time in the soil.

To obtain the soil water content from the dielectric constant there are two main kinds of models: a) empirical models (Topp et al. 1980) and b) mixture models. We used a mixture model based in Roth et al. (1990) which has given the best results in our experimental data:

$$K = Kw + (1 -) Ks + (-) Ka$$

or

$$\Theta = (K - (1 -) Ks - Ka) / (Kw - Ka)$$

Θ = volumetric moisture content (cm^3/cm^3)
η = soil porosity (cm^3/cm^3)
K_a= relative dielectric constant of the air
K_s= relative dielectric constant of the solids
K_w= relative dielectric constant of the water
α = geometry parameter

To calculate the soil water content with this model we assumed the following values: a) porosity has been taken equal to the moisture of saturated soil or n = 0.5 if saturation was never reached; b) = 0.5 which indicates an isotropic soil; c) Ks = 4; d) Ka = 1; e) relative dielectric water constant depends on water temperature (Handbook of Physics and Chemistry, 1986):

$$K_w=78.54 [1 - (4.579\ 10^{-3}(t - 25)+1.19\ 10\text{-}5(t - 25)^2 - 2.8\ 10^{-8}(t - 25)^3)]$$

t = temperature (°C)

2.3. Monitoring

The instrument used to measure TDR signal was the Tektronix 1502C Metallic Cable Tester. The catchment monitoring design to study the field soil water content consisted of two different systems: transects and profiles.

1) Transects: These were designed to evaluate surface soil water content (from 0 to 20 cm) along different environments in the catchment. We installed 108 probes along 5 transects (Fig. 1). These transects are 2160 meters long and cover about 30 ha. The measurements have been made every three months. The probes consisted of three rods with 8 mm diameter, 200 mm length and 40 mm of separation.

2) Profiles: The objective of these profiles was to evaluate temporal variation of soil water content and water storage and to characterize these variations in selected representative catchment environments. Six stations were installed (Table 1) where measurements have been made every week. These probes had three rods with 6 mm diameter, 200 mm length and 25 mm of separation; these probes are permanently buried in the soil, with only the coaxial cables emerging from it.

Every station comprises four probes of 20 cm, vertically inserted in the soil at four different deeps (0-20, 20-40, 40-60 and 60-80 cm). The stoniness of the soil forced us to make boreholes with an auger and afterwards to introduce the same soil without stones bigger than 1 cm before inserting the probe. The surface probes were directly introduced in the natural soil.

2.4. Calibration

We carried out three kinds of tests with the TDR method to verify its validity and the error range: 1) a laboratory test with a disturbed soil cylinder; 2) an evaluation of the bulk density and the gravimetric soil water content in the field; and 3) an operator error analysis.

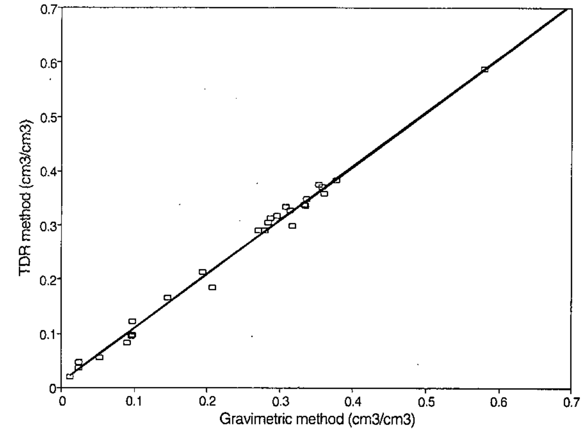

Figure 2. Volumetric soil water content obtained by TDR versus volumetric water content obtained by gravimetric method in the laboratory.

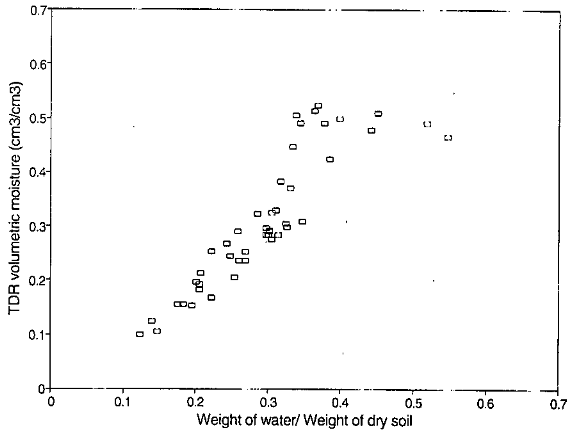

Figure 3. Comparison between gravimetric soil water content and volumetric soil water content (TDR method) near "Wet terrace" TDR station (0 a 20 cm). Each method was applied to different points, this fact explaining the high scatter of data.

2.4.1. Laboratory test

A soil sample taken near the "Wet terrace" TDR station was dried and sieved to 2 mm. Soil was introduced and packed in a container of known volume (3632 cm3). The sample soil bulk density was 1.1 g/cm3 and the porosity 0.59 cm3/cm3. We performed both a wetting and a drying sequences. Before taking TDR measurements we recorded the sample temperature to control the dependence of water dielectric constant.

The regression between two methods (gravimetry and TDR) gave us a very low average error (0.0125 cm^3/cm^3). The intercept is not significantly different from 0 and the slope is not different from unity:

$$\theta\ TDR = (0.995\ \theta_{grav}) + 0.010$$

θ_{grav}= Volumetric moisture (gravimetry) (cm^3/cm^3)
θTDR= Volumetric moisture (TDR) (cm^3/cm^3)

R squared $R^2 = 0.995$
Number of samples= 28

2.4.2. Field verification

During the period 17/02/1993 to 06/05/1994, every week we took 0-20 cm deep soil samples near the "Wet terrace" TDR station to compare the gravimetric and the TDR methods and to estimate the bulk density of these soils and its possible variation for different moisture ranges. We obtained 45 data pairs from two methods. The results were (Fig. 3):

Table 1. TDR stations installed in the Cal Parisa catchment to monitor soil water content in profiles 0 - 0.8 m depth.

Station	Vegetation	Gradient(%)	Drainage	Geomorphical unit
Wet Terrace	Mesophile grassland	1-20	Convergent	Terrace
Saturable terrace	Molinia coerulea	21-30	Convergent	Terrace
Divide forest	Grassland with Pinus sylvestris	11-20	Divergent	Old mudflow
Divide clearing	Mesophile grassland	11-20	Divergent	Old mudflow
Hillslope forest	Pinus sylvestris	41-50	Lateral	Little terraces
Hillslope clearing	Xerophile grassland	41-50	Lateral	Little terraces

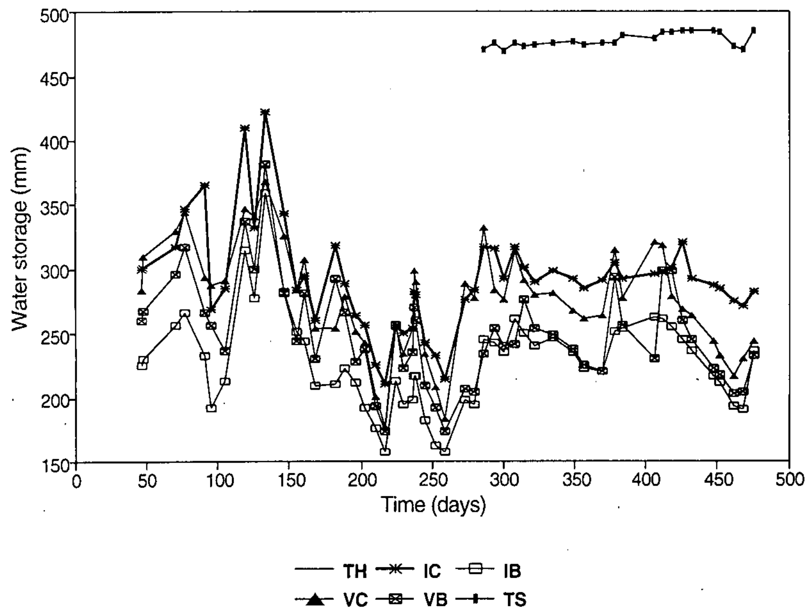

Figure 4. Total water storage from TDR station profiles (0 a 80 cm) during the period from 13/05/93 to 13/04/94. TH, wet terrace; TS, saturated terrace; IC, divide clearing; IB, divide forest; VC, hillslope clearing and VB, hillslope forest. TS was instrumented later (07/10/93).

1) The regression between the two methods shows a good dependence (R^2 =0.78) in spite of the spatial soil water content variability.

2) The dispersion between two methods is maximum when soil water content is high (when moisture is mainly controlled by soil structure), and minimum in low soil water content (when moisture is controlled by soil texture). Nevertheless, the relative error is nearly constant along all the soil water content range.
3) The slope of the regression between gravimetric soil water content and volumetric soil water content, which corresponds to soil bulk density, is 1.22 g/cm^3 (Fig. 3) and shows a good similarity with other field data. The linear regression shows a good fit between two methods, this would demonstrate that field bulk density is not significantly affected by moderated soil water content in these soils, although there is some uncertainty for high values of soil water content (fig. 2).

2.4.3. Operator error

To evaluate the error that the operator can make in the interpretation of the TDR curve, we repeated three measurements at the six TDR stations. First we made a measurement at these stations, and afterwards we repeated twice the measurements with the same order of the first course. The time between two successive measurements in the same point was about one hour.

The differences between these three measurements represent a standard error of ± 0.005 cm^3/cm^3. This error range represents about the half of the laboratory error.

3. RESULTS

Data were studied from TDR stations during the period May 93 to April 94. This period had normal weather conditions. The "Saturated terrace" was installed later than others (October 93). We are going to analyze both aspects of catchment soil water content: temporal variation and spatial distribution.

3.1. Catchment water reserve: temporal variation

We have compared the total water storage of TDR profile stations to analyze the temporal variation through the year:

$$S=\Sigma_1^4 \ (\theta_i \ \lambda_i)$$

S = Storage water volume in all the profile (mm)
θ_i= Volumetric water content at every depth (mm^3/mm^3)
λ_i= Probe length (200 mm)

Table 2. Maximum, minimum, mean and deviation of water storage for every TDR station (from 0 to 80 cm). Period May 93 - April 94. (1) The "Saturated terrace" station has a shorter recorded period (October 93 to April 94).

Station	Max. storage (mm)	Min. storage(mm)	Mean storage (mm)	Std. deviation (mm)
Wet terrace	407	198	291	48
Divide clearing	420	209	283	35
Divide forest	359	157	224	36
Saturated terrace [(1)]	484	448	476	8
Hillslope clearing	368	178	268	39
Hillslope forest	397	144	218	50

Over the year we observed six periods: three wetting periods and three drying periods (Fig. 4). The wetting periods occur during spring and autumn storms and during the snow melting at the end of winter. The highest water storage corresponds to spring (375 mm), the other two wet periods are similar in storage (280 mm in autumn and 305 mm in winter) (Fig. 4).

The drying periods occurred during transitions between altum and winter, winter and spring, and spring and summer. The lowest water storage was recorded during summer (180 mm), the second lowest during winter-spring (230 mm) and the last during altum-winter (250 mm). The quicker drying period was observed during spring. The other two periods are similar.

3.2. Spatial behaviour: profiles

We observed three groups of the same behaviour in the six stations according to total storage water of every station (Table 2) and the soil water content profile at every depth (Fig. 5).

a) Saturable zones: This group is represented by "Saturated terrace" station. Its maximum storage is high (476 mm). The moisture profile remained saturated and constant with a deviation of only 7.9 mm. Nevertheless there are no data of the late spring when the station is assumed to desaturate at the surface (Fig. 5). The TDR station has recorded soil water content values between 0.55 and 0.65 cm^3/cm^3.

b) Clearing zones: This group is formed by "Hillslope clearing", "Divide clearing" and "Wet terrace". Their mean storage is 275 mm, which is 200 mm lower than "Saturated terrace" one. On the other hand, we can observe along the profile of these stations, an increasing soil water content with depth (Fig. 5).

c) Forest zones: "Hillslope forest" and "Divide forest" TDR stations from this third group which has recorded the lowest soil water content values, with mean water storage of 220 mm and soil water content profiles between 0.24 and 0.29 cm^3/cm^3 (Fig. 5). Nevertheless, the water storage variation between the two forest stations are different: "Hillslope forest" presents a deviation of 46 mm and "Divide forest" of only 35 mm.

3.3. Spatial variation: surface transects

The profile stations data show a great variation in surface water content (0.02 to 0.13 cm^3/cm^3) while in depth they are more similar (0.04 to 0.08 cm^3/cm^3) (Fig. 5). For this reason we analyzed the surface soil water contents obtained in transect 3 to see the variation range. This transect contains 25 probes along 322 meters and crosses the basin from south to north connecting the following TDR stations:"Divide forest", "Wet terrace", "Saturated terrace" and "Hillslope forest". At the same time this transect crosses four different catchment environments: wet terraces, saturated terraces, forest zones and hillslope zones.

At this moment we have moisture data from eight different moments during the year. We choose the driest and the wettest points of 05/08/93 and 16/02/94, respectively (Fig.6). Along this transect we can observe: a) The hillslope zone has the lowest homogeneous soil water content A similar fact happens in the forest zones but

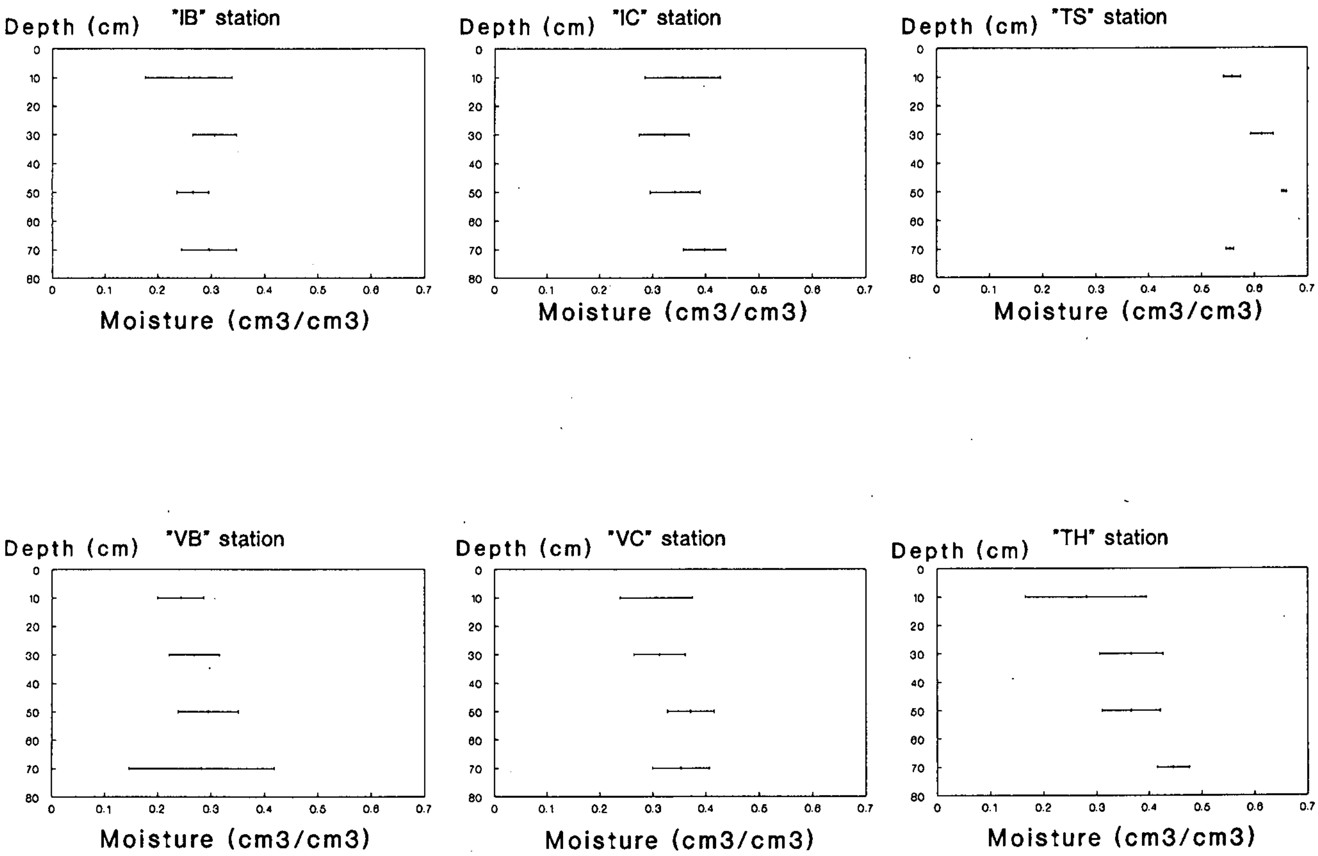

Figure 5. Soil water profiles (means and deviations) from Cal Parisa catchment during the same periods of figure 4.

the soil water content is a little wetter; b) The terrace zones show a higher soil water content with a great variation among recorded points along the transect. The maximum soil water content was recorded in the saturable zones covered by Molinia coerulea.

4. DISCUSSION AND CONCLUSIONS

4.1. Method

The soil water content error obtained in the laboratory with the TDR method (0.01 cm^3/cm^3) confirms that this is a good method to measure the soil water content. This error range includes the operator error (0.005 cm^3/cm^3) and the remaining errors can be attributed to mistakes in the parameter data used for the mixing function (porosity, geometric factor, dielectric solid constants and temperature). The computation of soil water content from TDR readings is sensitive to the porosity, which was obtained from saturated values. The presence of expansive clays in the soil could involve a variation of the porosity with dependence of soil water content, but the experimental data show that the surface soils with a great content of organic matter have a low variation in porosity and bulk density. Therefore the TDR application in this kind of soils does not need a continuous correction of the porosity parameter in the mixing function.

4.2. Monitoring

The moisture profiles do not attain a depth of constant under moisture. This fact indicates that we are underestimating the soil storage and its temporal variation. On the other hand, the soil volume where the measurement performed with non-surface probes (20 to 80 cm) was disturbed to allow the introduction of the rods. For this reason the recorded water contents represent some equilibrium with the surrounding natural soil, but the absolute values could be a little different. The deepest probe from the "hillslope forest" station gave some odd readings (see fig. 5) that can be caused by this problem.

