Application of numerical modelling, isotope studies and streamflow observations for quantitative description of hydrogeology of the Kouris catchment (Cyprus)

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Application of numerical modelling, isotope studies and streamflow observations for quantitative description of hydrogeology of the Kouris catchment (Cyprus)

A dissertation submitted to the Swiss Federal Institute of Technology (ETH) Zuerich for the degree of Doctor of Natural Sciences

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SUMMARY

The Kouris catchment in Cyprus is experiencing water scarcity problems due to semi-arid conditions in its southern part and an increase in ground- and surface water extraction. Thus, quantification of the regional water balance is essential. But, due to rock heterogeneity and high spatial and temporal variability of water balance components, this is also a difficult scientific task.

The catchment is composed by two main geological zones: an ophiolitic complex in the North that contains the major groundwater resources of the island and overlying sedimentary rocks in the South with few water resources. The area is drained by three rivers - Kouris, Krios and Limnatis. Their valleys are filled with alluvium.

In this study, steady state and transient groundwater flow and transport numerical modelling was applied to quantify the regional water balance. Main input data (transmissivities, aquifer boundaries, storativity, recharge) were obtained from field investigations and historical records. During the years 1998-2002, 176 water samples (surface water, groundwater, precipitation) were analyzed for tritium and 224 – for stable isotopes. Additionally, all water samples were analysed for the major ions concentrations. These data allowed to validate model assumptions. The numerical models were calibrated by piezometric measurements in boreholes, baseflows, spring discharges and isotope data. During the calibration procedure sensitivity to boundaries, transmissivity, storativity and recharge were characterized. Recharge was additionally calibrated by deuterium data, while porosities were estimated from simulation of tritium transport.

Stable isotope data in the precipitation over the Kouris catchment allowed to construct a Local Regression Line described by the equation: $\delta D = 6.6\delta^{18}O + 10.9$. Seasonal and altitude variations of isotope content in the precipitation appeared to be different from those in the groundwaters. This fact can be explained by evaporation during infiltration. Thus, the deuterium altitude effect in recharge had to be derived from the data of 33 springs (rather than from precipitation) and was described by a linear regression equation. That equation allowed to calculate the altitudes of recharge for all sampling points and to understand the groundwater origin. It was demonstrated,
that the ground water in the alluvium originates mainly from the ophiolites, whereas, the groundwaters in consolidated rocks originate from local recharges.

The tritium input function for the precipitation in Cyprus was constructed to study tritium distribution in the aquifer. The groundwater residence times were calculated by the PMPATH advective transport model with input velocities from 3-D transient groundwater flow model of the Kouris catchment. The groundwaters of the ophiolitic aquifer discharging from springs had residence times of 1-25 years; the "young" waters appeared in wet seasons. Residence times of the groundwater in the sedimentary aquifer were estimated to be higher than 45 years.

Combination of the tritium input function in the precipitation and PMPATH residence times resulted in the calculations of the transient tritium concentrations in the aquifer. Finally, the calibration of the PMPATH model by observed tritium concentrations allowed to establish the optimal distribution of the porosity between 0.04 and 0.06 for the ophiolitic aquifer.

Numerical modelling under the steady state assumptions allowed to quantify the regional water balance for the hydrological year 1988/89. It was calculated that all recharged water from precipitation (30.2 Mm$^3$) was discharging via springs (4.1 Mm$^3$) and the rivers Kouris (10.4 Mm$^3$), Kryos (1.3 Mm$^3$) and Limnatis (12.7 Mm$^3$). Groundwater outflow to the Mediterranean sea was estimated to be lower than 10$^5$ m$^3$/year. The model results also included the calibrated set of hydrogeological properties and transient groundwater balance for 1984-2000. Actual evapo-transpiration from the alluvial aquifer during dry season was estimated using continuous streamflow records and was at the range of 1.7-3.3 Mm$^3$.

Combination of numerical simulations and isotope studies resulted in quantitative estimates of steady state recharge (90-140 mm/year).

Finally, a map of the areas endangered in case of increase of groundwater extraction was proposed.
RESUME

Le bassin du Kouris présente des problèmes de sécheresse du fait de conditions climatiques semi-arides dans sa partie sud et d’une exploitation croissante des ressources en eau du sol et du sous-sol. Par conséquent, la quantification du bilan hydrique régional est une question essentielle. Mais en raison de l’hétérogénéité géologique du sous-sol et de la grande variabilité spatiale et temporelle de tous les paramètres contrôlant ce bilan, cette question pose de nombreux problèmes scientifiques.

Le bassin versant est constitué de deux grands ensembles géologiques. Sa partie nord consiste en un complexe ophiolitique qui contient la majeure partie des ressources en eau de l’île. Dans sa partie sud les formations sédimentaires qui recouvrent les ophiolites ne présentent pas de réserves importantes. La zone est drainée par les trois rivières Kouris, Krios et Limnatis. Le remplissage des vallées est constitué d’alluvions.

Dans cette étude, la modélisation numérique à l’équilibre puis en système transitoire de l’écoulement et du transport a été utilisée pour quantifier le bilan hydrique régional. Les principales données entrées dans le modèle (transmissivité, limites des aquifères, storativité, recharge) proviennent d’une part grâce des études de terrain et d’autre part de données publiées. De 1998 à 2002, 176 échantillons d’eau (eau de surface, eau souterraine et précipitations) ont été analysés pour le tritium et 224 pour les isotopes stables. Ces données ont permis de valider les concepts sous-tendant les modèles. De plus, les modèles ont été calibrés à partir de mesures piézométriques dans les sondages, les nappes phréatiques et les résurgences, ainsi que par des données isotopiques. Pendant la procédure de calibrage, la sensibilité du modèle aux limites géographiques, à la transmissivité, à la storativité et à la recharge ont été analysées. La recharge a de plus été calibrée à partir des données sur le deutérium, alors que les porosités étaient estimées à partir de simulations de transport du tritium.

Les compositions des isotopes stables ont permis de calculer une Droite de Régression Locale décrite par l’équation : \( \delta D = 6.6 \times \delta^{18}O + 10.9 \). Il est apparu que les variations saisonnières et en fonction de l’altitude différaient dans les eaux souterraines et les précipitations. Ceci peut s’expliquer par un processus...
d'évaporation pendant l'infiltration de l'eau. Pour cette raison, il a été nécessaire de déduire l'effet d'altitude sur le deutérium à partir de l'étude de 33 sources. Cet effet peut être décrit par une équation de régression linéaire. Cette équation a permis de recalculer l'altitude de la recharge pour tous les échantillons, et de déduire son origine. Il a ainsi été démontré que l'eau souterraine dans les alluvions provient essentiellement du complexe ophiolitique, alors que la recharge est locale dans les formations indurées.

La fonction d'entrée du tritium a été élaborée pour étudier la distribution du tritium dans les aquifères. Les temps de résidence ont été calculés par le modèle de transport advectif PMPATH en utilisant les valeurs de vitesse utilisée dans la modélisation 3D en régime transitoire des écoulement dans le bassin versant du Kouris. Dans le complexe ophiolitique, les eaux souterraines au niveau des sources ont un temps de résidence variant de 1 à 25 ans, les eaux les plus « jeunes » apparaissant pendant la saison humide. Dans l'aquifère sédimentaire, les temps de résidences calculés sont supérieurs à 45 ans.

La combinaison d'entrée du tritium dans les précipitations et des temps de résidences calculés par le modèle PMPATH a permis de calculer la concentration transitoire du tritium dans l'aquifère. Enfin, le calibrage du modèle PMPATH par les données de tritium mesurées a permis d'établir une distribution optimale de la porosité entre 0,04 et 0,06 pour l'aquifère ophiolitique.

La modélisation numérique à l'équilibre a permis de quantifier le bilan hydrique pour l'année hydrologique 1988-89. Il est apparu que toute l'eau rechargée (30,2 Mm$^3$) s'écoule par les sources (4,2 Mm$^3$) et les rivières Kouris (10,4 Mm$^3$) Kryos (1,3 Mm$^3$) et Limnatis (12,7 Mm3). L'écoulement souterrain vers la Méditerranée est estimé inférieur à $10^{5}$ m$^3$/an. Les résultats du modèle incluent également un ensemble de propriétés hydrogéologiques et le bilan hydrique pour la période 1984-2000.