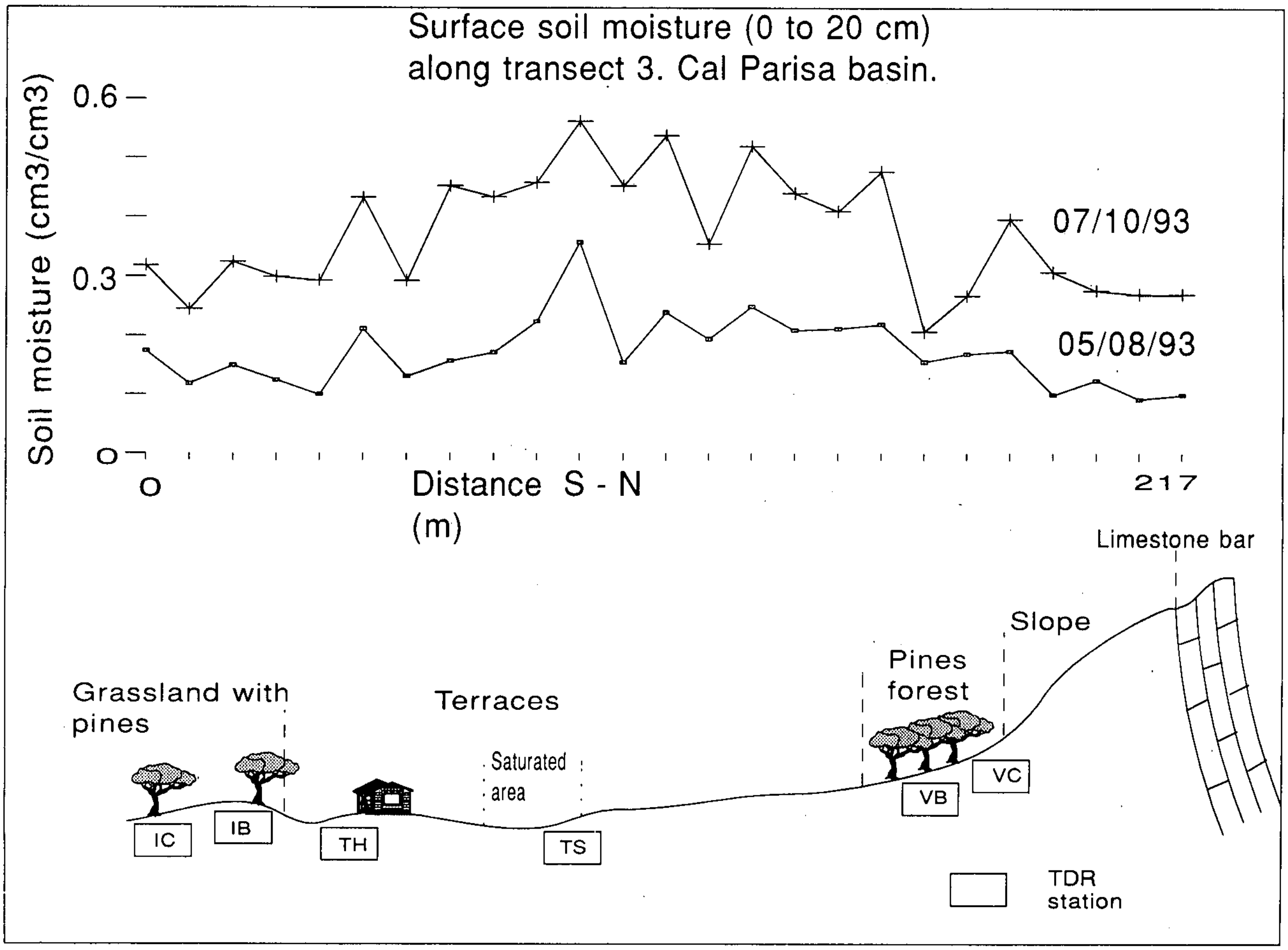

Figure 6. Shallow soil water content (0 to 20 cm) of the transect number 3. This transect traverses different catchment environments: pine forest, terraces, saturated area and hillslopes.

4.3. Results

The temporal variation of the soil water content in the Cal Parisa catchment shows three drying periods and three wetting periods. These periods are controlled by precipitation and evaporative potential of every season. The maximum water storage in not saturated stations was recorded during spring (375 mm) while the minimum was recorded during summer (150 mm).

The spatial variation of soil water content is very high. Three factors: vegetation cover, topography and terraces, seem to control this variation. These factors involve different processes, like a higher water loss (forest), and lateral transfer of water (from dry upper hillslopes towards saturable hollows). From wet to dry, the environments can be sorted as saturable terraces, wet terraces, forest zones and high hillslope zones of transect 3.

The temporal variation of water storage in every zone is found to depend on lateral drainage processes. For this reason in the high slope zones the deviations are very high (46 mm in "Hillslope clearing" and "Hillslope forest") and on the other hand, "Wet terrace", very near to a natural drainage line, recorded a high level of variation in water storage. On the contrary, the most constant water storage recorded has been "Saturated terrace" station which belongs to a saturable area by convergent hypodermic slow flow.

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Influence of wildland fire on surface runoff from a hillslope

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ABSTRACT

Surface runoff from three adjacent hillslope plots, unburnt or subjected to light or moderate prescribed burns, was monitored over a four-year period. All plots initially bore Ulex scrub. In the first year, runoff from the unburnt plot was about 3.5% of rainfall, while that from the burnt plots was between 5.2 and 6.6% of rainfall. Runoff from the burnt plots gradually dropped over the study period, and we estimate that preburn values are reattained within four to five years (corresponding to the time required for re-establishment of the vegetation cover). The water repellency of the soil was not appreciably increased by either light or moderate burns, indicating that wildfire-induced increases in runoff in this region are due largely to reduced interception as a result of the destruction of vegetation. Finally, although burning caused clear relative increases in runoff, the absolute increases were small (2 - 3% of rainfall), suggesting that wildfire-induced increases in runoff are unlikely to have major effects on the flow regimes of receiving rivers. The estimated postfire increases in infiltration, on the other hand, were much more marked, and it is possible that wildfire-induced increases in subsurface flow have important effects on watershed hydrology.

INTRODUCTION

One of the most evident effects of wildland fire on hillslope hydrology is to increase surface runoff. A number of authors have attributed this to a fire-induced increase in the water repellency of the surface layers of the soil (Ferreira, 1990; Imeson et al., 1992; Shahlaee et al., 1991). However, other studies have indicated that the physical properties of the soil are scarcely affected by fires of light to moderate intensity (Hudson et al., 1983; MarquÇs and Mora, 1992); in such cases, the time taken for surface runoff to return to prefire levels will be largely governed by the rate of recovery of vegetation cover and by the interception characteristics of the vegetation.

In the northwest Iberian Peninsula, Ulex scrub occupies a total of about one million ha. Wildfires are very frequent in these areas, as a result of the high flammability of this vegetation type (Casal et al. ,1984; Vega, 1985), the region's pronounced summer drought, and the high incidence of fire-starting due to negligence or criminal intent.. Ulex scrub has high biomass (25 - 60 t ha-1) and a characteristically stratified structure, generally comprising a herb layer (species such as Agrostis canina, A. setacea and Pseudarrenatherum longifolium), an intermediate layer (species such as Daboecia cantabrica, Erica cinerea and E. umbellata) and a shrub layer dominated by Ulex europaeus (Basanta et al., 1988). This well-developed stratification leads to high ground cover, with rainfall interception typically about 40%.

The characteristic pattern of postfire recovery of scrub communities of this type has been described by Casal et al. (1984). In the initial phase, typically lasting a few months, plant growth is largely restricted to resprouting of species such as Ulex europaeus and Agrostis setacea. Subsequently, the burnt area begins to be recolonized as a result of germination of heat-resistant seeds in the soil or of seed-led or vegetative expansion of nearby vegeta-

Table 1. Physicochemical characteristics at depths of 0-2.5 cm, 2.5-5.0 cm and 5.0-10.0 cm. of the soil at the study site.(standard deviation, into brackets)

	Depth		
	0-2.5	**2.5 - 5.0**	**5.0 - 10.0**
pH H20	4.44(0.14)	4.40(0.07)	4.41(0.07)
pH ClK	3.48(0.19)	3.51(0.06)	3.68(0.07)
% C	6.74(0.55)	7.91(1.18)	7.42(0.26)
% N	0.68(0.10)	0.58(0.06)	0.54(0.04)
C/N	10.1	14.0	13.8
Inorganic nitrogen (mg/100 gr)			
N-NH4+	2.00(0.79)	2.00(0.02)	1.82(0.16)
N-N03-	0.32(0.09)	0.32(0.04)	0.46(0.05)
Exchange cations (mg/100 gr)			
Na	5.01(0.75)	3.57(1.30)	3.30(0.88)
K	6.81(1.26)	3.73(0.60)	2.76(0.53)
Ca	4.56(1.16)	1.58(0.80)	0.62(0.40)
Mg	1.18(0.37)	0.13(0.10)	< 0.1
Total	17.57(2.87)	9.04(2.53)	6.78(1.75)
Available P (Bray-2) (mg/100 gr)			
2.75(1.29)	1.91(0.59)	1.20(0.35)	
Particle size distribution (%)			
Coarse sand	38.89(0.71)	38.81(4.17)	39.06(2.68)
Fine sand	26.31(1.41)	25.66(1.19)	27.96(1.65)
Coarse loam	8.32(0.22)	9.13(2.00)	8.42(0.98)
Fine loam	11.15(0.15)	11.93(0.94)	10.11(0.77)
Clay	15.32(0.78)	14.46(2.18)	14.44(1.09)

tion. Eventually, the scrub becomes so dense that the majority of the opportunist herb species disappear. The loss of this vegetation cover following burning leads to a marked increase in the proportion of rainfall reaching the soil surface, and a consequent increase in the risk of erosion due to surface runoff.

In the work reported here we monitored surface runoff from three Ulex-scrub hillslope plots (one unburnt, one subject to a light burn and one subject to a moderate burn) over a 4-year period. Interception, evapotranspiration and water repellency were also monitored, with the aim of investigating the causes of fire-induced modifications of surface runoff patterns.

MATERIALS AND METHODS

Three adjacent 20 x 4 m plots, all bearing 8-year-old Ulex scrub and with 30% slope, were marked out on the west-facing slope of Monte Pedroso (Santiago de Compostela, northwest Spain; 4¯ 53' W, 42¯ 54' N) and separated by metal sheets. The soil is an umbric Leptosol over semioriented granite. Basic physicochemical properties at depths of 0-2.5 cm, 2.5-5.0 cm and 5.0-10.0 cm are listed in Table I. Each value is the mean for 5 samples (_+ standard deviation). Each set of 3 samples (for 3 depths) was obtained by driving an open-ended 20 x 20 cm. metal box into the ground, and collecting all soil within the box at each depth. Both the visual homogeneity of the study plots and the low standard deviations of the physicochemical properties determined (see Table I) indicate that these samples are representative of the study area.

Precipitation was recorded with two rainfall gauges, one at the top and one at the bottom of the study slope. Troughs for runoff collection (feeding via a 1/9 divider

Table 2. Monthly rainfall and monthly maximum and minimum temperatures recorded over the study period.

		Month											
		S	O	N	D	J	F	M	A	M	J	J	A
	Rainfall (mm)	0.0	233.5	117.4	58.4	28.5	177.5	109.3	87.1	191.7	0.0	25.2	42.8
88/89	T. max. (¯C)	24.4	19.5	17.7	14.1	13.7	11.5	16.2	14.5	23.6	24.2	28.0	27.3
	T. min. (¯C)	12.7	10.6	8.4	4.9	3.8	3.0	6.4	6.2	12.1	13.0 •	15.7	5.6
	Rainfall	0.0	177.7	280.9	661.5	193.4	155.7	48.7	74.6	31.9	12.0	0.0	0.0
89/90	T. max.	24.4	22.2	15.0	14.7	12.0	9.7	18.1	15.6	22.3	21.7	28.5	28.5
	T. min.	12.0	12.0	9.6	9.6	4.8	6.5	6.9	7.3	11.3	12.4	14.9	12.2
	Rainfall	138.3	333.9	167.0	159.9	251.2	212.0	139.4	20.9	55.2	40.0	46.2	33.8
90/91	T. max.	25.4	18.2	14.3	11.9	11.8	7.3	14.0	16.1	21.6	21.3	24.3	26.7
	T. min.	14.5	10.9	6.3	3.8	4.4	2.4	7.0	6.6	9.7	11.3	14.4	14.9
	Rainfall	137.4	129.4	195.1	87.3	78.3	0.0	72.5	102.5	104.5	43.0	30.8	121.4
91/92	T. max.	24.0	16.9	14.1	13.7	12.4	10.1	16.2	16.6	21.0	19.8	26.2	24.3
	T. min.	14.3	9.1	7.3	6.0	2.0	3.3	6.1	7.8	11.2	11.2	14.4	14.2

to a collection tank outside the plot) were installed at the bottom of each plot. Each plot additionally contained two 0.11 x 4 m troughs (30 cm above ground level to avoid splash-in, feeding to a common collection tank outside the plot): for each rainfall event,apparent interception by vegetation was estimated as the difference between gross rainfall and rainfall collected in the troughs (throughfall). Note that apparent interceptation is likely to be higher than interception sensu strictu.(i.e., rainfall minus , troughfall + stemflow). Stemflow is probably significant in this vegetation type, due tp high density of ascending stems; however, major-stem morphology was scarcely affected by our prescribed burns , so that the observed effects of fire on apparent interception (see below) are likely to be indicative of effects on interception sensu strictu.

Monthly rainfall, and maximum and minimum monthly temperatures, for the four years of study are listed in Table II.

Prescribed burns were carried out in September 1988. Plot L was burnt downslope early in the morning, with the aim of simulating a light-intensity fire. Plot M was burnt upslope at midday, with the aim of simulating a moderate-intensity fire. In both cases maximum temperature reached at the soil surface was about 250¯C, with temperature rising only about 10¯C at a depth of 2 cm (Díaz-Fierros et al., 1990). Plot C (control) was not burnt.

Water repellencies of the 0 - 2.5 cm, 2.5 - 5 cm and 5 - 10 cm layers of the soil were determined by King's (1981), modification of Watson and Letey's method. Briefly, droplets of ethanol at increasing concentrations (0.2 M steps) are pipetted onto the soil until that concentration is reached at which the droplet remains unabsorbed for 10 sec or more. On the basis of the molar concentration of ethanol required, we classify water repellency as low (0 - 1 M), moderate (1.2ˇ-ˇ2.2 M), severe (2.4 - 3 M) or very severe (3.2 M or more). For the purposes of this study, we have considered only severe or very severe water repellency to constitute a significant barrier to infiltration.

Water repellency was determined before and 24 h after burning (3 samples per burnt plot). In addition, repellency determined in another 30 samples taken from the control plot at different times of year showed a close linear relationship with soil water content (Fig. 1). Soil water content was thus used to predict water repellency in the three plots over the four years study.

Soil water content was determined from soil water tension (as measured with tensiometers at depths of 2.5, 5, 10, 20, 40 and 60 cm)with the aid of calibration curves constructed previously. Since the bedrock lies 60 cm below the surface, our soil water data can be considered representative of the root zone.

RESULTS

Before burning, water repellency in both plots L and M was "very severe" in the 0 - 2.5 cm layer and "severe" in the 2.5 - 5 cm and 5 - 10 cm layers. This high water

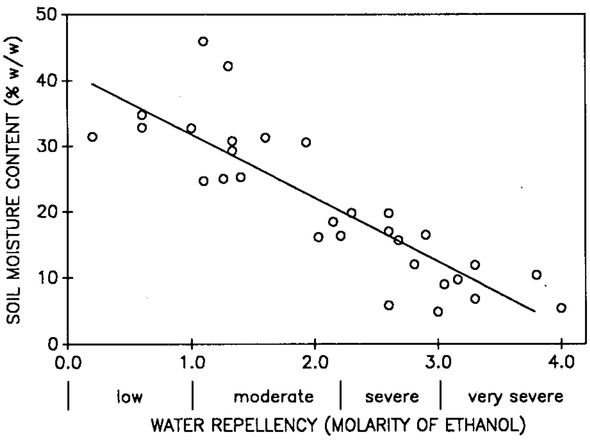

Figure 1. Water repellency (WR; estimated by a modification of Watson and Letey's method; see text) plotted against soil water content (SWC) for 30 top 2 cm soil samples taken from the study site at different times of year. Linear regression gives WR = 4.14 - 0.97 SWC, with r2˘=˘0.74.

repellency is as expected given that soil water content was very low (below wilting point) on the day of the measurement (10 September 1988). Twenty four hours after burning, water repellency in plots L and M was severe (2.9 M ethanol) in all three layers. Burning did not therefore lead to any increase in water repellency.

Actual evapotranspiration (ETa) over 16 rain-free periods (of between 2 and 9 days, in June 1989, March 1990 or May 1990) was determined for plots C and L on the basis of tensiometer measurements and of potential evapotranspiration (ETp) over these periods (estimated by the Penman method as per Smith, 1990). Mean ETa/ETp ratios in each plot in each month are listed in Table III. In plot C, ETa was higher than ETb in June 1989 and May 1990, though slightly lower than ETp in March 1990. In plot L , ETa never exceeded ETp. In June 1989, ETa/ETp was significantly higher ($p < 0.01$) for plot L than for plot C, during the other two pwriods, ETa/ETp was lower in plot L than in plot C, but the differences were not statisticallly significant at the 5% level.

Apparent interception in each plot in each year of study is listed in Table IV.Apparent interception was low in both burnt plots in the first and second years after burning, but by the third year was similar to that recorded

Table 3. Mean ratios of actual to potential evapotranspiration from plots C (control) and L (light burn) for the three periods during which evapotranspiration was estimated. The number of rain-free subperiods from which each mean value is calculated is also shown.

Period	ETa/ETp		Number
	Plot C	Plot L	of subperiods
June 89	0.88	1.28	6
Marzo 90	0.89	0.91	3
May 90	1.00	1.13	7

in the control plot. The between-year differences in interception in the control plot can be attributed to between-year variations in rainfall distribution (Rutter and Morton, 1977). Cumulative gross rainfall and througgfall curves for each plot and each year are shown in Fig. 2, and again illustrate how interception in the burnt plots approached near-normal values within three years of burning.

Surface runoff over the four years of study were divided into 100 events (Fig. 3). Runoff events during which the soil displayed severe or very severe water repellency (estimated on the basis of soil moisture content as shown in Fig. 1) are marked '1' in Fig. 3. These events (occurring mostly in the period May - October) accounted for 27.9%, 40.4% and 32.9% of total runoff over the four years from plots C, L and M respectively. The corresponding rainfall events, however, accounted for only about 15% of total rainfall over the four-year period . Runoff events during which soil water content was at saturation are marked '2' in Fig. 3. As is apparent from the figure, the majority of major rainfall events are attributable either to high water repellency or to soil water content being at saturation. During the first year post-burn, high water repellency appears to have been a more frequent cause of high runoff.

Runoff percentages for the burns plots showed a tendency to decrease over the years subsequent to the fire, reflecting the gradual establishment of vegetation cover

Table 4. Percentage apparent interception (see text) of rainfall in each plot in each year of study.

Year	Total rainfall	Interception (%)		
		C plot	L plot	M plot
1988/89	1071.4	43.7	13.8	10.7
1989/90	1636.4	42.3	24.3	23.8
1990/91	1597.8	38.4	31.9	36.3
1991/92	1100.2	51.0	40.0	48.6

Table 5. Runoff (expressed as a percentage of gross rainfall) from each plot in each year of study.

Year	Control Plot	L Plot	M Plot
1988/89	3.5	6.0	5.2
1989/90	2.8	4.0	2.9
1990/91	1.5	2.8	1.8
1991/92	1.9	2.4	2.3

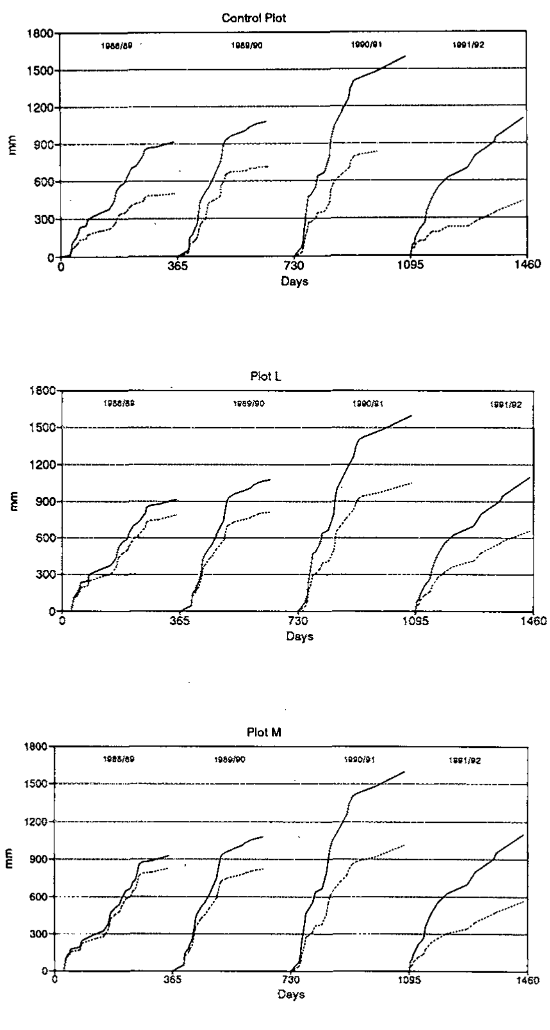

Figure 2. Cumulative gross (solid line) and net (dotted line) rainfall curves for each plot and each year of study.

(Table V). Note, however, that runoff percentages for the control plot also show a decreasing trend (which is, however, less marked than of the burnt plots): this can be attributed to between-year variability in rainfall distribution and thus in the factors affecting interception.

CONCLUSIONS

Various authors have reported increases in the water repellency of the surface layers of the soil following fire (see for example DeBano et al., 1981). The prescribed burns carried out in this study did not cause increased water repellency at any depth, which is probably attributable to the low maximum temperatures attained. In fact, burning caused water repellency in the 0 - 2.5 cm layer to drop slightly, possibly due to volatilization of the apolar compounds responsible for water repellency.

In the experiments reported here, burning led to an approximately twofold increase in runoff. However, the absolute increase was small (about 2 - 3% of rainfall), suggesting that wildfire-induced runoff is unlikely to have much influence on the flow regimes of Galician rivers, since quickflow (generally accounting for about 20 - 25% of total flow in this region) largely takes a subsurface route. The effects of wildfires on the water repellency of the soil, leading to increased surface runoff, can thus be expected to have little impact on river flow regimes; this contrasts with previous studies (such as Scott and Van Wyk, 1990), which have found that wildfires, by increasing the water repellency of the soil and thus increasing runoff, have a major impact on the flow regime of receiving rivers.

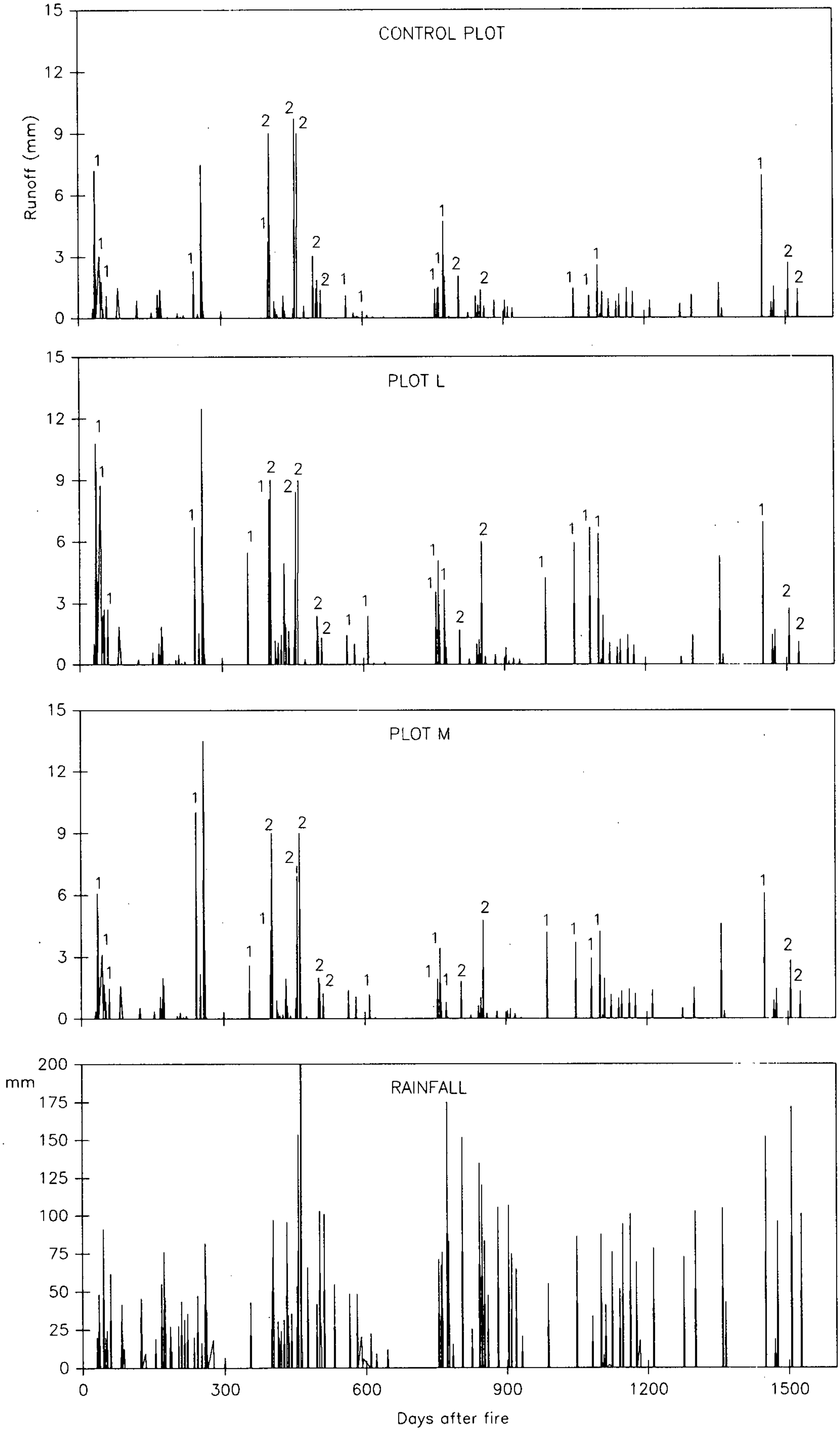

Figure 3. Temporal distribution of runoff from each plot over the study period. Runoff events marked '1' occurred during periods when water repellency of the soil was severe or very severe (see text). Runoff events marked '2' occurred during periods when soil water content was at saturation.

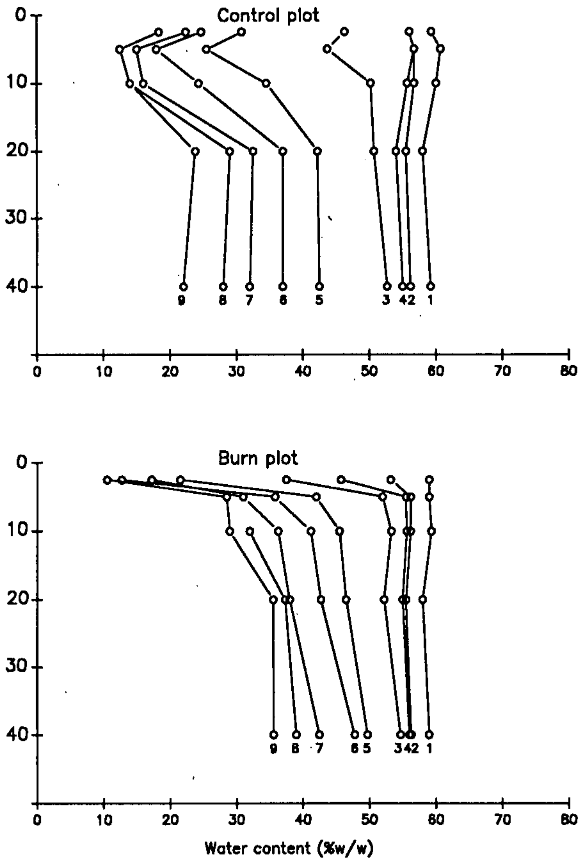

Figure 4. Soil water content profiles for the control plot C and the burn plot L, 9 months post-burn (june 1989). Soil water content was determined on the basis of tensiometers readings.

The same cannot be said of the effects of wildfire on evaporation from the vegetation canopy (i.e. interception) and on evapotranspiration of the soil water reserve. In the first year of the experiments reported here, loss of water to the atmosphere as a result of interception was 10.7 - 13.8% of rainfall in the burnt plots (as opposed to 43.7% in the unburnt plot). Similarly, burning led to reductions of about 30% in evapotranspiration losses over the first year. In view of the relatively minor increases in runoff from the burnt plots, these data suggest major increases in infiltration; this in turn suggests that wildfires may have a major impact on subsurface and groundwater flow, and thus on the flow regimes of Galician rivers. Responses to wildfire of this type may lead to increases in annual contribution of the order (about 25 - 30%) reported by Lavabre et al. (1993) for the south of France.

In cases in which the hydrological effects of wildfire are largely due to reductions in vegetation cover, the time required for recovery of prefire hydrological conditions can be expected to depend on the time required for recovery of the vegetation. In a study carried out in a region of Mediterranean climate and vegetation, Helvey (1980) found that prefire conditions had still not been attained seven years after the fire. The Ulex scrub studied by us, however, regenerates more rapidly (Casal, 1987), and we consider it likely that prefire conditions are reattained within about five years.

In Galicia, over the period 1985-1993, approximately 500.000 has were affected by wildfire. The degree to which the conclusions of the present study are valid for the region in general is currently unclear. Certainly, most fires lead to reductions in vegetation cover of similar magnitude to that observed in the present study, suggesting that similar effects on interception and evapotranspiration may be expected. On the other hand, fire characteristics, together with topografic , geological and climatic conditions , may all show considerable variation; furthermore, the eventual effect on watershed hydrology will depend

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Remote sensing techniques in the mapping of vegetation and their application to runoff evolution in burnt areas

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ABSTRACT

South of experimental forested catchments named "Réal Collobrier" was destroyed by fire in August 1990. Partially or entirely burnt catchments became an interesting field to study links between vegetation and runoff. An experimental program was built on two axis :

- use of remote sensing techniques to map limits of burnt areas and recovery of vegetation, year after year.
- link between vegetation and runoff. The first investigations were done on two burnt catchments (Rimbaud and Meffrey), using a runoff model fitted with data before fire. This model has permitted to compare new data (observed after fire) with the ones we would have observed without fire, for the months just after fire (1990, direct effect of fire) and the year after (1991, effect of vegetation recovery).

Keywords : hydrology, runoff, remote sensing, forest fires.

1. INTRODUCTION

Hydrological influence of vegetation on runoff for small catchments is relatively little-known. It's one of the most important preoccupation of the Scientific Group Real Collobrier which tries to understand hydrological behavior of small Mediterranean catchments. The forest fire of august 1990 was a scientific opportunity to experiment these catchments in order to better understand the vegetation effects on runoff using remote sensing.

This project has two major objectives interesting the management of Mediterranean area : 1) possibility of characterizing the vegetation recovery after fire by remote sensing methods and 2) hydrological study of vegetation effects on runoff. Moreover, this technology offers synoptic and frequent views, which can be very useful for further regionalisations of results.

Here we present and discuss the firsts results obtained for two years of observation after fire.

2. GENERAL METHODOLOGY

The different stages of the study are :

- installation of experimental plots in the burnt area, in order to observe year after year, the evolution of vegetation;
- cartography of fire limits and annual cartography of vegetation after fire, using plots observations and remote sensing imagery ;
- link between these cartographies and runoff evolution.

2.1 Study area

These experiments are undertaken on the BVRE (Experimental and research basins) of Real Collobrier, in Maures Mountains (France), where, since 1966, the CEMAGREF has collected data through an important hydrological network (LAVABRE et al., 1991a). The area is a metamorphic one, and forest covered, essentially by pines, cork-oaks and chestnut-trees.

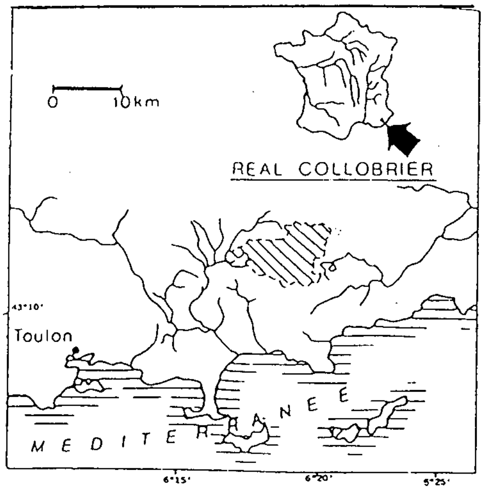

Figure 1. Real Collobrier basin

The annual rainfall is quite favorable (mean of 1000 mm) due to the mountainous character of the basin (70 to 800 m) and its proximity to Mediterranean sea (Lavabre, 1980). Since 1988 and 1989 were two years affected by drought, forest fires were larger in 1990. At the end of August 1990, fire spread over 8000 ha during 3 days and affected the south part of our experimental basins (LAVABRE et al. 1991b).

2.2 Hydrological data

The 100 km≤ of the Real Collobrier basins are covered with 17 rainjauges and a climatologic station. 11 flowjauges determine 11 small catchments from 0.7 to 70 km≤.

Table 1. Some characteristics of Rèal Collobrier catchments

N°	Basins	Area (km^2)	begin of data
1	Pont de Fer	70,6	1966
2	Collobrière	29,5	1972
4	La Malière	12,3	1965
5	Valescure	9,4	1967
6	Maurets	8,4	1968
7	Vaubarnier	1,5	1968
8	Rimbaud	1,4	1967
9	Davids	9,7	1966
10	Cogolin	5,5	1969
18	Boussicaut	0,7	1980
19	Meffrey	1,5	1985

2.3 Remote sensing data

We used multispectral SPOT data for this operation, due to their precise resolution (20*20m) and the possibility of satellite programming : one image just before fire, one just after, and then, one image each year at the same period

Table 2. SPOT images - XS KJ 52/263, 1B

Date incidence	Angle
21/08/90	17,5°(left)
04/09/90	10,1°(left)
26/07/91	8,8°(left)
13/07/92	8,3°(left)
16/06/93°	8,3°(left)

2.4 Geographical data

From previous works we had :

- two DEM (Digital Elevation Models) : a first one at 20 m of resolution, derived from IGN map at 1:25000 and covering only the 11 catchments ; a second one at 30 m of resolution, derived from IGN map at 1:50000 and covering the whole Maures Mountains;

- a cartography of forest species obtained by 1986 and 1989 SPOT images processing (WEESAKUL, 1992).

2.5 Experimental observations

In order to perform a significant image processing, it is important to have precise and representative land observations. So we have set up some observations plots with great care.

A first set of observation plots was proposed in 1991 and completed in 1992. The plot itself is a square of 20*20m corresponding to usual forested observations areas (BRETON, 1992) and also to the SPOT pixel. The criteria of selection are :

homogeneity : the plot must be chosen in an homogeneous area, minimum 60*60m, in order that an imprecise position does not affect the information. Indeed, in image processing we assume that localization is imprecise at more or less one pixel (20 m).

statistical representativity : plots of observation compose a sample as complete as possible for combinations of forest species, slopes, and aspects

precise localization : defined through 1:25000 IGN map and by using a GPS (Global Positioning System) ; easy access

For each plot the observations are :
- topographical characteristics (high, slope and aspect on a 20*20 m square) ;
- geographical position ;
- percentage of bare soils or rocks ;
- description of vegetation : names, height, number or percentage of cover, and an index of burnt violence (0 to 5) assessed just after the fire ;
- local map of situation ;
- general information : forestworks, erosion .. ;
- snapshot.

For a complete description of vegetation, special floristic observations were done by St. Jerome University in Marseille (THINON et BRUGER, 1993), in summer 1993.

3. MAPPING BURNT AREAS AND VEGETATION RECOVERY

3.1 Preprocessing of satellite images

Several images of different dates must be combined to obtain vegetation recovery cartographies. Thus, the first step is to transform images into compatible forms by using geometrical and radiometrical corrections.

Geometrical corrections

The images are rectified using a common reference tool : the IGN map at 1:25000. Correction appears necessary in mountainous areas if the incidence angle is significantly different from zero. This correction has been done using a software which determines and eliminates the parallax effect using a DEM and a set of reference points - common points on map and image - (PROY, 1986 ; PUECH, 1993).

Radiometrical corrections

Two radiometrical rectifications have to be done.

- The first one concerns the relief effects on reflectance. These effects appear in mountainous areas when the solar elevation angle is low, and give sides lightened differently.

The elimination of this effect needs an internal correction of the images. The theoretical models of correction (YANG, 1990) were tested on burnt areas for the 09.1990 image but gave poor results (LEPORI, 1991). The reflectance in these burnt areas depends not only on slopes and solar angle, but also on intern shadows due to the structure of the cover, and these effects are difficult to integrate in theoretical approaches.