L'évapotranspiration réelle à partir de l'aquifère alluvial pendant la saison sèche a été estimée entre 1,7 et 3,3 Mm$^3$ en se basant sur des enregistrements en continu du débit des rivières. La combinaison des simulations numériques et des études isotopiques a permis une estimation quantitative de la recharge à l'équilibre entre 90 et 140 mm/an. Enfin, une carte des zones à risque est proposée dans le cas où l'exploitation des ressources en eau viendrait à augmenter.
Chapter 1

Introduction
1.1. GEOGRAPHY AND GEOLOGY OF CYPRUS

Cyprus is situated at the north-eastern part of the Mediterranean basin, 33° east of Greenwich meridian and 35° north of the Equator (Fig. 1.1). It is the third largest island in the Mediterranean Sea with an area of 9251 km². It is dominated in its topography by two mountain ranges: the Troodos range in the central part of the island, rising to a height of 1952 meters and the Pentadaktylos range in the north of the island, rising to a height of 1085 meters. The Morphou and Messaoria plains lie between the two ranges.

Geophysical studies indicate that Cyprus is located on the upper plate of a northward dipping, active subduction zone (Kempler and Ben-Avraham, 1987; Anastakis and Kelling, 1991). The trench, associated with this subduction zone, runs approximately parallel to the southern and south-western coastline of Cyprus. The island of Cyprus and its immediate surroundings appear to be a segment of an oceanic crust and mantle which separated from the edge of one of the two major plates by rift-faulting, possibly during late Cretaceous times (Gass and Smewing, 1973; Robertson, 1975). The Troodos ophiolite complex (which is widely accepted to be one of the best exposed on the Earth) occupies an area of nearly 3000 km² in the south-western part of the island. It is an oval shape, approximately 130 km in length and 35 km in width. The massif is dissected by a number of normal, reverse and minor thrust faults and comprises a dome with a displaced slab of altered ultra basic and basic plutonic complex, stratigraphically overlain successively by a sheeted dike complex, extrusive sequence, and pelagic sediments (Moores and Vine, 1971; Gass, 1960). The plutonic complex is exposed in a central uplifted core, where it consists of tectonised harzburgites, dunites and serpentinites, stratigraphically overlain by gabbroic cumulates. Extensive late tertiary differential uplift, centered on a serpentine diapir, is responsible for the domal structure of the Troodos Complex (De Vaumas, 1959; Gass and Masson-Smith, 1963; Robertson, 1977). Due to subsequent erosion, the original upward succession of plutonic rocks, sheeted dike complex and pillow lava, was arranged in an outward succession from centrally exposed mantle rocks to the plutonic rocks, the sheeted dike complex and peripheral Pillow Lava.

Along the margins of the Massif, the volcanic rocks are unconformably overlain by prevailing calcareous sediments of Maastrichtian to Tertiary age which are followed by Pliocene-Pleistocene marls, sands and gravels.
1.2. CLIMATE AND WATER SCARCITY IN CYPRUS

The climate of Cyprus is typically Mediterranean, featuring essentially two seasons: hot and dry summers from May to October and mild, rainy winters from November to March (F.A.O., 1971). The average annual rainfall varies from less than 300 mm to more than 800 mm according to topography. The rain falling from November to March is about 80% of the total yearly amount. The average temperature in July and August ranges between 36°C on the central plain and 27°C on the Troodos mountains, while in January average temperatures are 5°C and 0°C respectively. Potential evaporation, estimated by Class A Pan varies from 1200 to 1750 mm per year, depending on surface elevation.

Since the beginning of the 20-th century the annual average rainfall amount for Cyprus decreased from 500 mm to 430 mm (informal communication with Dr. Stylianos Pashiardis, Meteorological Service, Nicosia, Cyprus). Additionally, the percentage of spring rainfalls in annual average increased (Kypris, 1995). This fact plays a negative role for water resources of the country since normally only winter rainfalls contribute to recharge, while those in spring were almost completely lost due to evapotranspiration. An increase of the average temperature in Cyprus (approximately in 1°C over the last century - Price et al., 1999) is worsening the situation, because of consequent increase of potential evapotranspiration.

Most of the water resources from Cyprus results from rainfall, only a few percent are obtained from desalination and recycled water. The annual total rainfall amount for the island is around 4600 Mm³ (Fig. 1.2) and 80% of it returns to atmosphere by evapotranspiration (Omorphos et al., 1996). From the remaining 900 Mm³, approximately 600 Mm³ contribute to surface runoff while 300 Mm³ recharge aquifers. Out of recharged groundwater, 270 Mm³ are usually pumped or discharged from springs and about 70 Mm³ are lost into the sea through the aquifers. Thus, Omorphos et al. (1996) estimates annual reduction of groundwater storage due to over-pumping as 40 Mm³ for the whole island.

Most of rivers have their sources in the Troodos Mountains. From 600 Mm³ of surface runoff, about 150 Mm³ are diverted from the rivers and used for irrigation. Reservoirs for collecting water during rainy seasons have been constructed on almost all rivers. The water from reservoirs is used during summers for irrigation and domestic
supply. The total storage capacity of surface reservoirs is around 300 Mm$^3$, but annually the available yield of these dams hardly reaches 190 Mm$^3$ (Omorphos et al., 1996).

Cyprus belongs to countries with scarce water resources. Margat and Tiercelin (1998) give an estimation of 900 Mm$^3$ per year of renewable water resources for the whole island. In comparison, Libya has 600 Mm$^3$ per year and Switzerland has 50 000 Mm$^3$ (Seckler et al., 1998).

A general management plan for surface and ground water does not exist yet for Cyprus (Matsis, 1999). In addition, because the water demand is higher than the available water resources, the situation is worsening. Recently, flow into the reservoirs dramatically diminished due to several dry years and an increase of pumping upstream of the dams. Additionally, continuous decreases of water levels (see Figure 1.3 as an example for the Kouris catchment) indicate a reduction of groundwater storage over the last 15 years. This fact resulted, for example, in drying riparian vegetation in the upper part of the Limnatis subcatchment and vanishing several springs in the same area.

Other water resources problems in Cyprus are related to saltwater intrusions into fresh aquifers due to overpumping, which affect almost all coastal regions.

During the last summers, water supply restrictions have been applied to save drinking water. Only in 2001, the construction of a second desalination plant improved the situation, however this way of solving water conflicts could not be considered as economically efficient and sustainable.

1.3. STUDY AREA

The Kouris catchment (an area of about 300 km$^2$) is located on the southern side of the Troodos Massif (Fig. 1.1). Within 30 km, the topography goes from an altitude of 2000 m in the North to sea level in the South (Fig. 1.4). The topography is typical of mountainous area with rapid changes in altitude from valley to valley.

The simplified geological map of the Kouris catchment (Fig. 1.5) shows the two main geological zones: the ophiolitic complex in the North and the overlying sedimentary complex in the South. The schematic crossection of the area is presented in Figure 1.6.

The ophiolitic sequence is ranging from Pillow Lavas at the stratigraphic top through a Sheeted Dyke Complex, Gabbros and Dunites to tectonised Harzburgites and Serpentinites at the north-western part of the catchment.
The rocks of the Mafic and Ultramafic Complexes are faulted, brecciated, highly fractured and sheared (Fig. 1.7). Jointing is very well developed and most probably extends to depth. Joints and faults of different scales are fairly open at ground surface and appear to be the main pathways for infiltration (Afrodisis et al., 1986). All major springs within this complex appear to be fault controlled.

In contrast to the Plutonic Complex, no large springs are found in the diabase of the Sheeted Dyke Complex - Diabases (Fig. 1.8). Infiltration appears to be lower due to less pronounced faulting and alteration. The Sheeted Dyke Complex is heterogeneous because it is constituted by a series of dykes of different ages with different compositions.

The Pillow Lavas (Fig. 1.9) are generally of limited hydrogeological importance, probably because the faults and fractures are usually filled and cemented with zeolic, tuffaceous or calcareous material.

The ophiolites have been altered either by hydrothermal circulation and weathering during their formation in an active rift or by recent weathering at surface. The alteration areas appear in the surface at scales from square meters to square kilometres (Fig. 1.10) and there is no information concerning their extending to depth.