So we used an empirical correction, based on statistical links between CN (Numeric Counts) and AIS (solar incidence angle), on the burnt parts of the images, the correction being redefined for each image and each radiometric band (NGUETORA, 1993).

- the second is the atmospherical correction which is necessary, when two or more images are used.

Here too, the theoretical approaches are often not available due to the lack of information on atmospherical parameters. So, empirical approaches have been developed ; they are based on the assumption of a linear perturbation of atmosphere on radiometry, and on the existence of peculiar areas with radiometric stability all over the year, "invariant zones", such as bare soils or roads for instance (ABEDNEGO, 1989 ; SEGUIS et al. 1992). The comparison, band by band, of the CN (Numeric Counts) on these invariant zones permits a relative correction, by taking one image as the reference.

3.2 Cartographies

This first step give the external limits of fire, statistics on burnt areas in each catchment, and a discrimination of the different burnt areas on a criteria of violence of fire. It has been done using the 24/08/90 and 4/09/90 images and ground observations of July 1991.

Limits of fire

We first tried to subtract the two images of 1990, one just before fire, the other just after.

For the pixels inside the burnt area, the results are correct, but outside this zone, the problems of exact superposition of the two supports gave poor results (BRETON, 1992).

So we only used the image after fire, studying the histogram on the PIR band (Near infrared). Burnt areas appear to have the lower CN, and the discrimination is quite good.

Results indicate 19% of burnt areas inside the whole experimental zone, some catchments being intact (basins 2, 5, 6 and 7), some a few burnt (basins 1, 4, 9 and 10) and some quite integrally burnt (basins 8, 19) (Fig 3 et Tab 3).

Table 3. Burnt areas on each catchment

N°	Basins	Burnt Area	
		(ha)	(%)
1	Pont de Fer	1294	18 %
2	Collobriére	0	-
4	La Maliére	458	37 %
5	Valescure	0	-
6	Maurets	0	-
7	Vaubarnier	0	-
8	Rimbaud	127	82 %
9	Davids	309	32 %
10	Cogolin	130	20 %
18	Boussicaut	0	-
19	Meffrey	141	89%
	Total	1603	19%

Statistics on burnt vegetation

Each forest specie has a different behavior (Tab 4). For example, pines are well burnt, but chestnut trees are not. Three years after fire, ground observations on burnt areas indicate that, in the burnt area, pines have completely disappeared and that it remains only three land components in variable proportions : cork-oaks, shrubs and grass, bare soils and rocks.

Table 4. Burnt areas by soil occupation

Soil occupation	burnt Areas Ha	% burnt(*)	% burnt(**)
Chestnut trees	39	5 %	2 %
Chestnut+ oaks	83	10 %	5 %
Oaks	514	18 %	32 %
Oaks+Pines	334	43 %	21 %
Pines	377	49 %	23 %
Pines + chestnut	7	34 %	0 %
Tracks+bushes	207	5 %	13 %
Vineyard	32	0 %	2 %
Urban areas	0	22 ,%	0 %
Burnt area 1986	10	3 %	1 %
Total	1603		100 %

(*) in reference of the same soil occupation
(**) in reference of the whole study area

Cartography of recovery

This cartography has been done for each image after fire by using data collected each year on experimental plots.

Due to the fact that, after fire, the pixels of burnt area are no more homogeneous, we used a sub pixel processing. The process define each pixel as a mixing of pure entities : bare soil, herbaceous, shrubs and trees.

Results are cartographies of recovery year by year : percentage and localization of each entity. Statistics by catchment can be obtained. For example, remote sensing processing give for 1993 image, a mean of 43 and 38 % of bare soils, respectively for the Rimbaud and Meffrey burnt catchments This is consistent with ground observations on a few plots (tab 5).

Table 5. Mean percentage of bare soil and herbaceous (ground observations)

Catchment	1990	1991	1992	1993
Rimbaud	90	80	65	50
Meffrey	90	70	50	40

The differences between catchments are of great importance to understand the different hydrological behaviors after fire.

4. EVOLUTION OF RUNOFF AFTER FIRE

To characterize evolution of hydrology before and after fire, the direct study of chronicles is not good enough, because climatologic perturbations have a great importance on runoff.

We study these evolutions for events and at monthly and annual time steps.

The method of comparison uses two ways :

- a direct comparison between runoffs on reference basins and runoffs on burnt ones ;
- use of an hydrologic model fitted on years before fire and applicated to years after fire, in order to study the differences between calculated and observed runoff values. This has been done on burnt and not burnt basins all together.

The two burnt basins studied are "Rimbaud" and "Meffrey" and the reference (not burnt basin) is "Valescure". The characteristics of these 3 basins are shown in tab 6.

Table 6. Characteristics of the 3 basins (in WEESA-KUL, 1992 and BRETON, 1992)

	Rimbaud	Meffrey	Valescure
Surface (km²)	1.5	1.5	9.4
Mean altitude	550 m	274 m	466 m
Mean slope	9°	17°	21°
Mean aspect	235°	203 °	195°
Bed rock	Gneiss°	Phyllade	Micaschiste
Vegetation	Bushes + Chestnut	Pines + Oaks	Oaks+ Chestnut

The reference period before fire is 1967-1990, the hydrologic year is considered from august year i to July year i+1. For more information on reconstitution of rains and runoffs see BRETON (1992).

Table 7. hydrologic characteristics of catchments

Basins	Rimbaud	Meffrey	Valescure
begin of observations	**1967**	**1985**	**1967**
PA (1967-1989)	1186		1200
PA (1985-1989)	901	655	913
LA (1967-1989)	664		523
LA (1985-1989)	474	105	332
PA 1990	983	948	1171
LA 1990	645	198	312
PA 1991	755	692	859
LA 1991	359	29	60

PA: mean annual rainfall (mm)
LA: mean i.e. LA = volume of flow/surface annual high of flow (mm)

4.1 Events

Fire increased considerably the runoffs in the few months after fire

- On "Rimbaud" basin a first study has been done by Lavabre et al. (1991b) giving some information on its behavior during the months after fire. Lavabre and al (1991b) indicate that, for the 3 months after fire, runoff overpassed 3 times the decennal value
- On Meffrey catchment, for the same period, runoff overpassed 6 times the maximum observed since 1987,

These exceptional runoffs occured with normal rain events.

4.2 Monthly runoff

We used a mathematical model of runoff : the GR3M model (MICHEL, 1991), a conceptual model using 3 parameters and a monthly period. It was fitted on data before fire (Tab 3), then used on months after fire.

At this period of time, comparison between calculated and observed values gives the following results :

- for the Rimbaud catchment (Fig 4) a great increase in runoff is observed just after fire (09.1990) and confirms the global increase for the whole year 1990; *the observed values remain greater than calculated values for more than one year, but, after December 1991 they become lower* ;

- for the Meffrey catchment (Fig 5) an increase is also observed just after fire, but after April 1991 (i.e. 6 months after fire), *the observed values become lower than calculated ones.*

The difference in behavior of these two catchments is interesting for many reasons :

- first, it reveals a great increase in runoff just after fire; this was expected and has been yet observed on daily data

- secondly, it reveals a rapid evolution of runoff conditions on these burnt areas. After a few months the increase disappears and the runoffs are lower than before fire. Furthermore, the duration of the period of increase appears different from one catchment to the other: it seems difficult to describe catchments after fire by a unique state and we must consider transitory states, rapidly variable.

Exact reasons for these differential evolutions are difficult to establish perfectly with so few observations, but it seems logical to attribute the evolution of runoff conditions to vegetal growth, which can restrain runoffs during the first spring after fire.

4.3 Annual runoff

At annual time step, no precise evolution is observed.

- the two burnt catchments present opposite results : small increase for Rimbaud catchment (fig 6 : 20% in 1990 and a few percents in 1991), and decrease for Meffrey ;

- not burnt catchments present a decrease in 1990 and 1991

At annual time step (fig 7) two effects are combined : a decrease due to dry years just before fire, and an increase due to fire conditions.

Moreover, a cover without no vegetation at all (giving great runoff) lasts just a few months. After one month many plants have begun their growing, and at the first spring the herbaceous strata is sufficiently important to influence hydrological behavior. It is not possible to consider catchments covered with bare soils for the whole year after fire.

So we conclude that an annual interval is too long to define a clear evolution.

The differential evolutions observed at annual time step explain that *annual values cannot be simply related with percentage of burnt areas.*

5. CONCLUSIONS AND PERSPECTIVES

These results already allow us to tell that a precise maps of soils cover can help runoff studies. For this, remote sensing techniques appear as an appropriate tool. Here, we only used information about total surface of burnt areas by basins, but we have planned to define finer maps of vegetal recovery, year by year.

It appears a great increase of runoff just after fire. This increase is observed during several months, but its duration depends on the catchments : it can be greater (Rimbaud) or smaller (Meffrey) than a year. After this period, we observe a decrease of runoff, with values possibly lower than the ones before fire.

These strong differences in runoff evolution from one basin to the other, during the months after fire, explain that at annual time step, no clear evolution can be observed, due to changes quicker than a year.

It is then important to notice the rapidity of evolution :

- vegetation recovery concerns herbaceous for the first year ; In the years after, progressively bushes take the place of herbaceous strata ; trees grow much more slower and their evolution is not sensitive in the first 3 or 4 years after fire.

- hydrological response increases during maximum 2 years, depending on the catchment ;

- erosion increase during the first months and is completely stopped after one year.

These observations shows the importance of herbaceous strata on runoff conditions; Two others conclusions can be defined :

- at annual rate (or more) it appears impossible to define hydrological conditions of basins after fire by a unique state

- it would be dangerous to conclude on evolution of runoff conditions by analyzing only one burnt basin.

ACKNOWLEDGMENT

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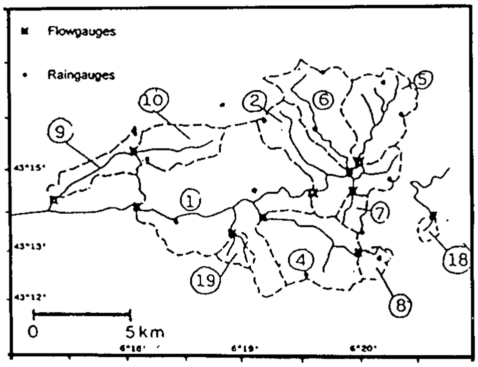

Figure 2. Réal Collobrier Catchments

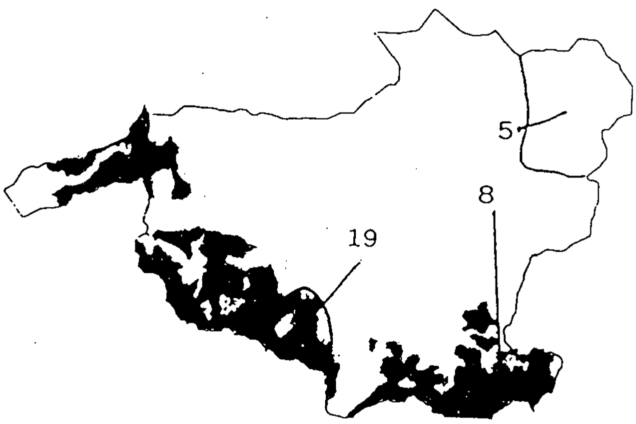

Figure 3. Cartography of burnt areas using histogram of the 09.1990 image.

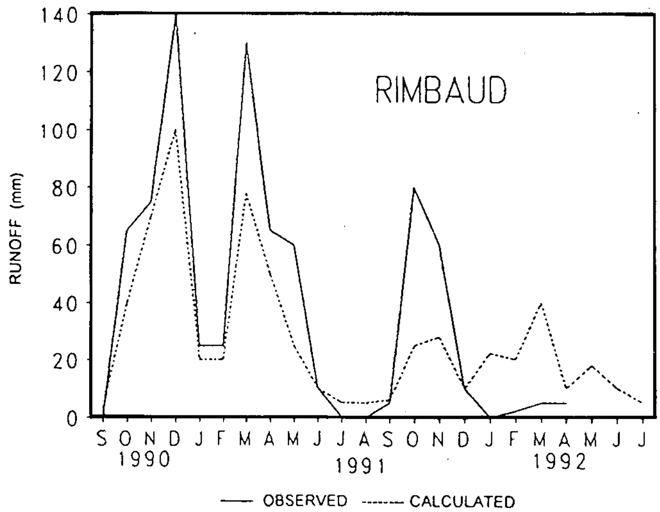

Figure 4. Monthly runoffs calculated and observed : Rimbaud catchment

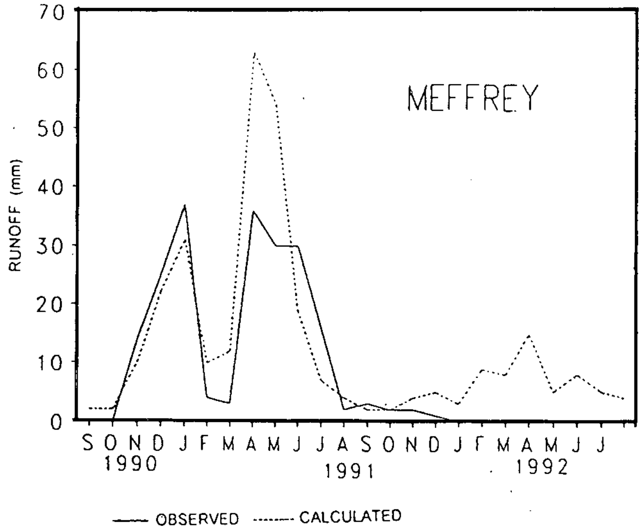

Figure 5. Monthly runoffs calculated and observed : Meffrey catchment

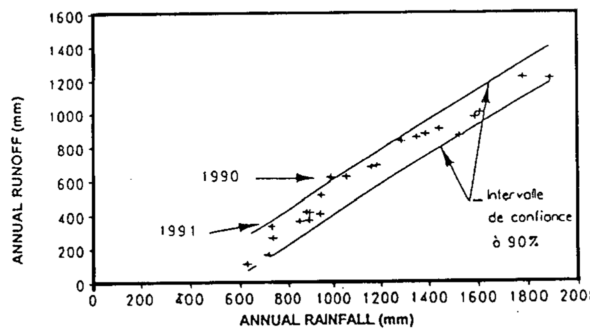

Figure 6. Linear link annual rainfall / runoff. Rimbaud catchment

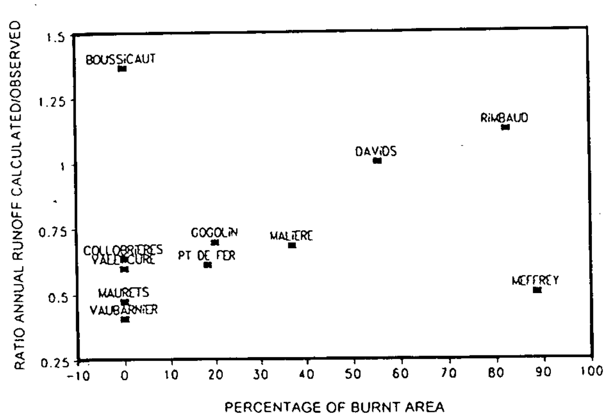

Figure 7. Annual runoffs calculated and observed in year 1990

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Six years of study on fast growing forest plantations catchments in the Northwest of Spain

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ABSTRACT

Data on water balance in three fast growing forest plantations experimental catchments in the northwest of Spain are presented. Two watersheds are covered by Eucalyptus globulus and other is covered by Pinus pinaster.

During the six years of study several perturbations occurred. In 1989 two consecutives wildfires affected to one E. globulus watershed. The second fire also burned the other eucalyptus watershed. All eucalyptus were felled since 1991 to 1992. Also 25% of the watershed area in the pine catchment was cutted in 1991. Quick changes in the hydrologycal regimen took place after these perturbations.

The very fast recovering capacity after fire and cutting of Eucalyptus globulus facilitated the return to pre-fire hydrologic parameter values within only a few years.

Before perturbations occurred, nutrient balances were very conservative, and were similar among the cathments. This indicates a good performance on the part of both tree plantations. Water comsumption for pine and eucalyptus stands was very similar.

INTRODUCTION

The eucalyptus plantations in Galicia were initiated in the forties, expanding very quickly, due to the excellent yield of this fast growing species, especially in the coastal area. Another factor in this expansion has been the occurrence of forest fires which, from the seventies until now, have dramatically affected Galician forests. The pyrophytic character of *Eucalyptus globulus* has given rise to an explosive expansion of this tree into an area initially covered by shrubs and *Pinus pinaster* stands. Consequently, at this time -with only 40,000 ha. of pure *Eucalyptus globulus* plantations- the total area that has been invaded by this myrtacea species is nearly 200,000 ha., in which the *E. globulus* has mixed mainly with *P. pinaster*. This significant expansion from its original planted area has caused an increase in concern about the assumed detrimental ecological effects of this species, especially with regard to the excessive water consumption in plantations of these species.

Some factors that are important to consider are the enormous variability that the mountainous terrain in Galician forests has, and the very short rotation that is applied to the *E. globulus* plantations in the northwest of the Iberian Peninsula (usually, 10 to 15 years). Both of these factors complicate the evaluation of the ecological effects of these species,and at the same time support the choice of watershed experimentation that allows integration of most of variability previously mentioned.
When this approach was initiated, information existed on the effects of eucalyptus forest plantations on soil in Galicia (Diaz Fierros, et al.(1982); Rodriguez Fernandez (1984); and Bará et al.(1985). However, very little knowledge existed on the hydrological effects of eucalyptus. Poore and Fries (1985), Florence (1986), Lima (1987), and Adlard (1987) had compiled most of the hydrological information at a world-wide level. The ex-

periments of Van Lill et al. (1980) on *E. grandis* plantations in South Africa, Soares David, et al. (1986) on *E. globulus* plantations in Portugal, and the work of the C.S.W.C.R.I. of Udhagamodalam (1987) on plantations of this same species in India constituted the data available to us at the time on eucalyptus plantations in experimental watersheds. Only a few studies on some aspects of water relations of *E. globulus* had been carried out on experimental plots in Galicia (Calvo et al., 1979; Paz Gonzalez, 1982).

Consequently, in the mid-1980s, the C.I.F. de Lourizan made the decision to establish a series of experimental catchment basins in eucalyptus forest plantations to study water balance and water consumption. It was deemed necessary to compare the data from the eucalyptus experiments with information obtained for other species; similar installations were put in place for *Pinus pinaster. Pinus pinaster* was chosen as the most traditional and most important commercial forest tree, with a habitat similar to that of *E. globulus.* Both species are currently the principle source of raw material for forest industries in Galicia.

These fast growing tree plantations have a very short rotation period, and are subject to frequent fires. They induce some rapid changes in the hydrological regime, and watersheds covered with these species provide good examples of temporal hydrological variability; in contrast to catchments for other, more slow-growing species that have a steadier hydrologic regime. This paper discusses the installation of three experimental watersheds covered with *E. globulus* and *P. pinaster.* It will emphasize the annual variability of several hydrologic parameters affected by a wild fire, partial cutting, and a clearcut, in a short period of six years.

Methodology and description of the experimental watersheds

The method employed in this work is the experimental watershed water balance method (Toebes and Ourivaev, 1970). This well known method has been in use for many years (Bormann and Likens, 1967, 1970, 1979; Likens and Bormann, 1972; Bosch and Hewlett, 1982),etc. and more recently in Spain (Escarré, Gracia, Rodá and Terradas, 1984).

Three experimental watersheds were installed:

In 1987, the Castrove Eucalyptus globulus watershed

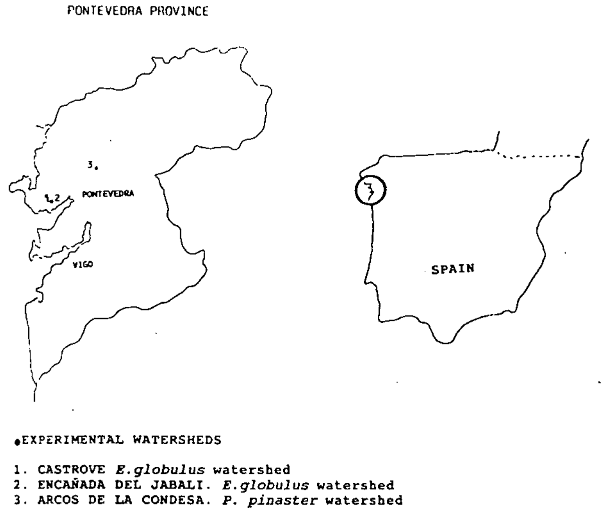


Figure 1. Experimental watersheds in the Pontevedra province: 1) castrove E. Globulus watersheds, 2) Encañada del jabalí. E. Globulus watershed, 3) Arcos de la condesa, P. Pinaster watershed

was located on the south slope of the Castrove mountain ridge, near the shore of the Ria of Pontevedra.(fig. 1)

The catchment is totally occupied by a *Eucalyptus globulus* plantation that was 15 years old in 1987 (this age is the most common for the final cut for this species in Galicia). Under the eucalyptus plantation *Acacia melanoxylon* grows (mainly in the most humid parts of the catchment), as does *Ulex europaeus* and *Pteridium aquilinum.*

We have measured a strong pluviometric gradient with altitude in this area, comparating it with the nearest pluviometric stations; for a long period of years, mean annual precipitation in the watershed probably was more than 2100 mm.Other data are in tables 1 and 2.

In the same year,1987, the Arcos de la Condesa Pinus pinaster watershed was installed.This basin is a small valley on the east face of Monte Axeiros (fig.1). All the catchment is covered by a *Pinus pinaster* plantation that was 30 years old at the beginning of the experiment. Also present are young plants of *Quercus robur* (very frequent in the part of the catchment where soils are deep and fresh), *Pteridium aquilinum* (very thicket and high in spring and summer), *Ulex europaeus* and *Rubus sp.* in the understory. Grass (*Agrostis curtisii* and *Arrenatherum sp.*) abounds in the watershed.

Complementary data are summarized in tables 1 and 2.

In 1989,other experimental watershed,Encañada del Jabalí Eucalyptus globulus watershed was established in

Table 1. Catchments location and main physiographic characteristics.

Name	Situation Lat N	Lon W	Altitude (m)	Mean Slope(%)	Exposition	Area (ha)
CASTROVE	42º26'40" - 27'00"	8º43'30" - 55"	348-445	22	S-SE	9.9
ARCOS	42º34'28" - 48"	8º36'57' - 37'8"	210 -251	10	S-SE	6.74
ENCAÑADA	42º25'49" - 27'9"	8º43'4" - 44'4"	150-475	13	S	194

Table 2. Catchments climatic,edaphic,and vegetation characteristics

Name	Parent Material	Soil Texture	Ta (ºC)	Tc (ºC)	Th (ºC)	Climate	Annual Mean P (mm)
CASTROVE	granitic	sandy	14.1	8.8	20.2	maritime warm-humid	1940
ARCOS	granitic	loam	14.2	9.1	20.6	"	1540
ENCAÑADA	granitic	sandy	14.1	8.8	20.2	"	1940

Ta.- annual mean temperature. Tc.- mean temperature in the coldest month Th.- mean temperature in the hottest month. P .- precipitation.

the big Eucalyptus plantation area of Monte Castrove ridge.(fig.1)

Almost all the catchment area was covered by 15-30 year old *Eucalyptus globulus* plantations. This watershed contained in its interior the smaller Castrove watershed already described.
Other characteristics are in tables 1 and 2.

Equipment in the watersheds

The three catchments have stream gauging stations with standard ink scripture limnigraphs (OTT Kempten). In Castrove and Arcos de la Condesa watersheds, v-notch weirs are installed, and Encañada del Jabalí has a rectangular weir. There are automatic water sampler pumps in the three stream gauging stations (HCV Struers). In Encañada del Jabalí there is also a proportional water sampler (Aquapropor PRT 41 Struers).

Two meteorological stations were installed, one in the interior of the Encañada del Jabalí *Eucalyptus globulus* watershed, and the other in the *Pinus pinaster* watershed. In both there are pluviometers, a pluviograph, a windmeter, a hygrothermograph, maximum and minimum thermometers, and a Piché evaporimeter. Devices for taking rain water samples are installed in the eucalyptus and pine watersheds. Three plots per catchment were used to make water content determinations (gravimetric method). Data are taken weekly.

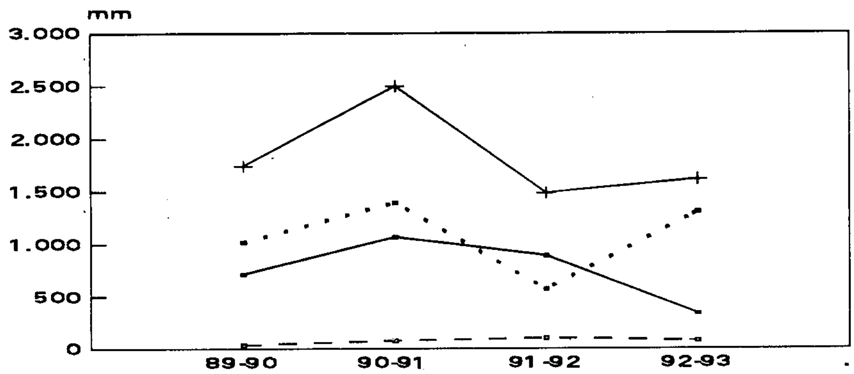
Standart pluviometers under trees (twelve per catchment) and stemflow rings (ten per catchment) were instaled to measure rain interception in Castrove watershed and in Arcos de la Condesa watershed. Data were taken weekly or more frecuently in some periods of very heavy rain.

Changes in the watersheds during the study period

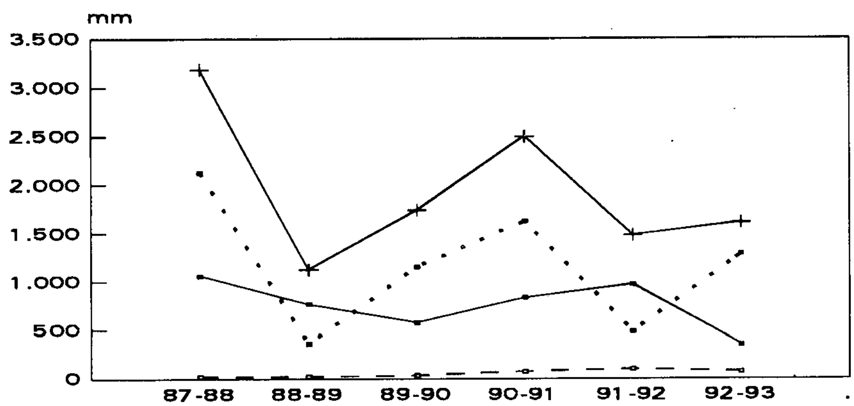
During the first hydrological year studied (87-88) no perturbation or change occurred in the two first installed watersheds. Both forest plantations were growing in a natural way, without human intervention or perturbation.

During the very dry 88-89 hydrological year, the Castrove eucalyptus catchment was affected by two consecutive forest fires. The first one happened at the beginning of spring, with a very high soil moisture level. It

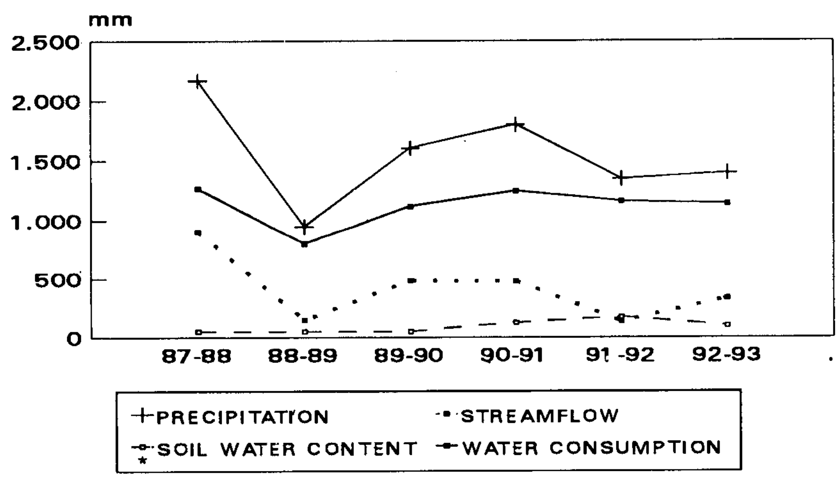
ENCAÑADA EUCALYPTUS WATERSHED WATER BALANCE



CASTROVE EUCALYPTUS WATERSHED WATER BALANCE



ARCOS P. PINASTER WATERSHED WATER BALANCE



* at the end of the summer.

Figure 2

Table 3. Nutrient balance in Castrove eucalyptus watershed during a non alteration period: Hydrologic year 87-88. (Values in Kg. per ha. and year)

	Ca	Mg	K	Na	PO4-3	NO-3
INPUT	9,6	12,7	6,4	132,5	0,22	0,0
OUTPUT	1,5	8,2	1,0	124,7	0,20	0,0
BALANCE	+8,1	+4,5	+5,4	+7,8	+0,02	0,0

Table 4. Mean nutrient concentrations in Castrove eucalyptus catchment water streamflow during hydrologic year 87-88. (Values in parts per million.)

	Ca	Mg	K	Na	PO4-3	NO-3
p.p.m	0,07	0,39	0,05	5,9	0,009	0,0

Table 5. Nutrient balance in Arcos de la Condesa Pinus pinaster watershed during a non alteration period: hydrologic year 87-88. (Values in Kg per ha. and year)

	Ca	Mg	K	Na	PO4-3	NO-3
INPUT	5,4	10,2	4,3	90,1	0,15	0,0
OUTPUT	4,0	4,7	1,2	69,5	0,05	0,0
BALANCE	+1,4	+5,5	+3,1	+20,6	+0,10	0,0

Table 6. Mean nutrient concentrations in Arcos de la Condesa Pinus pinaster watershed during a non alteration period: hydrologic year 87-88.

	Ca	Mg	K	Na	PO4-3	NO-3
p.p.m	0,44	0,52	0,13	7,7	0,005	0,0

Table 7. Maximum nutrient concentrations in stream water flow Castrove eucalyptus watershed after summer 1989 forest fire. (Values in parts per million)

	Ca	Mg	K	Na	PO4-3	NO-3
p.p.m	1,5	1,3	2,3	15,0	0,04	1,28

was a very low-intensity fire and only a small part of the litter was consumed; 20% of the total area of the catchment was burned. Understorey recovery was very fast.

At the end of July, after a long, dry and windy period with high temperatures, a tremendous forest fire took place on the south face of the Castrove range. All, of the Castrove watershed was affected and an important part of the Encañada del Jabalí watershed was also affected by the fire. Most of the litter and shrubs were consumed. Eucalyptus trees were injured in different ways, depending on their position in the catchments.

Extensive eucalyptus defoliation occurred. Some eucalyptus were badly damaged, and died in a short time, but the majority of trees resisted fire and began to regrow very early, slowly recovering the foliar surface destroyed by fire during the next few years. All of the eucalyptus trees in the Encañada del Jabalí watershed affected by fire were felled during the hydrological year 91-92, and the same occurred at the end of summer 1992 in the Castrove watershed. Trees on 25% of the total area in Arcos de la Condesa *Pinus pinaster* watershed were also felled at the end of Summer 1991.

In the eucalyptus watersheds, slash burning was applied after clearcutting.

Chemical water analysis

Rainfall and water streamflow have been sampled and analysed periodically (weekly, when possible) for nutrient content. Cations analysed are: Calcium, Magnesium, Potasium and Sodium; and the anions: PO_4^{-3} and NO_3-.

RESULTS

Water balances in experimental watersheds

These data appear in the fig.2. Looking at the results of the two eucalyptus catchments, we can point out the following aspects:The hydrologic year 87-88 was tremendously wet in this region, and no water deficit occurred throughout all the year. So, the obtained water consumption (1061 mm) is probably very close to the potential evapotranspiration for a representative eucalyptus plantation in the coastal area of Pontevedra. The next hydrologic year (88-89) was very dry, and so the water consumption was very low (766 mm), and far from the potential consumption.

After the fire in the summer of 1989, a strong reduction in water consumption occurred because an important transpirant foliar surface had been destroyed (Langford, 1976). But the incredible recovery capacity of eucalyptus makes it possible for water consumption to increase in the following years (Langford and O'Shaughnessy, 1977). So, in the hydrologic year 91-92, water consumption in Castrove watershed reached 972 mm, which is a considerable value, indicating a strong recuperation of the eucalyptus trees.

The Encañada del Jabalí eucalyptus catchment was less affected by fire than the Castrove watershed over the majority of its area and so in the first year after the fire (89-90), water consumption in Encañada was 715 mm; 139 mm more than in Castrove watershed (576 mm). But on other hand, on some steep slopes in Encañada, severely affected by fire, soil erosion after the first rains was considerable.

Finally, what is remarkable in both catchments is the large reduction in evapotranspiration after tree felling. So, in the first hydrologic year after cutting, 92-93 a moderately rainy year, water consumption for Castrove and Encañada were only 352 mm and 334 mm, respectively. The almost identical consumption for a very similar reduction of vegetation indicates the equality of hydrological behaviour between both watersheds.

In the Arcos de la Condesa *Pinus pinaster* watershed, the high water retention capacity of soils makes possible a high level of consumption. The canopy interception that was measured gave the values of 15% and 25% of rainfall for *E. globulus* and *P. pinaster* respectivelly. So for a moderate precipitation year (1750 mm aproximately), the *P. pinaster* plantation intercepted 175 mm more than the *E. globulus* plantation.

Both soil, and interception, contributed in an important way to the resulting total water consumption in the Arcos de la Condesa Watershed.

Nutrient balances and nutrient concentrations in water streamflow

During a period without any alteration, nor human intervention, in the *P. pinaster* and *E. globulus* catchments (hydrologic year 87-88), nutrient balances were very conservative, and were similar between the catchments. Mean nutrient concentrations in water streamflow during this period were low, and extremely low for PO_4^{-3} and NO_3-. All of this indicates a good performance on the part of both tree plantations.

In the opposite direction, after the 1989 forest fire in the eucalyptus catchments, significant changes in nutrient concentration in water streamflow occurred.

In tables 3, 4, 5 and 6, appear values on nutrient balances and mean nutrient concentrations in Castrove and Arcos de la Condesa watersheds. In table 7 appear values of maximun nutrient concentrations in stream water flow in Castrove eucalyptus watershed after summer 1989 forest fire.

CONCLUSIONS

Quick changes in the hydrological regimen take place just after an intense forest fire and a clearcut in eucalyptus plantations.

An important reduction in water consumption in the first post fire year occurred as a consequence of the drastic reduction of foliar surface in the eucalyptus stand.

The extraordinary recovering capacity of *Eucalyptus globulus* after fire facilitates the return to hydrologic parameter values that are very close to the ones of the prefire situation within only a few years.

Water potential consumption in a representative and undisturbed galician *Eucalyptus globulus* plantation was around 1000-1100 mm per year. This figure was similar to that of *Pinus pinaster* in another experimental watershed from the same area.

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Effects on rainfall gradient on tree water consumption and soil fertility on *Quercus pyrenaica* forests in the Sierra de Gata (Spain)

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ABSTRACT

Calculation of water consumption and nutrient fluxes were made for four Quercus pyrenaica forests along a rainfall gradient, located in the "Sierra de Gata" mountains (CW Spain), to obtain information about the evolution of fertility in long terms.

It is concluded that the water content at the beginning of the active growth period of the vegetation depends mainly of the soil characteristics; for that, there was a positive correlation between the annual evapotranspiration and the precipitation in the May-August period, but not with annual precipitation. So, the greater abundance of rainfall in the wet season did not tend substantially to increase water consumption by the vegetation. On the other hand, there was a high correlation between the volume of annual rainwater and that of drained water; the excess of water in the soil produced on winter gives as result a nutrient leaching of the soil and a consequent loss of fertility. This was confirmed by the net balance of several bioelements, the Ca/Al ratio and pH of the soil solution, and canopy leaching values.

Key words: water balance, nutrient balance, soil fertility, rainfall gradient, forest , ecosystem.

1. INTRODUCTION

Evapotranspiration is a relevant parameter in the understanding of terrestrial ecosystems, especially in Mediterranean areas, where water availability is scarce during dry periods (Piñol *et al.* 1991). The soil behaves as a buffered system which receives water intermittently and releases it continually by evapotranspiration (Garnier *et al.* 1986). Thus, in climates with a Mediterranean influence, a greater winter rainfall may positively affect soil moisture during the active period. Piñol *et al.* (1991) point out that annual evapotranspiration is positively correlated to annual rainfall in the Mediterranean area. This possible variation in water availability may cause differences in both photosynthetic efficiency and light interception (Jarvis & Leverenz 1983, quoted in Tenhunen *et al.* 1990), as a consequence of a lesser limitation in transpiration.

However, the differences in the volume of rainfall affect the evolution and properties of the soil profile (Birkeland 1984). Low saturation of bases and greater weathering of the original material is usually associated with humid regions (Quilchano 1993); this is due to the differences in the excess of water in the soil produced by the different pluviometry, giving rise to leaching processes of nutrients in the soil, with the resulting loss of fertility.