The Ophiolitic Complex contains the major groundwater resources of Cyprus; the water is stored in fractured and altered zones of the Harzburgites, the Dunites, the Gabbros and the Diabase Dykes. A statistical study, carried out by Afrodisis et al. (1983) indicated, that the Gabbros were the best aquifers, the Pillow Lavas - the poorest and the Diabase and ultramafics were of intermediate quality.

Groundwater in the ultramafic rocks, the Gabbros and the Diabases is generally of CaMg-HCO₃ and Na-HCO₃ types with low to moderate salinities (200-600 mg/l), cation and anion compositions vary considerably with local lithology. The groundwater is usually of drinking quality, except for some springs (Kaoras, Agious Theodoras springs), already polluted due to leakages from the surface. The amounts of TDS are higher in Pillow Lavas, than in the ultramafic, plutonic and intrusive rocks (500 mg/l to 950 mg/l) and, within the anion content, SO₄ - plays the major role.

The contact between the ophiolites and the sedimentary rocks has a tectonic origin and has been investigated within quite a number of studies (Gas et al. 1994), however its slope, orientation and hydrogeological properties within the area are still unknown.
The Sedimentary complex (Fig. 1.11) in the southern part of the Kouris catchment unconformably covers the Pillow Lavas and includes the Lefkara sequence (upper Cretaceous – Tertiary) and the Miocene Pakhna formation. The limy sediments of the Lefkara are very heterogeneous, they consist of massive limestones, limy marls and calcarenites. The thickness of the formation is more than 300 meters. The Pakhna sediments include considerable amount of gypsum, and its thickness can reach 300 meters. The sedimentary complex is considered to have a low groundwater storage due to high actual evaporation and both low permeability and rainfall. From boreholes, drilled in these rocks for irrigation purposes, 70% were unsuccessful. Groundwaters in the hard sedimentary rocks are generally moderately mineralised to salty, the salinity varies between 600 and 1600 mg/l. They are of Na-Ca-HCO₃-SO₄, Na-Ca-SO₄-HCO₃ or Ca-Na-HCO₃-Cl types and in some areas are not suitable for drinking.

The alluvium aquifer (sands, gravels – Fig. 1.12) constrains river beds of Kryos, Kouris and Limnatis in the southern part of the Kouris catchment. It is discontinuous, narrow (width up to 50 meters) and has a thickness of a few meters. It contains considerable amount of water, comparatively to the whole sedimentary complex. Groundwaters in the alluvium are less mineralised (500-600 mg/l), than in the Lefkara and the Pakhna formations; they are of Ca-Mg-HCO₃, Mg-Ca-HCO₃ or Mg-Na-HCO₃ types.

The thicknesses of the main rock aquifers are not known exactly but can be inferred from geological observations, and by analogy with other mountainous groundwater flow systems in fractured rocks. The total thickness of the Sheeted Dykes is around 500 m, while the thickness of the Gabbros is up to kilometres (Gass et al. 1994). The hydraulic conductivity, caused by fracturing, decreases with depth (van Everdingen 1995). In the Alps, fractured crystalline rocks can be conductive even at depths exceeding a kilometre (Ofterdinger et al. 2003). Therefore, we envision the ophiolitic system to be a very thick body with a rather high degree of fracturing and alteration for the first 100 to 200 metres and a progressive decrease of conductivity up to a depth of several kilometres. We can then expect very deep groundwater flow systems interacting with faster shallow systems. The Pillow Lavas formation has a thickness varying between 50 and 200 metres. The thickness of the sedimentary rocks varies between a few tens of metres in the north to more than 600 m in the vicinity of Lofou village, 9 km north-west of the Kouris Dam (Gass et al. 1994).
The Kouris catchment is drained by three rivers – Kouris, Kryos and Limnatis – and their tributaries (Fig. 1,12, 1.13, 1.14). The main river of the catchment, and also the largest in Cyprus, is the Kouris River. It had an average yearly runoff of 36 Mm$^3$ during the last 30 years. All rivers in the natural conditions discharge groundwater from springs during almost all summers. However, during the last years all water during dry seasons was either caught by dams (Trimiklini and Kouris dams at the Kouris River) or diverted to the irrigation fields (Limnatis and Kryos rivers); thus, in summers the rivers have not been flowing naturally any more. The response of the rivers to rainfall is quite fast – the maximum surface runoff appears after several hours downstream the Kouris catchment. In the lower part of the area, the Kouris dam was constructed in 1988. It has the highest storage capacity of Cyprus (110 Mm$^3$), although the maximum amount of collected water only reached 70 Mm$^3$ at the end of 80’s and it since then has had a negative trend. The water is collected for drinking purposes for the whole country.

1.4. SUMMARY OF PREVIOUS RESEARCH IN CYPRUS

Ophiolitic complex

Tectonics and lithology of the Troodos mountains (as well exposed examples of ophiolites) were attractive subjects for researchers over the last decades (Gillis and Sapp 1997; Varga et al. 1999; Gass and Masson-Smith 1963; Gass and Smewing 1973; Robertson 1975; Robertson 1977; Moores and Vine 1971).

On the contrary, no complete hydrogeological concept of the Troodos mountains existed, though some hydrogeological and hydrochemical studies were conducted in the beginning of 80-th within a Cyprus-German Geological and Pedological Project. The results of the studies were presented in technical reports (Afrodisis, 1983; Afrodisis et al., 1986) that included some information concerning the yields of boreholes and springs and the description of the chemical composition of groundwaters from different rock types. The authors related water occurrence to certain rock types and density of faulting, consequently they presented lineament maps and described geology with more or less details. Though, at that stage it seemed not to be possible to understand even conceptually how the tectonics and faulting influenced groundwater flow and whether it was reasonable to distinguish aquifers according to every rock type.

The LIFE project (Charalambides et al., 1998) assessed the existing and potential ground water contamination due to the asbestos mines in the north-west of the
Kouris catchment. Within this project, water and rock samples from the igneous part of the Kouris catchment were analysed. Lineament analysis, based on aerial photography, was also carried out, however later, in our studies, we did not observe any correlation of hydrogeological properties with distance from lineaments. Generally, the project was related to analysis of existing groundwater contamination, rather than to the description of hydrogeology.

**Sedimentary complex**

Coastal sedimentary aquifers in Cyprus have been for many years an important subject for investigation. The reason of such interest is, on the one hand high density of population, while on the other hand those regions have always experienced water scarcity problems due to semi-arid climate and salt water intrusions.

For the last 40 years, several attempts were made to describe quantitatively and to model the hydrogeology of coastal aquifers. The results of these studies are included in a few technical reports of the WDD of Cyprus; see Milnes & Renard (2003), Plothner et al. (1985) and Schmidt et al. (1988) for the Kiti area and Fink (1965) for the Akrotiri area as examples.

Jacovides (1997) roughly describes the hydrogeology of several riverbed, delta and pond areas (Yermasoyia, Xeropotamos, Kouris, Phassouri and Lanitis) related to the possibility of artificial recharge in wet seasons. The study of Kitching et al. (1980) deals with estimates of recharge from precipitation in the southern Mesaoria area. The authors compared lysimeter measurements with chloride mass balance and tritium peak methods in geochemical profiles. Finally, they obtained quite a good agreement of geochemical methods and too low estimates from lysimeters. The main conclusion of their study is the first estimated recharge rate in Cyprus: 50 mm per year that seems to be reasonable for the sedimentary complex at low altitudes.

**Isotope surveys**

In the 1970's–80's, big efforts were put on water isotope investigations in Cyprus. The aim of the studies was not only to make the first regional environmental isotope survey, but also to use $^{18}$O, $^2$H, $^3$H and $^{14}$C for obtaining new information concerning hydrogeology of the island.

In 1971-78, 175 water samples from Cyprus were analyzed for $^{18}$O, $^2$H and $^3$H within the research contract 1039/9 “Environmental Isotope Survey (Cyprus)”, carried
out by WDD of Cyprus with the technical and financial assistance of the International
Atomic Energy Agency. Water samples were taken from springs, boreholes, rivers and
snow in the Troodos Massif. The description of sampling points, analysis itself and
detailed results of the investigation are presented in a technical report (Jacovides 1979).
This research was very important as the first survey of environmental isotopes in
Cyprus, however it did not considerably improve understanding of hydrogeology of
ophiolites because of the high rock heterogeneity and large spatial area coverage.