This double influence of rainfall on hydric and nutritional availabilities appears to have positive and negative effects, respectively, on forest productivity. The first results found in four *Quercus pyrenaica* forests, situated along a pluviometric gradient, indicate that neither productivity (Martín *et al.* 1993) nor the leaf area index (Gallego *et al.* 1994) respond to that gradient.

This study is part of a research project on the ecology of *Quercus pyrenaica* forests. Water fluxes have been con-

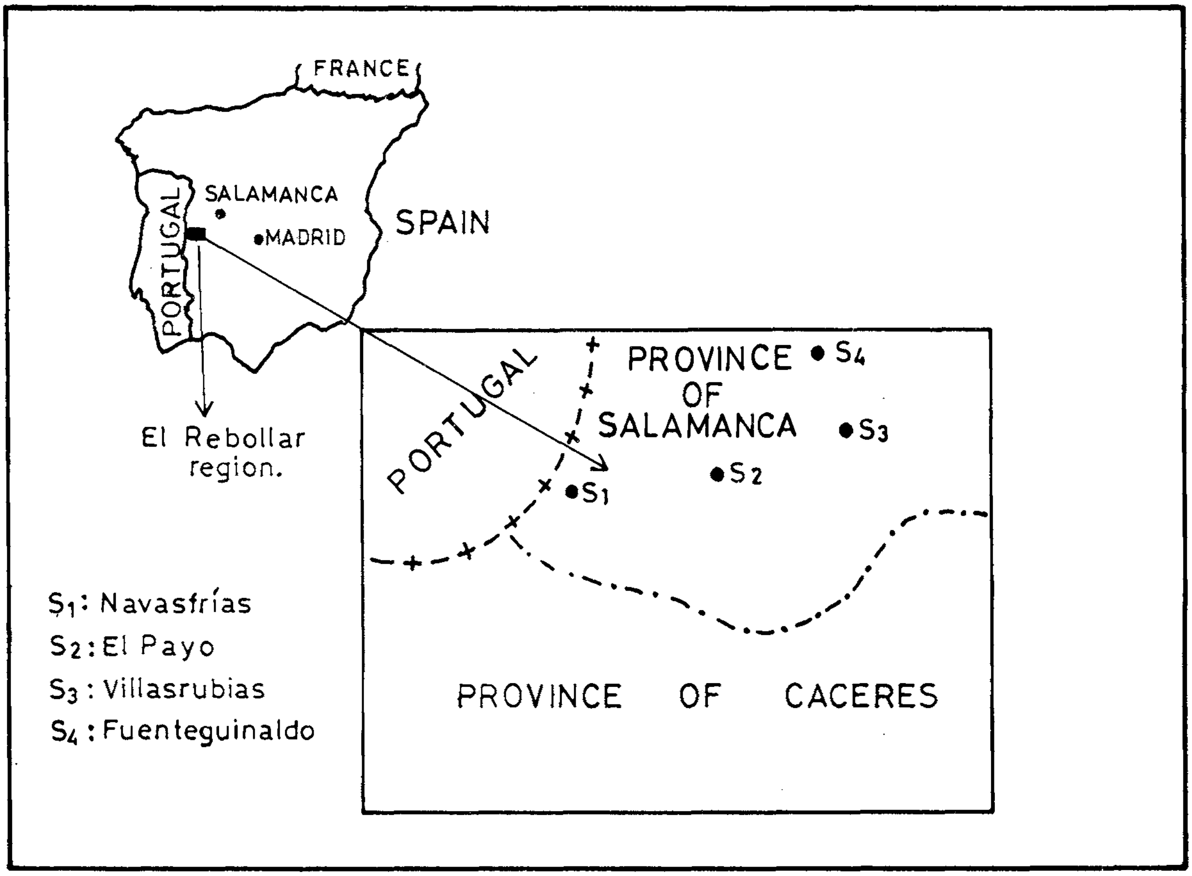

Figure 1. Localization map of the studied area.

sidered as a major determinant of vegetation growth as well as a vector for nutrient transport to and across the ecosystem. In this study we have tried to establish water and several bioelements balances in four *Quercus pyrenaica* forests along a marked pluviometric gradient, in order to obtain information on the effect of rainfall amounts on annual and summer evapotranspiration, and on nutrient leaching from the soils, and their fertility. Results are also related to productivity of these forests.

2. METHODS

2.1. The study area

This study was carried out in *Quercus pyrenaica* natural forests, classified as *Quercus robori-pyrenaicae* communities, located on the Northern face of the Sierra de Gata (40_ 2' 40" N; 3_ 0' 50" W, Salamanca Province, CW Spain; Figure 1). *Quercus pyrenaica* is a deciduous Mediterranean species, whose chorology corresponds to the southwestern region of Europe. The climate is humid Mediterranean, according to Emberger's climogram, most of the rainfall being concentrated in the cold part of the year, and dryness coinciding with the warmer season and the growing period. The soils are generally humic Cambisols (Gallardo *et al.* 1980), over Paleozoic granites and slates.

Four experimental plots, situated close to one another (maximum 15 km), were selected along a pluviometric gradient. The major characteristics of the plots are summarized in Table 1.

2.2 Field sampling procedure

The devices used in each plot for collecting water to chemical analysis are:

a) Above the canopy or in a large forest gap close to the plot:

- Three aerodynamically shielded rain gauges ("open gauge") for collecting bulk precipitation (**Bp**);

- Three funnels surmounted by an inert wind filtering

Table 1. Specific characteristics of the plots.

Experimental plot	Navasfrías (S1)	El Payo (S2)	Villasrubias (S3)	Fuenteguinaldo (S4)
Altitude (m.a.s.l.)	1000	940	900	870
P, mm	1,580	1,245	872	720
tm _C	11.4	N.d.	N.d.	13.3
Substrate	Slates and grauwackes	Calc-alkaline granite	Slates and grauwackes	Calc-alkaline granite
Dominant vegetation	Q. pyrenaica+ Pteridium+ grasses.	Q. pyrenaica+ grasses.	Q. pyrenaica+ scrubs+ grasses.	Q. pyrenaica+ abundant scrubs.
T.d. (N_/Ha)	820	406	1043	738
M.t.h., m	13	17	8.5	12
D.B.H., cm	15.2	25.4	11.0	16.5
L.A.I.	1.8	1.9	2.0	2.6

P, mm: Mean annual precipitation;
tm, _C: Mean annual temperature.
T.d.: Tree density.

M.t.h.: Mean tree height
D.B.H: Mean tree diameter at a height of 1.3 m.
L.A.I.: Leaf area index.

The S1, S2, S3, and S4 notation in Table 1 follows the decreasing order of precipitation and will be used hereafter in the text.

of polyethylene coated wire mesh ("filter gauges"), collecting bulk precipitation plus certain amount of dry deposition (**Fg**).

The 'filter gauge' enhaces the aerosol impaction, and the 'open gauge' minimizes this component in bulk precipitation (Miller & Miller 1980).

b) Beneath the trees:
- 12 standard rain gauges, randomly located, for collecting throughfall (**Tf**);
- 12 helicoidal gutters, around trunks, covering the basal area size ranges, for collecting stemflow (**Sf**).

c) On and in the soil:
- Six non-bounded gerlach type collector troughs in each plot (Sala, 1988) for measuring surface runoff (**Sr**).
- Soil solutions were collected using zero-tension lysimeters; six installed 20 cm below the soil surface, to collect water draining from the humic horizon (**SSr**), and other six installed at 60-100 cm of depth, to collect water of deep drainage (**D**)

The lysimeters were made from PVC, and the different type of rain gauges consisted of a polyethylene funnel. All devices were connected to a 5 l collecting bottle. Different filters were used to prevent contamination of water.

Water precipitation was recorded hourly with two tipping-bucket rain-gauges located above the crown in S1 and S4. Global shortwave radiation, air temperature, relative humidity and wind velocity were recorded as hourly means, using a data logger (Starlog 7000B Unidata) also only in S1 and S4.

Soil water content was measured with a neutron moisture gauge (Troxler 3321 A 110 mC of Americium/Berilium) at 12 access tubes in each stand. Soil moisture was measured every 20 cm from 20 to 100 cm depth, as a maximum, according to the depth of the soil. On the surface, soil moisture was measured by gravimetric method. Measurements were taken approximately once a month (occasionally every two weeks) over forty months: from March 1990 until September 1993. The calibration curves were determined from gravimetric samples and dry bulk densities, according to Vachaud *et al.* (1977).

The physical and chemical soil characteristics were studied in three selected profiles of each plot. The results have been discussed in previous papers (e.g. Moreno *et al.* 1993 and Quilchano 1993).

3.3. Calculation of water balance

Taking as a basis the hourly record in S1 and S4, the daily distribution of rainfall on S2 and S3 was estimated, once the high correlation existing between the distribution of rainfall on the four sites was verified. The records were also used to estimate the daily distribution of throughfall, taking into account the crown capacity for water retention (Zinke 1967). The Penman potential evapotranspiration was estimated following the recommendations of the FAO (Smith 1991).

The water balance equation was used as a basis: $dS/dt = Bp - EA - Sr - D$; (1)

where S is the soil water storage, Bp the precipitation, EA the actual evapotranspiration, Sr the surface runoff, and D the deep drainage, i.e. the flow of water below the root zone (including almost all the roots). This notation will be used hereafter. The precipitation, runoff and changes in soil water storage are readily measurable, but both EA and D are difficult to measure or to calculate.

Hence a water balance model was used which employed a simplified relationship between the drainage component and soil water content, characterizing the downflow of water across a certain level according to the water content existing above that level (this function is called the drainage characteristic; more detailed information in Rambal 1984 and Joffre & Rambal 1993).

The equation (1) is solved iteratively, for each period comprised between two readings of soil moisture, with a time step of one day (during periods of heavy precipitation, time step of one hour), i.e., starting at Sn and ending at Sn+1, fitting the term EA, the only unknown. The iterations continue until the measured and calculated value of Sn+1 coincide. It is always considered that $EA \leq EP$ (potential evapotranspiration).

When it is not possible to obtain this equality (1), a term known as "others" is introduced. It is probably a combination of EA, Sr and D. This is because in a rainy period EA can be greater than EP (Rutter 1975); moreover, even when the soil is not humid enough, deep drainage can exist, flowing through paths of rapid circulation, i. e. macropores (Beven & German 1981), which is not included in the calculation model used.

3.4.Calculation of nutrient fluxes

Above ground, water fluxes were measured, on an event basis (64 cases), immediately after each rainfall event, over the period 21 September 1990 - 20 September 1993. In 23 events, water was collected for chemical analysis.

The fluxes on a mass basis (kg ha^{-1}), for each parameter, were calculated multiplying the average weighted concentration (mg l^{-1}) by amount of water (mm), either measured (aboveground parameters and surface runoff) or calculated (deep drainage).

The net forest water (i.e., deposition in throughfall + stemflow, minus bulk deposition) is regressed against the gain in the deposition resulting from the aerosol deposition on the "filter gauge" (Bp - Pf). This regression results in an intercept term representing the mean value of canopy exchange (CE) for equal time periods (Lakhani & Miller 1980). Dry deposition (Dd) is calculated thus: $Dd = Tf + Sf - Bp - CE$, where Tf, Ef, Bp and CE are known. Total deposition (Tdep) from atmosphere is Bp + Dd. More detailed information can be found in Lakhani & Miller (1980) and Bellot (1988).

3.5. Laboratory analytical procedure

pH was measured with a pHmeter Beckman 3500, and dissolved organic carbon (D.O.C.) was measured on a T.O.C.A. (315A Beckman). These analyses were performed as soon as possible after collection (within first day). Na and K were analyzed by flame emission; Ca and Mg by atomic absorption spectrometry; Fe, Mn and Al, by ICP; P were determined spectroscopically by molybdenum-blue method; Cl^-, $NO3^-$, $SO4^{2-}$ and $NH4^+$, were analyzed by ion-chromatography. Complete analysis were generally done within about one week.

3.6. Statistical Analyses

Statistical analyses were performed by one-way analysis of variance with repeated measures, followed by a posteriori contrasts (Tukey-test) for multiple comparison of means, in order to detect variations between sites and between years. The results were expressed as equality probabilities (p). Simple regresion models were used to assess relationships between P and EA or D. The results were expressed as regression coefficients (r^2) and level of significance (p).

Table 2. Annual water balance of the four plots, in mm / year. % in relation to precipitation. The years refer to the period between 21/IX to 20/IX.

Year	Flows	S1 mm	S1 %	S2 mm	S2 %	S3 mm	S3 %	S4 mm	S4 %
1990 -91	Precipitation	1306		1188		1045		782	
	Interception	144	11	95	8	112	11	76	10
	Surface Runoff	2.1	0.2	1	0.1	3	0.3	0	0.0
	Deep Drainage	868	66	791	67	626	60	443	57
	Others	16	1.2	0	0.0	20	1.9	0	0.0
	Actual Evapotr.	408	31	382	32	385	37	325	42
	Potential Evapotr.	817		820		822		825	
1991 -92	Precipitation	777		658		610		442	
	Interception	147	19	160	24	112	18	96	22
	Surface Runoff	4	0.5	2	0.3	1	0.2	0	0.0
	Deep Drainage	212	27	170	26	111	18	19	4
	Others	49	6.3	15	2.3	25	4.1	13	2.9
	Actual Evapotr.	506	65	475	72	475	78	421	95
	Potential Evapotr.	806		819		832		845	
1992 -93	Precipitation	1086		953		820		650	
	Interception	163	15	131	14	127	15	125	19
	Surface Runoff	6	0.6	1	0.1	1	0.1	0	0.0
	Deep Drainage	404	37	373	39	239	29	106	16
	Others	54	5.0	18	1.9	20	2.4	8	1.2
	Actual Evapotr.	525	48	506	53	536	65	509	78
	Potential Evapotr.	753		763		773		783	

4 RESULTS AND DISCUSSION

4.1. Water balance

Precipitation

Table 2 shows the annual rainfall values for the three years, which, from the point of view of pluviometry and comparing them with the mean annual values (see Table 1) can be defined as normal (1990-91), very dry (1991-92) and moderately dry (1992-93). During the three studied years, the pluviometric gradient from which we started *a priori* was maintained; the differences between plots remained fairly constant during the three years, in relative terms (88, 78 and 59% of the rainfall in S1, for S2, S3 and S4, respectively). Annual precipitation differed significantly, among all plots and among all years (ANOVA, $p < 0.001$ in both cases). Nevertheless, daily rainfall distribution is similar in all the plots, with correlation coefficients around 90%. The seasonality of the rainfall and its acute irregularity are outstanding features; in this sense, in the 1990-91 period, rainfall was very abundant during autumn-winter but no important precipitations were recorded after mid-March. On the other hand, over the following two years the rainfall, although less abundant, was distributed more regularly with important precipitation recorded until the beginning of June. Averaged across all plots and years, the intensity of the precipitation was generally moderate and it was never above 14 mm h^{-1}. The intercepted water was 16% of the annual rainfall, and 14% during leafless period.

Evapotranspiration

EA only differs significantly among S4 - S1 plots (ANOVA, $p < 0.05$); S4 generally gives lower values, due to less precipitation and its smaller soil water storage (Table 2). These differences are mitigated even more if we subtract from EA the intercepted water, of little value for the vegetation (Rutter 1975). Between years significant differences are, in fact, found (ANOVA, $p = 0.001$), but with the lowest during the year of highest precipitation. Pooling together all plots, anual Bp and EA are uncorrelated ($r^2 = 0.17$, $p < 0.05$). The annual values of EA in relation to EP average 46% (range 39-50%), 55% (range 50-61%) and 72% (range 70-74%), in the years 1990-91, 1991-92 and 1992-93, successively. The moderate amount of water stored in the soil (Moreno *et al.* 1993) does not prevent there being a large water deficit during the active period. The maximum values of actual evapotranspiration in absolute terms were generally reached in June (sometimes May or July). Consumption begins to decrease generally in July, reaching very low values as early as August, when available soil water is practically depleted, remaining almost constant for approximately one month, depending on when the first autumn rains occur. Paz & Díaz-Fierros (1985) found in *Quercus robur* that the soil remained dry for two months in a year with 1368 mm of rainfall in the northwest of Spain; Joffre & Rambal (1988, 1993) obtained similar results in southern Spain. The daily mean values of EA, for the May-July period and for August, are shown in Table 3. Polling all plots, a positive relationship ($r^2 = 0.72$, $p < 0.001$) between total precipitation in the May-August period and the EA during this same period was found.

Comparing the monthly EA values of the four plots, significant differences were obtained between S4 and the other three plots, which showed slightly higher values; however, if the water intercepted is subtracted, these differences become considerably reduced. Daily EA distribution in the four plots is very similar, with correlation coeficients above 0.85.

Drainage

Drainage (D) increased with rainfall (Table 2), and significant differences were established both on the level of years and of plots (ANOVA, $p < 0.001$ and $p = 0.05$, respectively). Thus, pooling all plots, the existing relationship is: $D = 0.997*Bp - 494$ ($r^2 = 0.85$; $p < 0.001$), expressing D and Bp as mm year^{-1}. This relationships did not be set up for each plot indvidually due to a insufficient number of data (only 3 years).

These results indicate how easily an excess of water, and therefore deep drainage, can occur in these soils; a precipitation above 490 mm year^{-1} causes an excess of water in the soil, and practically all the rainwater over that figure will be drained. In other papers consulted, this limit was (mm·year^{-1}) 360 (Avila 1988), 400 (Piñol *et al.* 1991), 578 (Rambal 1984), 450 (Likens *et al.* 1977), 420 (Hudson 1988), 470-500 (Lewis 1968). Additionally, the rapid moistening of the deep horizons could imply the existence of water loss by drainage, due to rapid circulation in macropores (Beven & German 1981), although the soil is still far from reaching field capacity, a fact that is not included in the model used.

Surface runoff

The results indicate that the runoff decreases according to the following order: S1>S3>S2>S4 (table 2), and is well correlated with the slope of the plots (Table 1); the latter plot, with almost no slope, has null runoff. The greatest amounts were obtained for the year with the least precipitation. It occurs mainly in autumn (when rain intesity is greater); at no time the runoff was due to saturation of water in the soil. In these plots, according to our observations, this water did not transport appreciable amounts of sediment. The volume of water lost through surface runoff can also be said to be very low, as is frequently true in forest ecosystems (Rambal 1984; Francis & Thornes 1990; Soler & Sala 1992).