Other isotope investigations were conducted in 1979-81 on the Lefkara chalk, 30
km to the East from the Kouris catchment boundary, by the Geological Survey of
Cyprus in cooperation with the Federal Institute for Geosciences and Natural Resources,
Hanover. $^{18}$O, $^2$H, $^3$H and $^{14}$C contents for 45 groundwater samples and consequent
conclusions about hydrogeology of the area are published by Verhagen et al. (1991).

Conclusions from the general state of research
Despite a few good attempts made to describe hydrogeology of Cyprus, it has not been
fully understood. For example, the sedimentary complex, which was studied in several
regions of Cyprus, was never considered to be connected to the ophiolites; additionally
it was often studied locally – with uncertain inflows/outflows through areal boundaries.
Calculations of water balance for the construction of dams in Cyprus, in most cases, did
not take into account the fact that major amount of water in the rivers of the sedimentary
complex originate from baseflow and springs discharge in the Troodos ophiolites. The
hydrogeology of the ophiolites was never carefully studied, though these rocks contain
the major groundwater resources of Cyprus.

1.5. THE OBJECTIVES AND THE SCOPE OF THE STUDY

Our studies aim to contribute in the understanding of Cyprus hydrogeology, particularly
the hydrogeology of the ophiolites.

The Kouris catchment was chosen for investigations for several reasons.

First, this area has a great importance for Cyprus because of the Kouris Dam.
Calculation of the total discharge in the Kouris Dam area must be based on surface and
groundwater modelling, while no groundwater models of the area existed before starting
the present study. The study was initiated by the EU project “Integrated water
management in Cyprus - Economic and institutional foundations”, where groundwater
models of Cyprus had to support economic estimates and finally result in new water pricing policy.

Second, the Kouris catchment is typical for Cyprus (including two main types of rock sequences – ophiolites and sediments), so the methodology and general conclusions can be applied elsewhere in the island.

Third, a quantitative description of the hydrogeology of this type of region is a challenging scientific task due to a mountainous environment, high heterogeneity of rocks and semi-arid climate in the southern part of the catchment. Assessment of water resources under semi-arid climatic conditions requires completely different methods and approaches (compared to traditional ones, developed for temperate climates). For example, the estimation of recharge, which under a temperate climate is simply the difference between rainfall, run-off and potential evapotranspiration, becomes a big and still unsolved scientific problem in semi-arid environment. Since recharge is small and highly variable in time and space, uncertainties of its estimation are often counted by hundreds of percents, while in humid conditions they can be easily reduced to a few percents.

The general aim of our study is, first, to create a conceptual hydrogeological model for the Kouris catchment, supported by all available and collected information. Second, this model must result in quantitative estimates of the main hydrogeological parameters and water balance components. Third, the parameter estimates and sensitivity analysis are used to give recommendations for further water resources studies in the region and for water management.

In parallel, the study attempts to use environmental isotopes for conceptual support of model assumptions and for calibration of numerical models. These ideas appeared decades ago, however every successful application of environmental isotopes for groundwater model calibration (see Kirk and Campana 1990; Mattle et al. 2001; Herweijer et al. 1985; Leduc et al. 2000 as examples) is still useful, because it is often very problematic to make quantitative conclusions concerning hydrogeology with environmental isotopes.

Another aim of the study is the quantification of actual evapotranspiration in semi-arid conditions during a dry season. This term of the water balance is usually very difficult to estimate, and we suggest a new methodology by means of continuous streamflow observations, which could be applied elsewhere.
1.6. MATERIALS AND METHODS

For the description of the groundwater flow and the calculation of the water balance we applied numerical modelling, hydrochemistry, isotope investigations and streamflow observations.

For numerical modelling finite difference MODFLOW, MT3D and MODPATH codes were used in the PMWIN environment (Harbaugh and McDonald 1996; Zheng 1990; Chiang and Kinzelbach 2001). Four types of models were developed:

1) A steady state 2-dimensional groundwater flow model (MODFLOW) was developed to check data compatibility, to calculate steady state groundwater balance, estimate the range of recharge and perform sensitivity analysis; additionally, locations of endangered areas, due to over-pumping, were obtained;

2) A steady state 2-dimensional transport model (MT3D) to simulate deuterium transport in the aquifer and to calibrate a steady state recharge rate;

3) A transient 3-dimensional MODFLOW model to simulate groundwater flow with transient recharges and river leakages and to calibrate storage coefficient;

4) A MODPATH model, based on transient velocities from the 3-D MODFLOW model, allowed us to simulate tritium transport in the aquifer and to calculate residence time for the groundwaters, discharging from springs in the ophiolitic complex.

We also used box models (Maloszewski and Zuber 1982; K. Zoellmann - http://www.baug.ethz.ch/ihw/boxmodel_en.html;) for calculating groundwater residence times under the piston flow assumption (Chapter 4).

The records from 76 pumping tests in single boreholes were re-interpreted with the PC software AQUITEST (Sindalovski 1999) using the Theis solution superimposed in time for variable rates and recovery.

Surface elevations, which were necessary for calculations of piezometric heads, numerical modelling and stable isotopes studies, were obtained from the 25 meters resolution Digital Elevation Model provided by Hall (1998). Geological boundaries and rivers locations in the Kouris catchment were corrected with a LANDSAT 5 satellite image of Cyprus taken the 11-th of July 1987. This LANDSAT image was also used for
constructing a map of Normalized Difference Vegetation Index (NDVI) for the Kouris catchment.

All data for the region were combined in several ARCGIS (http://www.esri.com/) projects for visual analysis, data interpolations and consequent input to numerical models.

For hydrochemical and hydroisotopical description, 224 water samples from the Kouris catchment were analysed for stable isotopes and 176 - for tritium. Totally, waters from 114 locations were sampled. A part of the analysis were done during diploma studies by Jorin (2001), Moll (2000), Oertli (2002) and Steiner (2000). Hydrogen and oxygen isotope ratios are expressed by $\delta$D and $\delta^{18}O$ respectively, where $\delta = [(R_{sample}/R_{standard})-1] \cdot 1000$ (%o), $R$ is the ratio of D/H or $^{18}O/^{16}O$ in sampled water ($R_{sample}$) or in Standard Mean Ocean Water ($R_{standard}$). All analyses were performed at the Institute of Hydrology of “Gesellschaft fur Strahlen- und Umweltforschung” (GSF) at Munich; the analytical errors ($2\sigma$) were 0.1 %o for $\delta^{18}O$, 1 %o for $\delta$D and 0.7 – 2.0 TU (1 TU = 0.118 Bq/l) for $^3$H. All samples were also analysed for chemical macro-components in the Hydrochemical Laboratory of Engineering Geology at ETHZ.

For the determination of actual evapotranspiration from the alluvial aquifer, we applied continuous measurements at one flow-gauging station from Water Development Department observation network, constructed downstream the River Limnatis. The station was equipped with a mechanical water level recorder; additionally, two loggers P-Log520-PA (Driesen + Kern GmbH) were set up at the same location for pressure and temperature record with the interval of 10 minutes. The measured data covered the period 30.10.2001 – 26.11.2001. One of the loggers was used to record atmospheric pressure and temperature, while another one was measuring water pressure in the river. Additionally, some manual measurements of a river flow were made at several locations in the River Limnatis by universal anemometer MiniAir2 (Schiltknecht).

1.7. STRUCTURE OF THE THESIS

The thesis consists of an Introduction (Chapter 1), four scientific chapters, and a Conclusions (Chapter 6).

The Chapter 2 corresponds to the paper “Groundwater resources in the Kouris catchment (Cyprus): data analysis and numerical modelling”, published in the Journal
of Hydrology (Boronina et al., 2003a). The preliminary version of this paper was presented at the International Groundwater Conference of IAH “Balancing the Groundwater Budget” in Darwin (Australia) in May, 2002. This paper includes a detailed description of the study area: geography, climate, hydrology, hydrogeology, water conflicts. Hydrogeological information is integrated into a regional 2-dimensional numerical groundwater flow model. The calibrated model allowed to estimate the range of steady state recharge, and to select the parameters limiting the uncertainty of the water balance calculations. Additionally one scenario of future groundwater extraction was simulated, showing locations of endangered, due to over-extraction, areas.

In the second and the third papers (Chapters 3 and 4), water isotopes were applied for the validation of the hydrogeological concepts of the Kouris catchment and for the calibration of numerical models.