4.2. Nutrient balances

Atmospheric deposition

Amounts of atmospheric deposition can be described as moderate to scarce, in almost all the elements (Figure 2), when compared with those obtained in the northeastern Spain (e.g., Avila 1988; Bellot 1988; Belillas & Rodá 1991), Europe (e.g., Miller *et al.* 1987; van Breemen *et al.* 1988; Tietema & Verstraten 1991), U.S.A. (e.g., Likens *et al.* 1977; Lindberg & Owens 1993), and with mean values obtained by Parker (1983). They are especially low for ions such as $SO4^{2-}$, $NH4^{+}$ and $NO3^{-}$, which are mainly anthropogenic (Belillas & Rodá 1991), which in more industrialized regions show clearly higher values. There is no evidence of acid or polluted depositions of anthropogenic origin in this region. Regarding the differences along the pluviometric gradient, when the total deposition is considered (bulk and dry) the amounts were similar among plots.

Losses from the soil

The leaching of elements, nutritive or otherwise, from the soil, is a process controlled mainly by the content in anions and hydrogen ions, as the electrochemical equilibrium in the solution draining from the soil is maintained (Johnson *et al.* 1986).

The results obtained are shown in Table 4 and Figure 2.

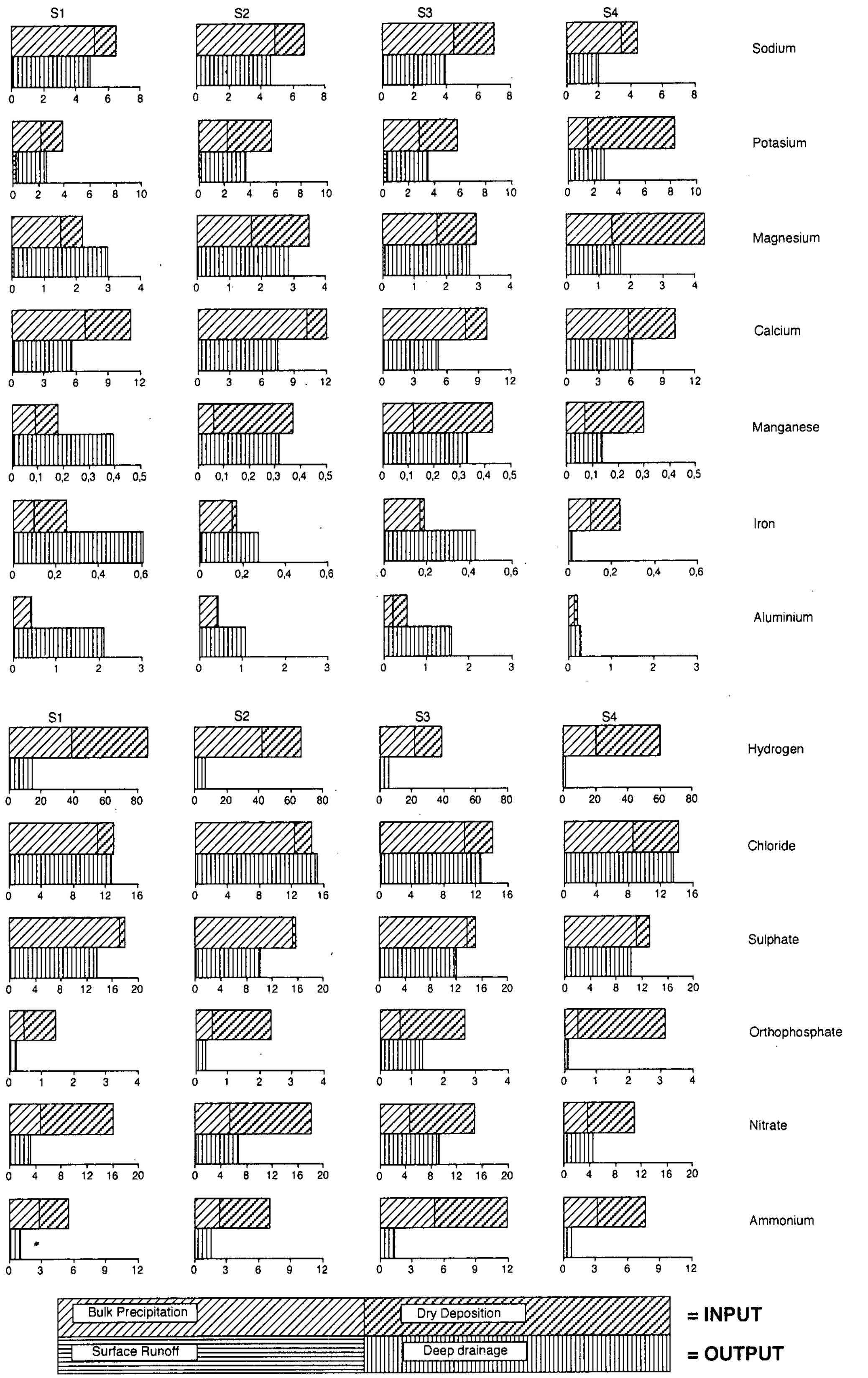

Figure 2. Annual balance of dissolved elements in four different plots (S1 to S4), calculated as the difference between input (bulk and dry deposition, upper bars) minus output (surface runoff and deep drainage, lower bars).
Values expressed as kg ha-1 year-1, except H+ (g ha-1 year-1).

Table 3. Mean daily values of actual evapotranspiration in two different periods, May-June-July (M-J-J) and August (A). Values expressed in mm day-1.

EA	S1 M-J-J	S1 A	S2 M-J-J	S2 A	S3 M-J-J	S3 A	S4 M-J-J	S4 A
1990	1.85	0.90	1.84	0.76	1.83	0.86	1.42	0.62
1991	1.83	0.43	1.49	0.74	1.76	0.94	1.33	0.45
1992	2.28	1.05	2.09	0.80	2.06	0.81	1.67	0.43
1993	2.59	1.14	2.59	0.99	2.73	0.90	2.65	0.80
Means	2.14	0.88	2.00	0.82	2.10	0.88	1.77	0.57

The scanty change produced in the pH of the water when passing through the soil is indicative of the reduced effect of the hydrogen ions on the leaching process of bases in the soils studied, except on the surface, where there is an important increase in the nutritive cations in the solution. In the rest of the soil the process is inverted, with a retention of bases and a leaching of hydrogen ions (Table 4). The increase in pH is higher in S2 and S4 (granite) than in S1 and S3 (grauwackes); also the increase is lower in the more rainy plots. These differences coincide with the content in base-cation and thus the buffering capacity of the soils (Quilchano 1993).

As for the anions, their behaviour is highly irregular, although all of them experiment a considerable increase on passing through the forest floor (Table 4). Almost all the phosphate is subsequently retained as a consequence of the high adsorption capacity of P that the soil minerals have (Yanai 1991). $NO3^-$, together with the cation $NH4^+$, show more complex behaviour, which will not be discussed in this paper. Cl^- and $SO4^{2-}$ pass through the soil with hardly any significant gains or losses; as a whole they may represent an important flow as regards the washing of cations; this is, however, not so much due to the fact that they represent fairly low atmospheric deposition rates, especially in $SO4^{2-}$, considered to be largely responsible for the leaching of bases (David *et al.* 1991).

Therefore, neither the hydrogen ions nor the anions seem to be an important source for causing the leaching of bases in the studied soils. Thus, although Na, K, Mg and Ca undergo a considerable increase on passing through the forest floor (especially K), together with an increase in pH, the concentration of all of them decreases again in both the humic and the exchange horizons (together with a decrease in pH) (Table 4). That is, the cations are retained in the soil (either by root absorption or by entry into the exchange sites), so that the losses from the soil are lower than the atmospheric inflow, there being thus a net gain (Figure 2). The order of these gains (% with respect to the Tdep) is: K > Ca = Na > Mg > Mn > Fe > Al.

The sum of the net losses (outputs - inputs) of the four major cations is known as the Cationic Denudation Rate (CDR, Avila 1988), expressed in keq ha^{-1} $year^{-1}$. In our case the results are: -0.24 (S1), -0.35 (S2), -0.24 (S3) y -0.57, i.e., the CDR is negative due to the existence of net gains. If we consider Bp instead of Tdep, in order to contrast it with the results in the literature, the results are: 0.14 (S1), 0.06 (S2), 0.07 (S3) y 0.10 (S4). In any case these values are well below the mean of 1.03 keq ha^{-1} $year^{-1}$, described by Avila (1988); of the 22 cases given by this author, 19 show net loss values higher than those obtained in our plots, two of them show similar values, and only one shows net losses lower than our case. The values in the 22 cases vary from -0.12 to 4.4. In the Mediterranean region, the values obtained are 1.3 in Montseny (Avila 1988) and 1.8 in Prades (Escarré *et al.* 1984).

The soils show a greatly decreased base saturation level, lower in the plots with higher rainfall (Quilchano 1993), hence the availability of bases, to be exchanged for H^+ or to accompany an anion, is scarce. This fact, together with the scant acidity of the precipitation and the anion content of the atmospheric inflow, contributes to the scant leaching of the bases of these soils.Another aspect of interest in obtaining a knowledge of the fertility conditions of the soil is the Ca/Al ratio (molar quotient) in the soil solution as Abrahamsen (1983) pointed out; this author, quoting Ulrich (1981), points out that a value of 1 or less indicates a degraded state of the forest soils, even phytotoxicity by Al. In our case the values obtained were:

	S1	S2	S3	S4
Ess	0.95	1.52	2.07	4.04
D	1.79	4.73	2.23	14.73

Table 4. Volume weigted mean annual chemical concentratios in four different flows (S1 to S4).

mg/l	pH	DOC	Cl	SO4	PO4	NO3	NH4	Na	K	Mg	Ca	Mn	Fe	Al
Tf-S1 (23)	5.45	9.9	1.9	1.61	0.15	0.53	0.26	0.53	0.85	0.50	1.36	0.050	0.030	0.05
Tf-S2 (23)	5.53	13.9	2.3	1.89	0.46	0.83	0.31	0.67	1.84	0.72	1.48	0.090	0.030	0.06
Tf-S3 (23)	5.64	15.7	2.5	1.92	0.43	0.83	0.32	0.70	1.85	0.85	1.62	0.130	0.040	0.10
Tf-S4 (23)	5.59	23.5	3.5	2.30	1.16	0.73	0.60	0.65	3.19	1.14	2.00	0.100	0.050	0.10
Sr-S1 (17)	5.73	48.8	3.07	2.73	0.40	1.56	0.78	0.73	5.87	1.59	2.58	0.176	0.120	0.34
Sr-S2 (16)	5.81	115.9	5.66	3.54	1.62	6.29	3.11	0.921	7.54	2.18	3.19	0.312	0.172	0.78
Sr-S3 (14)	6.22	81.0	6.75	6.92	2.50	4.12	2.53	1.131	6.35	3.55	4.06	0.348	0.154	0.60
Sr-S4 (7)	6.62	147.2	7.34	11.67	8.71	16.00	9.52	1.83	45.06	4.84	9.97	0.487	0.175	0.26
SSr-S1 (9)	5.74	21.0	3.09	3.81	0.03	0.49	0.60	1.02	1.72	1.08	1.08	0.034	0.038	0.77
SSr-S2 (10)	5.53	28.9	2.84	3.14	0.09	4.43	1.29	0.87	3.38	1.04	1.21	0.054	0.054	0.54
SSr-S3 (8)	5.89	57.2	3.29	2.93	0.25	6.67	1.22	1.00	2.88	1.89	2.74	0.060	0.080	0.90
SSr-S4 (5)	6.01	47.8	5.17	6.15	0.21	16.29	1.38	0.761	0.14	2.70	6.51	0.093	0.081	1.09
D-S1 (9)	5.49	8.4	2.81	2.96	0.04	0.67	0.23	1.06	0.54	0.63	1.22	0.085	0.079	0.46
D-S2 (5)	5.79	4.8	3.97	2.46	0.07	1.64	0.39	1.13	0.84	0.69	1.79	0.074	0.069	0.26
D-S3 (5)	5.72	17.2	4.46	4.22	0.45	3.28	0.49	1.41	1.14	0.94	1.85	0.116	0.152	0.56
D-S4 (3)	6.12	12.3	8.01	6.11	0.07	2.74	0.45	1.20	1.68	1.04	3.70	0.083	0.011	0.17

Tf= Througfall; Sr=Surface runoff; SSr= Subsurface runoff; D=Deep drainage.
The numbers of analysed events, for each flow and plot, are shown in n parentheses.

Ess is water draining from the humic horizons and D is water of deep drainage.

Although the Ca/Al values do not indicate a very unfavourable situation, common in areas with much more acid precipitation than that obtained in this study (Abrahamsen 1983), they do show the effect of the abundant precipitation on the soil fertility in this area, with S1 values which indicate a very low level of fertility, and with high levels of Al in the soil water. The situation improves (lower relative importance of Al) as the pluviometric gradient decreases.

In addition to the Ca/Al quotient, the net gains for K, Mg, Na, $NH4^{+}$, $H4PO4^{-}$, Fe, Mn, among others, increase when precipitation decreases (Figure 2).

5 CONCLUSION

Although the study dealt with situations where the precipitation was markedly different, both on the level of years and of plots (range from 442-1306 mm year^{-1}.), the soils had a similar water content in the different years at the beginning of the active growth period of the vegetation; the water content depends mainly on the soil characteristics and not so much on the precipitation received during the wet season (Moreno *et al.*, 1993).

Taking as a basis the soil water balance, we obtained decreasing transpiration rates over the active period, reaching acutely low levels. Unless the oaks have an efficient deep radicular system for extracting water from the weathered bedrock, they could be subjected to an important restriction of water during part of the summer season. Moreover, the vegetation shows great dependence on the rains of the dry season. The water gradient does not seem to define outstanding differences in the water consumption patterns for the vegetation in this area.

The relative facility with which an excess of water in the soil is produced in winter, together with the high correlation existing between the volume of rainwater and that of drained water, cause an important leaching of the soil and a consequent loss of fertility, which becomes greater as the pluviometry gradient increases. This fact is evident in the figures of the CDR and the Ca/Al ratio. In fact, both on the level of dry matter production and leaf area index, the higher values are obtained in the plots with less precipitation (Martín *et al.* 1993; Gallego *et al.* 1994).

It can be concluded that the greater abundance of rain in the wet season did not tend to increase water consumption by the vegetation, at least not substantially; on the other hand, it did entail a greater leaching of the soil and the consequent loss of fertility, which is especially manifested in a decrease of bases, degree of saturation and pH, as demonstrated by Quilchano (1993) in these same forests.

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Hydrological implications of afforestation of abandoned lands: water balance simulation of a small Mediterranean mountainous basin

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ABSTRACT

The intense human activity suffered by Mediterranean mountainous areas till 50 years ago induced important geoecological modifications that played a significant hydrological and geomorphological role during agricultural land use and still have relevant consequences after land abandonment. Reforestation of these areas is perceived to suppose in the next future a serious water management problem, but more research is needed to quantify the magnitude of the hydrological implications of this change.

This paper presents a simulation model made to analyze the role that vegetation cover changes can play on water resources. This simulation model compares the actual hydrological behaviour of a small mountainous catchment covered by grassland with its behaviour modeled considering that the whole area is covered by a *Pinus sylvestris* stand.

Simulation results during a sensitivity analysis, with forest water consumption parameters taken from literature, show that differences in vegetation cover can modify all the water balance components but effect especially both quickflow and baseflow. The differences depend much more on the parameters used for the simulation of interception than on those used for stomatal control of transpiration.

INTRODUCTION

The abandonment of marginal agricultural fields since early this century appears as one of the main geoecological modifications in mountainous Mediterranean areas. Moreover the actual agricultural policies of the European Community and the Mediterranean countries encourage agricultural land abandonment in low productivity areas, like mountain ones. This would lead to the abandonment in the next future of 15 million hectares of cultivated fields in the European Community countries (Miller, 1992).

The main geoecological adjustment that has or will be produced in these abandoned fields is related to afforestation, either spontaneous or promoted by the landscape management policies. Although the afforestation has been promoted as a good land conservation practice, mainly in degraded areas, it can result in important problems on water conservation. A greater loss of water to the atmosphere in forested areas caused a diminution of water discharge (Calder, 1993). Bosch & Hewlett (1982) summarized some studies over the world that showed that the greater water consumption by forests is sufficiently significant to stop the afforestation of headwater areas policy.

Taking into account the present importance of land use change in mountainous areas, and also considering recent predictions of climate change, more research is needed to quantify the magnitude of the hydrological implications of reforestation. The aim of this paper is to present a basin water balance simulation model that illustrates the potentially different hydrological behaviour between two possible types of vegetation covering abandoned fields: the grassland or a spontaneous reforested *Pinus sylvestris* stand, in a Mediterranean mountain area with mean annual rainfall of 850mm.

THE CAL PARISA BASIN: AN EXAMPLE OF HYDROLOGICAL RESPONSE IN MEDITERRANEAN MOUNTAINOUS ABANDONED AREAS

The small Cal Parisa basin (36 Ha), located in the Eastern Pyrenees (Spain) was selected as a representative of agricultural abandoned fields in Mediterranean mountainous areas. This basin has been monitored since 1988 to analyze the role of land abandonment in water and land conservation (Llorens & Gallart, 1992). The main geoecological implications of old agricultural land use are the modifications in vegetation cover, topography and drainage net.

The old *Quercus pubescens* forest was removed and a system of terraces was constructed. Due to the hydrological modifications induced by topographical changes, especially the outcrop of phreatic water in the inner part of terraces, a net of man-made ditches was constructed modifying significantly the total drainage length (Llorens et al., 1992).

As a consequence of these geoecological modifications, the hydrological behaviour of the basin should exhibit:

a) A greater water storage capacity due to increased soil thickness produced by the terraced system.

b) An important modification of water circulation due to the remodelling of topography. This hydrological change is characterised by the formation of frequently saturated areas in the inner parts of terraces, and subsequently by the role of the man-made ditches.

Data obtained from the catchment monitoring show that in this conditions its hydrological behaviour is characterized by (Llorens, 1991):

a) A hydrological response strongly regulated by the basin antecedent water reserve.
b) A runoff generation by saturation mechanisms.

c) A quick stormflow response after basin saturation.
A comparison of this hydrological functioning with the behaviour predicted for a natural catchment, with the same general topographic structure, was performed (Gallart et al., in press) using the fundamentals of the semidistributed hydrological model TOPMODEL (Beven & Kirby, 1979). This analysis confirmed that terraces clearly modify the pattern of saturated areas, and suggested that the water balance of the basin would be consequently modified towards a higher saturation storm runoff at the expense of lower baseflow and evapotranspiration.

SIMULATION OF CAL PARISA BASIN HYDROLOGICAL BEHAVIOUR: THE SIMBAL MODEL

The SIMBAL model was built as a tool to better understand the hydrological behaviour of Cal Parisa grassed basin (Llorens & Gallart, 1990). The initial structure of the model considering only grassland cover, is:

- Model input: Daily measured rainfall (P).
- Initial variable: Initial soil water deficit (S_i).