The Chapter 3 corresponds to the paper “Study of stable isotopes in the Kouris catchment (Cyprus) for the description of the regional groundwater flow”, submitted to the Journal of Hydrology. Some results of this chapter were presented at the International Symposium on Isotope Hydrology and Integrated Water Resources Management in Vienna in May, 2003 and published in the Proceedings of the Symposium (Boronina et al., 2003b). Chapter 3, in the first part, presents the analysis of stable isotopes in precipitation over the Kouris catchment. The isotope contents of precipitation are compared to those in groundwater and illustrate the role of evaporation. In the next part, a linear regression of deuterium excess with altitude is used for identification of recharge areas. Finally the calibration of recharge of the steady state groundwater model by deuterium concentrations in the aquifer is also performed.

The Chapter 4 corresponds to the paper “Application of tritium in precipitation and in groundwater of the Kouris catchment (Cyprus) for description of the regional groundwater flow”, which will be submitted to the “Applied Geochemistry”. The preliminary results of the studies were presented at the International Conference “Hydrology of the Mediterranean and semi-arid regions” in Montpellier in April, 2003 and published in the proceedings of the conference (Boronina et al., 2003c). For these studies, a temporal input function of tritium in precipitation is obtained by a regression analysis with tritium data from several surrounding meteorological stations. This function is further used to model tritium transport in the aquifer and to calculate residence times of different waters in the catchment. For tritium transport simulations, the MODPATH code was used; the model implied transient piezometric field, simulated
by the transient 3-dimensional MODFLOW model. The chapter presents also the results of transient MODFLOW and MODPATH model calibrations. As one result of model calibration, the estimates of the storativity of the aquifer were made.

The Chapter 5 presents the paper "Calculations of actual evapotranspiration from an alluvial aquifer of the Kouris catchment (Cyprus) using continuous streamflow records". The paper was submitted to the Hydrological Processes. It describes possible causes of regular daily variations in streamflows and concludes that they originate from daily actual evapotranspiration. After calculation of actual evapotranspirations for the observation days and establishing, for those days, a linear correlation with potential Pan "A" evaporation, the actual annual evapotranspiration from the alluvial aquifer for the whole Kouris catchment is estimated.

The Conclusions (Chapter 6) present the synthesis of our knowledge about hydrogeology of the Kouris catchment.

The thesis also includes a CD with all data used for this study. The CD contains six folders. The folder RAW DATA consists of initial data, obtained from the Water Development Department of Cyprus, the Forest Department of Cyprus and the Meteorological Service of Cyprus. The folder ETH DATA includes the data, obtained during the PhD project (chemical database, piezometry measurements, river observations). The folder GIS consists of georeferenced files, which can be used for GIS projects of the Kouris catchment. Numerical and box-models developed during the study are included in the folder MODELS. All published and presented materials from the PhD projects are given in the folder PAPERS AND PRESENTATIONS. The folder PICTURES includes scanned maps and pictures of the area. Finally, the final folder THESIS contains final version of this thesis in pdf-format.

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Figure 1.1. Simplified geological map of Cyprus showing the location of the Kouris catchment.

Figure 1.2. General water balance of Cyprus (after Omorphos et al., 1996); all numbers are given in Mm$^3$.
Figure 1.3. Progressive increase in the depth of the ground water table in the extraction boreholes with serial numbers 53/76, 31/76, 57/76, 21/82 and 69/79 during non-pumping seasons (October to March).

Figure 1.4. Digital Elevation Model of the Kouris catchment (Hall, 1998).
Figure 1.5. Simplified geological map of the Kouris catchment: black circles show locations of boreholes drilled for irrigation purposes (some of them were unsuccessful).

Figure 1.6. Schematic North-South hydrogeological crosssection of the Kouris catchment:
1) – sedimentary rocks; 2) – Pillow Lavas; 3) – Gabbros and Sheeted Dykes; 4) – Ultramafic rocks; 5) piezometry level of groundwaters; 6) - fault zones.
Figure 1.7. Fractured plutonic rocks of the Kouris catchment (Gabbro).

Figure 1.8. Intrusive rocks of the Kouris catchment (Sheeted Dykes).
Figure 1.9. Volcanogenic rocks of the Kouris catchment (Pillow Lavas).

Figure 1.10. Highly altered gabbro in the Kouris catchment.
Figure 1.11. Sedimentary rocks of the Kouris catchments (marls).

Figure 1.12. River alluvium of the River Kouris.
Figure 1.13. The River Limnatis upstream.

Figure 1.14. The River Limnatis downstream after the rain.
Chapter 2

Groundwater resources
in the Kouris catchment (Cyprus):
data analysis and numerical
modelling

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ABSTRACT

The Kouris catchment in Cyprus is currently experiencing a scarcity of water resources due to the semi-arid climate across the southern part of the region, a series of dry years, and recent surface/subsurface water over-extraction. The catchment consists of the upper part of an ophiolitic complex in the North, which is considered a very significant aquifer for Cyprus, and an overlying sedimentary complex in the South, which has low water storage capacity.

Water balance calculations are conducted using a steady state groundwater model. The recharge rate was calculated to be between 12 and 16 % of the total annual rainfall. This agrees, with an estimate based on the mass balance of chloride. When the rate of extraction was increased to a value close to the present water demand, river baseflow was reduced from 25 to 18 Mm$^3$ per year. Other negative impacts were extreme drawdowns and drying up of springs.

KEYWORDS: ophiolites, water management, semi-arid climate, Cyprus, mountainous region, recharge.
2.1. INTRODUCTION

The Kouris catchment encompasses an area of 300 km² and extends from the southern side of the Troodos Massif of Cyprus to the Mediterranean Sea (Fig. 2.1). The topography is typical of a mountainous area with steep valleys, and surface elevations drop within 30 km from 2000 m in the North to sea level in the South (Fig. 2.2).

The River Kouris, which drains the catchment (Fig. 2.3), is the largest in Cyprus and has had an average annual streamflow of 36 Mm³/year during the last 30 years. The Kouris downstream reservoir (Kouris Dam) has the highest storage capacity in Cyprus (110 Mm³), but it has never been filled to the planned level and has stayed almost empty during the years 1998 - 2000. At the same time, a drop of groundwater levels in the vicinity of irrigation boreholes located in the upper part of the catchment was observed (Fig. 2.4). Water conflicts appearing in the catchment motivated the investigations reported in this paper.

The major groundwater resources are associated with the ophiolitic complex, which comprises harzburgites, gabbros, sheeted dykes, and pillow lavas. It is characterised by strong heterogeneity in hydraulic conductivity and porosity related to tectonic fracturing and hydrothermal alteration. The climate is characterised by very high spatial and temporal variability.

There are many scientific publications related to the geology of the Troodos Mountains (Gass et al. 1994; Malpas et al. 1990; van Everdingen 1995; Gillis and Sapp 1997; Varga et al. 1999). However, there are no published results related to groundwater occurrence in these ophiolites even though the hydrogeology of the ophiolites was studied in the 70s and 80s and described in several technical reports (Jacovides 1979; Afrodisis et al. 1986). In the 1980s, most groundwater studies in Cyprus were concerned with isotope investigations (¹⁸O, ²H, ³H, ¹⁴C) (Jacovides 1979; Verhagen et al. 1991). The area studied included the part of ophiolites of the Kouris catchment. Other studies of the sedimentary complex of Cyprus have focused on estimating recharge from precipitation (Kitching et al. 1980) and the practice of artificial recharge (Jacovides 1997). The LIFE project (Charalambides et al. 1998) assessed existing and potential groundwater contamination due to the asbestos mines in the north-west of the Kouris catchment. Within the LIFE project, water and rock samples from the igneous part of the Kouris catchment were analysed, and lineament analyses based on aerial photography were performed.
This paper presents an analysis of existing and new hydrogeological data, and combines them in a 2-D regional groundwater flow model. The method of model calibration, which is based on hydrogram separation and spring discharge, provides a unique set of hydrogeological parameters that will be required for more detailed future coupled surface/subsurface modelling. The focus of the present study is to estimate a range of the rates of recharge and analyse the likely impact of additional groundwater extraction. The results of the modelling are compared with estimates of the long-term average recharge derived independently using the chloride method. Finally, the sensitivity of the regional water balance to the different model inputs is evaluated, thereby pointing the way for future work.

2.2. DATA ANALYSIS

Precipitation
The long-term average annual precipitation for the Kouris catchment is 700 mm with 80% of it falling from November to March. Daily rainfall has been recorded at 18 meteorological stations since 1917 (Fig. 2.3); continuous daily records from splayed base rain gauges have been available at 9 stations for the last thirty to forty years.

Annual rainfall correlates well with surface elevation (average correlation coefficient for 1970 – 1994 is 0.93), and is highly variable over the years (Fig. 2.5). No obvious trends can be identified during the period from 1970 to 1994. The maximum multi-year differences are found at high altitudes and can reach 900 mm (station N270, hydrological years 1988/89 and 1989/90). Some dry periods (for example in 1972/73 and 1990/91) are noticeable at all stations (Fig. 2.5). Daily precipitation is highly variable in time and space and cannot be described using simple correlation analysis.

Evapotranspiration
Cyprus has a typical Mediterranean climate with mild winters, long, hot, and dry summers, and short autumn and spring seasons. Climatic data (temperature, pan evaporation, humidity, wind velocity, etc.) have been recorded at 4 meteorological stations since 1984 by the Meteorological Department of Cyprus (Fig. 2.3). Daily potential evapotranspiration rates
were calculated using a version of Penman’s equation with the net radiation term estimated from sunshine hours, using the software INSTAT (Stern et al. 1991, p. 49-51). The calculated mean annual potential evapotranspiration rate for the Kouris catchment for 1986-96 varies from 1060 mm to 1360 mm for the stations at different surface elevations. Within an annual cycle, the monthly potential evapotranspiration rate changes from less than 50 mm to more than 200 mm (Fig. 2.6). It is phase-shifted by half a year relative to monthly precipitation. Annual values stay nearly constant (with less than 10 % variability) over the period of observations (Fig. 2.6).

Surface flow
The Kouris catchment is drained by three rivers: Kouris, Kryos and Limnatis (Fig. 2.3). Daily streamflow data is available from 4 gauging stations, although data collection at each began in different years. Three of them are located in the lower part of the three rivers and close to the Kouris Dam (Fig. 2.3). All the stations are equipped with permanently installed water level recorders and portable current meters that allow the measurement of flow velocities at different levels of water flowing through the station. The accuracy of the flow measurements is 5 %-10 %, and is affected by the infrequency of flows, the steep gradient of riverbeds, and the movement of bed loads during floods.

Flow in the Kouris and Limnatis rivers persist throughout the year, consisting of spring water during the dry season, while the river Kryos is normally dry in summer. Daily streamflow reacts rapidly to rainfall – the peak of runoff occurs downstream only several hours after rainfall. The annual streamflow depends exponentially on the amount of precipitation (Fig. 2.7).

Geological setting
Cyprus and its immediate surroundings are segments of oceanic crust and mantle that are separated from the edge of one of the two major plates by rift-faulting, that developed possibly during the late Cretaceous (Malpas et al. 1990). The ophiolitic complex of the Troodos mountains, which is widely accepted as one of the best exposed on Earth, occupies an area of nearly 3000 km² in the south-western part of the island (Fig. 2.1). The plutonic complex is exposed in a central uplifted core and consists of tectonised harzburgites, dunites
and serpentinites, stratigraphically overlain by ultramafic gabbroic cumulates. Due to subsequent erosion, the original upward succession of plutonic rocks, sheeted dike complex and pillow lavas is arranged in an outward succession from centrally exposed plutonic rocks to the sheeted dike complex and peripheral pillow lavas.

The simplified geological map of the Kouris catchment (Fig. 2.8) shows the two main geological zones: the ophiolitic complex in the North and the overlying sedimentary complex in the South. The ophiolitic complex is characterised by strong lithologic heterogeneity and is composed of different units: the pillow lava, the sheeted dykes, the gabbros and the peridotites. Inside these units, the sheeted dyke complex is heterogeneous because it is composed of a series of dykes of different ages with different compositions. These rocks have been altered either by hydrothermal circulation soon after their formation in an active rift or by recent weathering in surface. Finally, they are all fractured at different scales. The contact between ophiolites and sedimentary rocks has a tectonic origin and has been investigated by a number of studies (Gas et al. 1994), but the magnitude and the angle of the displacement within the area are still unknown.

**Spring flow**

Springs are mainly associated with plutonic and intrusive rocks (gabbro, sheeted dykes) in the upper part of the region. In total, 64 springs have been mapped within the catchment (Fig. 2.3) and 14 of the largest have been monitored monthly since 1986 by the Water Development Department. However, in the area with the most intensive groundwater extraction, spring discharge has not been measured. In addition, there are many small springs in the upper Limnatis subcatchment that are not mapped. The instantaneous discharge of one of the largest monitored springs (Mavromata) varies from 0.65 l/s to 215 l/s. Considering only the monitored springs, we calculate minimum total spring discharges for the area ranging between 2 and 5 Mm³ per year, depending upon the year. The ratio between this amount and measured total streamflow increases in dry years and can reach 20 %.
Groundwater quality and occurrence

Generally, the catchment can be divided into two main aquifers: a sedimentary aquifer (chalks, marls, calcarenites) and an igneous rock aquifer (ophiolites) consisting of mantle, plutonic and intrusive sequences.

The ophiolitic complex was identified as a major aquifer only comparatively recently when in the 70’s drilling machines were imported into the island. The exploited groundwater resources lie in the fractured and altered rocks where the aquifer stays largely confined. Groundwater is generally of the CaMg-HCO₃ and Na-HCO₃ types with low to moderate salinities (300-600 mg/l), although cation and anion compositions vary considerably with local lithology (Afrodisis 1986). The groundwater quality is generally adequate for domestic supplies and irrigation. The sedimentary aquifer in the south, which is composed mainly of chalks and marls, is less permeable and the groundwater has a higher salinity.

Table 2.1 shows a synthesis of drilling record information related to water occurrence in the different lithologies. In general, the boreholes are drilled to the second or third encountered fractured zone or up to a maximum depth of 250 metres. The location of the boreholes is shown in Figure 2.8. Information from a total of 166 boreholes is available. Of these, 64 boreholes were unsuccessful in finding sufficient water, and the rest found water in different quantities. Usually, water inflow during drilling occurs in the fractured zones in gabbros and sheeted diabase dykes, in the basal group of pillow lavas, in the river gravel near the surface, or in fractures in the sedimentary rocks. The few available records suggest a low probability of water occurrence in the pillow lavas.

In summary, the sediments can be considered to be an aquifer with little water available (30 % drilling success in finding water), while drilling in the ophiolitic aquifer has a 93 % success rate. These two aquifers are separated by Upper and Lower Pillow Lavas.

³H content in groundwater

Values of the tritium contents in groundwaters provided by Jacovides (1979) and Verhagen et al. (1991) allowed us to estimate residence times in the different aquifers. The calculation is based on the piston flow model (Maloszewski & Zuber 1982), implemented in an EXCEL spreadsheet by K. Zoellmann (http://www.baug.ethz.ch/ihw/boxmodel_en.html), and uses the rainfall isotopic data for Cyprus provided by IAEA (http://isohis.iaea.org).
In the ophiolitic complex, the tritium content sampled in 14 springs in 1976 – 78 varied over a wide range (15 – 62 TU), which shows the complexity of the groundwater flow system in the fracture network. The calculated residence times are less than 5 years for 8 springs, 4 –10 years for 4 springs and around 20 years for 2 springs. During the year, the tritium contents vary significantly. In March-April, the contents suggest that rainfall from the last rainy season contributes significantly to net discharge, while in autumn the vast majority of the discharging groundwaters tend to be several years older. The 20-year-old water seen at 2 springs appears only in the dry season. Boreholes tapping water in different depths show an even higher variability in tritium content than seen in springs. Boreholes which are pumped heavily during dry seasons (typically for irrigation) sometimes eventually can even yield water that is older than 50 years, indicating extraction from old reservoirs. Such water cannot be considered renewable in the short term.