- Evapotranspiration submodel:

- Potential evapotranspiration (E_t) is provided by a stochastic generator based on the annual distribution of Penman's potential evapotranspiration (Penman, 1948) measured in the basin (during 1989-90).

-The calculation of simulated actual evapotranspiration (E_a) is done considering that:

a) Simulated actual evapotranspiration is equal to the potential one for basin water deficit lower than 100 mm.

b) For basin water deficit higher than 100 mm, simulated actual evapotranspiration diminishes as a function of the basin water deficit (Hillel, 1971).

- Runoff and soil water deficit submodel:

- Baseflow runoff (R_b) depends on a negative exponential function of catchment water deficit based on the baseflow equation of TOPMODEL (Beven & Kirkby, 1979). After calculation of evapotranspiration and baseflow then basin water deficit is updated by losses.

- Quickflow runoff (R_q) is calculated using a simplified partial saturation model (one parameter). Basin water deficit is updated after the calculation of quickflow runoff to acount for infiltration.

- Outputs of the model: simulated actual evapotranspiration and runoff at daily scale.
The model was built using data measured from 1989 to 1990, and validated for the quickflow runoff with measured data from 1990 to 1992.

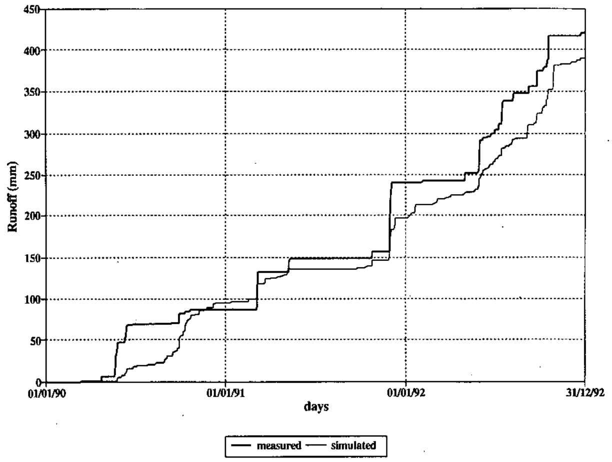

Figure 1. validation of simulated Quickflow Runoff using Cal Parisa measured data for the period 1990-1992

Figure 1 shows a cumulative graph comparing measured and simulated quickflow runoff (R_q) for the years 1990-92. The total difference between measured and simulated runoff is less than 8% for this validation period. The specific differences observed between measured and simulated runoff are attributed to the limitations of the single store model, made to simulate daily water balance. This is due to the occurrence of high and quick discharges produced by great rainfall events (up to 140 mm/event or up to 50 mm/day).

SIMULATION OF HYDROLOGICAL DIFFERENCES BETWEEN FORESTED AND NON FORESTED AREAS

Physical hydrological differences between grassland and forest

The hydrological differences between forest and grassland are deemed to be primarily due to two factors: the water losses by transpiration and by interception. Secondary changes induced by forest in soil hydrological behaviour (water retention, infiltration...) are not considered by the model.

- Water loss by transpiration:

In soil water deficit conditions transpiration of forested areas is certainly

lower than grassland one. Forest transpiration could represent 20-50% less than the grassland one (Wallace & Oliver, 1990), although forest has more evaporative potential because:

a) Its energetic budget is greater:

Forest albedo (about 7 to 8%) is lower than the grassland one (about 20 to 30%) due both to the dark colour of forest surface and to the radiation trap effect caused by forest structure (Oke, 1987).

The loss of long wave radiation over forest is lower than over grassland. Forests consist of rough surfaces generating greater eddies over them, which allow more sensible heat transfer and lower the surface temperature.

b) Its potential to transfer water vapour to the atmosphere is greater:

Atmospheric conditions over forest are more turbulent than over grassland, because forest is a rougher surface that has lower aerodynamic resistance, or atmospheric control of water vapour transfer (ra), (Monteith, 1965). For example for a wind speed about 3 ms-1, grassland aerodynamic resistance is about 50 sm-1 and conifers one is only about 3.5 sm-1 (Wallace & Oliver, 1990).

The difference in transpiration rate between forest and grassland for the same atmospheric conditions is controlled by the surface resistance, or vegetation physiological control (r_s), which is 3 times greater in forest than in grassland. Considering the one leaf model of Monteith (1965) grassland surface resistance is about 50 sm^{-1} (Wallace & Oliver, 1990) and *Pinus sylvestris* one is about 150 sm^{-1} (Gash & Stewart, 1977). Because resistances (atmospheric and surface ones) act in series, surface resistance that depends directly on soil water stress (Rutter, 1975) is the main limiting factor for transpiration and provides the difference in transpiration rate between grassland and forest.

- Water loss by interception:

When vegetation canopy is totally wet, surface resistance (r_s) is zero. In these conditions there are two phenomena causing greater losses to the atmosphere over forest than over grassland:

A forest is capable to intercept a greater volume of rainfall because of its structure. A descriptive parameter of vegetation structure is the LAI (Leaf Area Index) that shows the relationship between leaf and soil surfaces. Typical values of measured LAI are about 2-3 $m^2\ m^{-2}$ for different kinds of crops and range from 2 to 5 $m^2\ m^{-2}$ for *pinus sylvestris* (Breda, 1993).

The aerodynamic resistance (r_a) over forest is about 15 times lower than over grassland, in wet conditions fo-

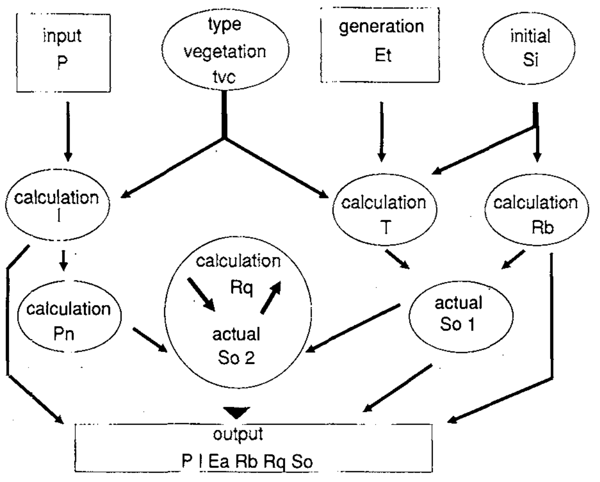

Figure 2. Flux diagram of the SIMBAL model. P= Precipitation; Ea= Actual evapotranspiration; Et= potential Evapotranspiration; I= Interception; Pn= Net precipitation; Rb= Baseflow; Rq Quickflow; Si= Initial soil water deficit; T= Transpiration; tvc= Type of vegetation cover.

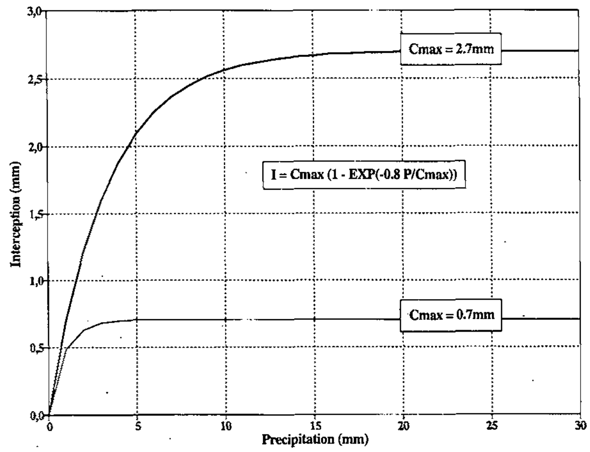

Figure 3. Relatioship between Precipitation and Interception used for the grassland and forest simulation. Maximum retention capacities are 2.7 mm for forest and 0.7 mm for grassland.

rest, and specially coniferous forest, are the best systems for water vapour transfer to the atmosphere.

Considering that for the same atmospheric conditions turbulent transfer is greater over forest than over grassland, the evaporation of intercepted water in grassland areas depends more on radiative energy supply than in forests. This difference is especially significant during or after rainfalls when the advective energy supply for evaporation is frequently greater than the radiative one (Oke, 1987; Rutter, 1975).

Simulation of the hydrological differences between grassland and forest

The simulation of these differences is performed by means of an interception submodel that simulates the loss of water intercepted by vegetation, and transpiration one that simulates the regulation of transpiration by stomatal control. (Fig. 2).

- The Interception submodel:

Daily interception (I) is simulated for rainy days using maximum retention capacity (C_{max}) and a negative exponential function (Fig. 3) depending on rainfall (Calder, 1990).

Maximum retention capacities in literature are about 0.7 mm for *Molinia caerulea* grass (Leyton et al., 1967), species representative of the basin grassland cover, and between 1.6 mm (Rutter, 1975) and 3.8 mm (Aussenac, 1968) for *Pinus sylvestris*. These maximum retention capacities were determined considering only continuous rainfalls with negligible evaporation. Using these parameters the model restricts evaporation of intercepted water to only evaporation between successive days.

Water volume intercepted by vegetation that exceeds maximum retention capacity is converted to a througfall-stemflow component and considered as a net precipitation (P_n). Difference in water volume between gross (P) and net (P_n) rainfall is evaporated differently in forest than in grassland areas:

a) Forest evaporates all the intercepted water. In these conditions the sum of daily transpiration plus interception can be greater than the Penman potential evapotranspiration calculated for grassland. This characteristic of the model allows the simulation of the evaporation by advection mechanism in forests that is underestimated by the classical Penman equation (Calder, 1990).

b) For evaporation of intercepted water in a grassed catchment, the model compares the volume of water intercepted with the meteorological evaporative demand (or the difference between actual and potential evapotranspiration). Then it evaporates a volume of intercepted water smaller or equal to this evaporative demand. This part of the model is based on the idea that in grassland areas intercepted water is evaporated depending mainly on radiative mechanisms that are suitably represented in the classical Penman equation (Calder, 1990)

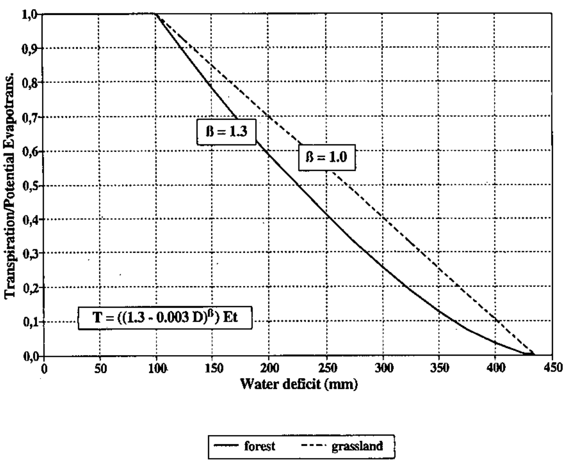

Figure 4. Relationship between water deficit of soil and the quotient transpiration versus potential Evapotranspiration used for the simulation of transpiration over grassland and forest

- The Transpiration submodel:

The decrease of transpiration in respect to the potential evapotranspiration acts, both in forest and in grassland, when basin water deficit is higher than 100 mm. This reduction of transpiration is as a function both of the basin water deficit and of the type of vegetation (Fig. 4).

This component of the model allows the simulation of a greater stomatal control by forest than by grassland in dry conditions. The equation presented in figure 4, and explicitly the ß coefficient (1.0 for grassland and 1.3 for forest) is used to simulate that transpiration of forest will be up to 50% lower that grassland one for the same atmospheric conditions (Wallace & Oliver, 1990).

Results of the simulation

To compare the simulated water balance for different vegetation cover in a sensitivity analysis exercise, the SIMBAL model was run using the 10 years daily precipitation data, since 1983 to 1992, collected by the Servicio Meteorológico Nacional (at the Vallcebre station near the basin). The initial conditions and parameters used in three executed runs are summarised in table 1.

The main results of the three model runs (table 1) considering the 10 years simulation are:

a) There are not important differences in simulated total water losses to the atmosphere (the sum of transpiration plus interception) between the two types of vegetation cover. Actual simulated evapotranspiration for each forest simulation are respectively 2 and 5% greater than grassland one (Fig. 5).- Mean forest simulated transpiration is about 87% of the grassland one.

- Simulated grassland interception is about 5.5% of the mean forest one (mean of forest1 and forest2).

b) There are not relevant differences in water deficit evolution between grassland and forest simulations (Fig. 6). Maximum annual differences are about 17 mm.

c) In spite of that there are very important differences in total runoff losses (the sum of quickflow plus baseflow) between the two types of vegetation cover.

This is due to that a little increase of water loss by evapotranspiration (transpiration plus interception) implies an important diminution in total runoff, because runoff is the smaller term of the basin water balance. Conside-

Table 1. Summary of parameters used in the three 10 years simulation runs of water balance.

Model runs	Period	Initial parameters		atmospheric losses parameters	
		Si (mm)	tcv	ß	C_{max} (mm)
Grassland	1983-92	100	grass	1.0	0.7
Forest 1	1983-92	100	forest	1.3	1.6
Forest 2	1983-92	100	forest	1.3	2.7

ß= Control of transpiration parameter; Cmax= Maximum retention capacity of intercepted water;

Si= Initial soil water deficit; tvc= Type of vegetation cover.

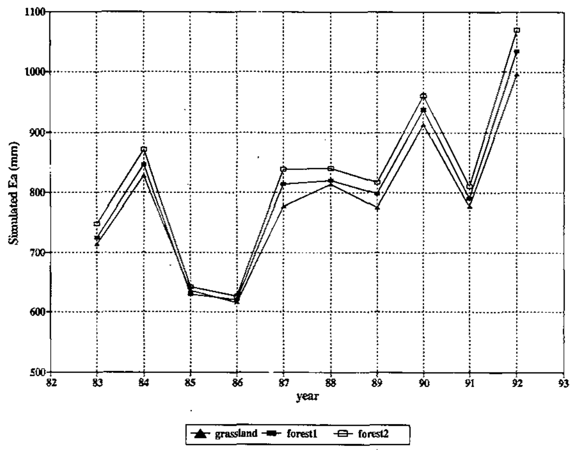

Figure 5. results of the 10 years simulation runs of water losses to the atmosphere. Simulated Actual Evapotransiration is the sum of Transpiration plus perception.

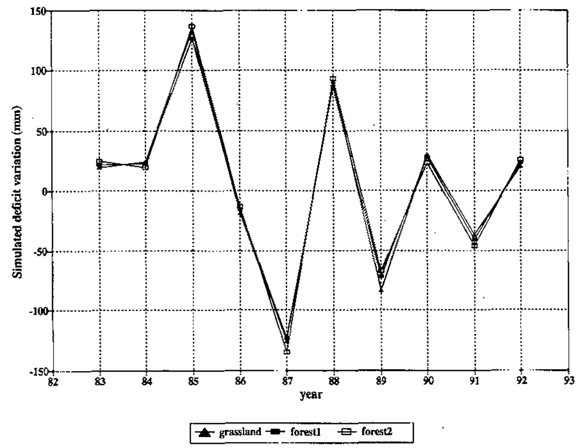

Figure 6. Results of the 1o years simulation runs of basin water deficit. The graphic presents the variations of catchment soil water deficit.

ring, as an example, the results of simulations of grassland and forest 2, the differences are: simulated actual evapotranspiration in forest2 is about 37.5 mm/year greater and runoff is about 35 mm/year smaller than in grassland. These similar figures represent an augmentation of 5% of evapotranspiration and a diminution of 32% of runoff.

Total runoff for each forest simulations are respectively 15 and 32% lower than grassland one (Fig. 7).

This difference is due mainly to the difference in the total simulated interception loss determined by the different maximum retention capacity parameters obtained for *Pinus sylvestris* from literature (Table 1).

Although simulated quickflow runoff is less affected by the change of vegetation cover than baseflow one, differences are not significant. Mean forest quickflow is about 78% of the grassland one and mean forest baseflow is about 74% of the grassland one.

d) The influence of vegetation on hydrological water balance is very sensitive to the used parameters, and specially to the interception maximum water retention capacity. A clear example of that is that difference between the two forest simulations depend directly on the maximum retention capacity figure used for the same type of tree species.

DISCUSSION AND CONCLUSIONS

As shown by the sensitivity analysis results, the variation of vegetation cover type can provoke important differences in total runoff due mainly to the increase of rainfall interception by the forest stand, even using an interception model that not allows evaporation of intercepted water during storms.

The simulation results using this conservative interception submodel predict a diminution of water yield due to afforestation of about 30%. This is about 10-20% lower than the figures presented by Bosch & Hewlett (1982) for some studies with similar mean annual precipitation.

Although in terms of total water losses to the atmosphere, part of interception loss by a forest will be balanced by a lower transpiration loss, a small modification of this water balance component implies an important variation of the runoff volume because it is the smaller component of the balance.

In order to design the most reasonable management of abandoned areas looking to the water conservation, it is necessary to better understand the hydrological mechanisms under different types of vegetation, by means of:

- The development of field experiments to obtain more accurate data of rainfall interception process. There are not sufficient experiments on conifers in the pluviometric range 600-1200 mm (Bosch & Hewlett, 1982) and especially under Mediterranean climatic conditions.

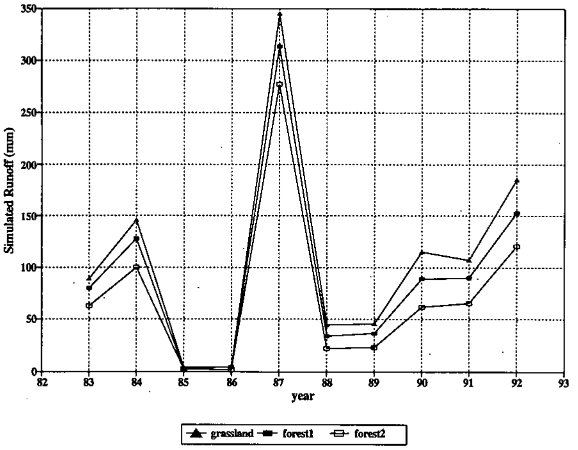

Figure 7. Results of the 1o years simulation runs of basin Runoff. Total Runoff is the sum of Quickflow plus Baseflow.

- The advancement of modelling runoff and actual evapotranspiration to lead as a final step to the construction of distributed models.

As described by Calder (1993), it would be possible to obtain a regulation of interception losses of a forest stand. Management strategies can consider, as an example, that forest can help water conservation in strong water deficit areas due to the better transpiration control by trees, whereas it could be a bad management practice in areas dominated by winds where trees are affected by higher evaporation of intercepted water.

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