In the sedimentary complex, there is only one spring that has been sampled for tritium within the alluvium in the Kouris catchment. The data indicate that the spring discharges recent water from the last flood. Verhagen et al. (1991) report analysis of groundwater from 28 boreholes that take water from the sedimentary complex to the west of Larnaka (approximately 30 km to the east of the Kouris catchment). Our calculations for 26 of these boreholes indicate residence times of about 25 years (1 – 6 TU in 1981-86), which is relatively long. This could reflect both low recharge rates and low flow rates coming from the ophiolites.

Aquifer properties
Data from 76 variable rate pumping tests are available. The drawdowns were recorded only in the pumping boreholes (i.e. not in observation wells). The duration of the pumping was usually 2-3 days, and the subsequent recovery was monitored. Some tests were repeated in different seasons and years. In many cases, the pumping rate varied considerably during the tests, sometimes by a factor of 5-6. Often, the water level in the borehole coincided with the level of the pump itself (which made interpretation more complex). All pumping test data were re-interpreted with PC software (Sindalovski 1999) using the Theis solution superimposed in time for variable rates and recovery.

The resulting transmissivities are presented in Table 2.2. Each interpretation is assigned a measure of quality between 0 and 4. The highest quality of 4 indicates that
differences in values obtained from different tests or from pumping and recovery periods are all less than 20%. Similarly, 3 indicates differences between 20% and 200%. A quality factor of 2 indicates that only the recovery tests could be used, and 1 indicates that only the order of magnitude can be estimated. 0 indicates that the data cannot be interpreted. Overall, 15 values had grade 4, 24 values had grade 3, 15 values had grade 2, 12 values had grade 1, and 10 values had grade 0.

This data set is probably biased, as the boreholes, especially in gabbros and sheeted dykes, were drilled in zones with potentially high transmissivity. Other concerns relate to the possible non-applicability of the Theis model because of well-bore effects, presence of single fractures, double porosity, non-integer flow dimensions, unconfined situations, spatial heterogeneities, etc.

For the gabbros and sheeted dykes, taking only the 40 values graded "2" to "4", the distribution of transmissivity values is asymmetrical, with a high number of small values and only a few high values. It is approximately, but not perfectly, log-normal. The geometrical mean is 20 m²/day with minimum and maximum values of 2 m²/day and 703 m²/day. The variance of the decimal logarithm of the transmissivity is 0.36. The spatial distribution of these transmissivities as described by their variogram does not show any clear spatial structure. Nor do we observe a correlation between transmissivity and the lineament locations provided by Charalambides et al. (1998). Analysis of the geological logs from the boreholes shows that higher transmissivity values tend to be associated with fractured zones located at depths of several tens of metres and having widths, in certain cases, from 20 to more than 100 metres. For the harzburgites, serpentinites, and pillow lavas, no pumping tests data are available within the catchment.

For the sediments, only 6 transmissivity values are suitable for interpretation (quality level 2 and 3). These vary from 2 to 23 m²/day with a geometrical mean of 9 m²/day. For the river bed alluvium, only two tests were conducted. These show high transmissivity values but with different quality: 32 m²/day (quality level 3) to 250 m²/day (quality level 1).

The thicknesses of the main aquifers is not known directly but can be inferred from geological observations, and by analogy with other mountainous groundwater flow systems in fractured rocks. The total thickness of the sheeted dykes is around 500 m and the thickness of the gabbros is up to kilometres (Gass et al. 1994). However, the hydraulic conductivity due to fracturing decreases with depth (van Everdingen 1995). In the Alps, fractured crystalline rocks can be conductive even at depths exceeding a kilometre (Ofterdinger et al. 2002). Therefore, we envision the ophiolitic system to be a very thick body with a rather high degree
of fracturing for the first 100 to 200 metres and a progressive decrease of conductivity up to a depth of several kilometres. We can then expect very deep groundwater flow systems interacting with faster shallow systems. The pillow lavas formation has a thickness varying between 50 and 200 metres. The thickness of the sediments varies between a few tens of metres in the north to more than 600 m in the vicinity of Lofou village, 9 km north-west of the Kouris Dam (Gass et al. 1994).

**Piezometric surface**

Measurements of the groundwater table were conducted by the Water Development Department of Cyprus and the Engineering Geology unit of Swiss Federal Institute of Technology as follows: a) monthly in 17 boreholes since 1984, b) continuously in 1 borehole since July 1992 and in 4 boreholes since spring 2000; c) during one non-pumping period in 64 boreholes in March 2000.

The piezometric surface (Fig. 2.9) was constructed for March 2000 using the interpolated field observations of depths to the water table, a Digital Elevation Model with spatial resolution of 25 m X 25 m (Hall 1998), and a survey of the springs. Figure 2.9 shows the piezometric surface of the catchment which follows the topography (Fig. 2.2) in a considerably smoothed manner.

Variations in piezometric heads over many years range between several metres and 100 metres. Annual variations normally do not exceed 30 m, although higher values can occur probably because of pumping (e.g. in borehole 109/77, where piezometric head variations of 60 metres were measured in 1993). The lowest piezometric heads occur in September-October, while the highest occur in February-March.

**Surface/groundwater relation**

Daily streamflow measurements in the lower part of the 3 rivers were used for hydrogram separation. We applied both the fixed-interval and the sliding-interval methods (Pettyjohn and Henning 1979; Sloto and Crouse 1996). The duration of surface runoff was estimated from the empirical relation (Linsley et al. 1982):

\[ N = 0.83 A^{0.2} \]
where $N$ is the number of days after which surface runoff ceases, and $A$ is the drainage area in km$^2$. Drainage areas of 120 km$^2$ (Limnatis), 100 km$^2$ (Kouris) and 70 km$^2$ (Kryos) were used. The results of hydrogram separation for the years 1988/89 and 1989/1990, presented in Table 2.3, indicate that baseflow represents 60 to 75% of total streamflow.

**Evaluation of recharge rate by the chloride method**

We measured the chloride content in rainfall for the period September 2000 to November 2001 at 7 locations. Precipitation samples were combined either monthly or by rainfall events. The data of 29 rainfall samples indicated a homogeneous spatial and temporal distribution of chloride content, with an average value of 3.8 mg/l.

The direct recharge, $R$ [m$^3$/year], was determined from the following chloride mass balance equation,

$$R = \frac{P \times C_p - S \times C_s}{C_r}.$$  \hspace{1cm} (2.1)

Here $P$ [m$^3$/year] and $S$ [m$^3$/year] represent, respectively, the long term averaged precipitation and surface run-off rates, while $C_p$, $C_s$, and $C_r$ [mg/l] represent the chloride concentrations in precipitation, surface run-off and recharge. It is assumed that the amounts of chloride leaving the system via evaporation and entering the system via rock dissolution are negligible.

Hydrogram separations by the fixed-interval and the sliding-interval methods (Pettyjohn and Henning 1979; Sloto and Crouse 1996) show that the surface runoff, $S$, is 30% of streamflow on average for the years 1986-1996. In addition, for this period the average annual streamflow versus rainfall is 0.16. Consequently we have $S = 0.16 \times 0.3 \times P = 0.05 P$. The chloride concentration in surface run-off is generally close to the concentration in precipitation ($C_s = C_p$); however, this assumption leads to slight over-estimation of recharge for first rains after dry seasons when the surface run-off contains dissolved solids. Using these results in Eq. 1, together with the assumption that the concentration in recharge is close to the concentration in groundwater, $C_r \approx C_g$ [mg/l], gives:

$$\frac{R}{P} = 0.95 \frac{C_p}{C_g}.$$  \hspace{1cm} (2.2)

We selected 70 sampling locations in the ophiolitic complex and 4 locations in the sedimentary complex chosen so that the samples were far enough from the rivers to avoid
mixing with surface water from floods. The measured concentrations were between 7 and 26 mg/l in the ophiolitic complex, and between 23 and 30 mg/l in the sedimentary complex. This leads to local recharge rates ranging between 14 and 52 % in the ophiolites and between 12 and 16 % in the sediments. An estimate of the average chloride concentration for the catchment is calculated by weighing the averaged values within each of the lithologies by the relative surface areas they occupy in the catchment. The results indicate $C_g = 21$ mg/l and an average recharge rate of 17 %.

**Land use and water extraction**

Water extracted upstream of the Kouris Dam is used both for irrigation (mostly) and for domestic needs. The irrigated surface is approximately 17 km$^2$ and is covered by deciduous trees, vineyards and to a lesser extent by seasonal crops. The exact amount of present groundwater extraction is unknown. However, water demand upstream of the dam, estimated by the Water Development Department based on types of crops, irrigation area and number of inhabitants, was more than 10 Mm$^3$ per year in 2000 (Fig. 2.10).

Water demand is satisfied either by stream water, diverted from rivers or collected at the local dams, or by groundwater, pumped from boreholes or diverted from springs. Stream water is normally collected during the whole year and applied in summer for irrigation, while groundwater is extracted during summer and autumn. A rough estimate of actual ground and surface water extraction suggests that the most intensive pumping occurs in the upper parts of Limnatis and Kouris subcatchments.

**2.3. GROUNDWATER FLOW MODELLING**

Despite all the evidence that the aquifer is fractured, we modelled it as a continuous porous medium. This approach is reasonably correct for a regional study of water balance (see NRC 1996 for a review). All the heterogeneities due to the presence of a dense network of fractures and dykes which have a scale of a few metres in width and between tens and hundreds metres in length are averaged in an equivalent isotropic transmissivity which is constant at the scale of the geological formation (several kilometres). In practice, the groundwater flow was simulated in steady-state with a finite-difference technique using the code MODFLOW.
(Harbaugh and McDonald 1996) with the PMWIN interface (Chiang and Kinzelbach 2001). The hydrological year 1988/1989 (01.10.88 – 30.09.89) was chosen because the rainfall in that year was close to the long-term average. Another reason is that human impact on groundwater was still small at that time, while it seems have been very high since 1995.

Model Configuration
The boundary of the surface catchment was derived from the Digital Elevation Model (Hall 1998) and assumed, in the first step, to coincide with the boundaries of the underground catchment. In the second step this assumption was tested. The catchment was modelled as a confined aquifer with recharge imposed on the top. The area was discretized as one layer with a regular grid (250 m x 250 m, 142 rows by 86 columns). Figure 2.11 shows the grid and the location of the different boundary conditions. The boundaries are taken as impervious except along the Mediterranean coast where a fixed-head boundary condition, h = 0 m, was imposed. The rivers Kryos, Kouris and Limnatis were incorporated into the model using the River Package of MODFLOW. The elevations of the river bottom were obtained from interpolation of the Digital Elevation Model; the water levels in the rivers were specified to be 0.5 m higher than the surface of the river bed; the hydraulic conductances of the riverbed was determined by model calibration. Forty-two spring areas were incorporated into the model with the Drain Package of MODFLOW. The elevations of the drains were derived from the Digital Elevation Model and modified during groundwater model calibration. The drain hydraulic conductances were determined during the calibration procedure. Forty zones of groundwater pumping were imposed with constant pumping rates equal to the estimated annual extraction amounts for 1988/89.

Recharge
Four zones of recharge were defined (Fig. 2.12a) based on the geology and on the assumption that recharge is controlled by precipitation and evapotranspiration, which have correspondingly positive and negative correlations with surface elevation.

The total regional rainfall for the modelled year was calculated with the ISATIS package (http://www.geovariances.fr). The annual rainfall from 9 meteorological stations was interpolated by linear kriging inside the catchment boundaries on a 50 m by 50 m grid (Fig.
2.3). The total rainfall was the sum of the raster values multiplied by the area of the grid cell and was 241 Mm$^3$.

Two variants of recharge were modelled. Variant 1 was the conservative one for water management (minimum recharge) and assumed that all recharged water left the area via baseflow upstream of the Kouris dam and springs discharged into the rivers. No water was presumed to leave the model area via underground flow, recharge spent for increasing groundwater storage was neglected, and spring flow was included in river baseflow. Hence, total recharge for the four zones was adjusted to 30 Mm$^3$ to agree with the results of hydrograph separation for the three rivers (Table 2.3). This amount equalled to 12.5 % of total annual rainfall, and was in agreement with estimates from other studies that suggest that recharge in the range of 10 % - 15 % of the total rainfall for some areas on the coast (Jacovides 1982; Schmidt et al. 1988). Higher values probably apply to the ophiolitic complex, as it is at a higher surface elevation, and this was suggested by the chloride balance calculation. The total recharge was distributed over the 4 zones (Fig. 2.12a) according to the assumption that recharge increases with altitude. The largest differences in recharge rates were expected between Zones C and D and were caused both by differences in altitude and the relatively small infiltration capacity of sedimentary rock compared to ophiolites.

Variant 2 was the scenario with high input recharge. It was developed to test the sensitivity of the model to the recharge rate and to check what could be the maximum reasonable recharge rate. According to the general water balance for Cyprus (Omorphos et al. 1996), 80 % of rainfall is evaporated, while the remaining 20 % constitutes the water resources that are available for recharge and runoff. In Variant 2, 20 % of the annual rainfall (i.e. about 48 Mm$^3$ for 1988/1989) was used as recharge.

**Transmissivity**

Five zones of transmissivity were distinguished according to regional geology and pumping test results (Fig. 2.12b). These were: Zone 1 - mantle rocks (harzburgites, serpentinites, dunites); Zone 2 - plutonic and intrusive rocks (sheeted dykes and gabbros); Zone 3 - volcanogenic rocks (pillow lavas); Zone 4 - sedimentary rocks (chalks, marls, calcarenites); Zone 5 - sedimentary rocks (alluvium in the river beds).

For zones 2 and 5, the initial values for the regional transmissivities were estimated by multiplying the percentage of successful boreholes drilled within a lithology by the...
the geometrical mean of the transmissivities obtained from pumping test interpretation. The resulting values were \( T = 0.93 \times 20 = 19 \text{ m}^2/\text{day} \) for the gabbros and sheeted dykes (Zone 2), and \( T = 3 \text{ m}^2/\text{day} \) for the sediments (Zone 4). The transmissivity of Zone 1 (mantle rock) must be lower than that of Zone 2 because the rocks are more massive, while the transmissivity of the alluvium (Zone 5) was expected to be higher than that of the surroundings (Zone 4). Thus, 9 \text{ m}^2/\text{day}, and 6 \text{ m}^2/\text{day} \) were used as initial values for zones 1 and 5. In the case of Zone 3 (pillow lavas), 7 out of 10 boreholes did not reach the water table, and one pumping test conducted outside the Kouris catchment gave a moderate transmissivity value (10 \text{ m}^2/\text{day}). Thus, an initial value of 1 \text{ m}^2/\text{day} \) was used for Zone 3. Note that fractures are expected in pillow lavas and therefore the regional transmissivity could be higher in this lithology.

**Model Calibration**

The model was calibrated manually by trial and error using the following procedure. First, the transmissivity of Zone 1 was adjusted such that the variance of the difference between measured and calculated heads was a minimum. Then, the same was done successively for the transmissivities of zones 2, 3, 4 and 5. When this was achieved, the values of the recharge rates were adjusted within the different zones to again minimise the variance of the head differences, subject to the constraints that the total recharge remained constant and that the zones at higher altitudes had higher recharge rates. The next step was to adjust the values of the river bed conductances and drain altitudes so as to match the discharge rates in rivers and springs with their measured values. Finally, the variance of the head differences was checked, and all the steps repeated in sequence until an acceptable variance was obtained. Three loops were required until the expected accuracy was achieved.

An important point was the criteria to decide when a reasonable fit was obtained. We arbitrarily considered the errors in heads below 100 m to be reasonable. Our arguments were first, that the topography is steep in the upper part of the catchment (local slopes between 15 to 30 \%, sometimes up to 40 \%). Thus, differences in surface elevations can reach 100 m within the distance of 250 m (the size of a model cell), which brings the uncertainty equal to 100 metres in piezometric heads, calculated from depths of the water table and surface elevations. Second, for the model calibration we used piezometric data from different years and 100 metres is considered to be the maximum of long-term variations of piezometric heads. Thus, attempting to increase model accuracy necessarily leads to a reduction in the
quantity of data used in the calibration. Finally, 100 m represents only 6 - 7 % of the difference between the lowest and highest piezometric heads of the area.

The resulting set of calibrated parameters is listed in Table 2.4. With the exception of those for the pillow lavas, transmissivity values were modified only slightly during the calibration procedure. The observed and simulated discharges agree to within 92 %, and the corresponding scatter diagram of heads for Variant 1 is presented in Figure 2.13. Table 2.5 lists estimated and simulated river and spring discharges for Variant 1, and shows that there is satisfactory agreement between modelled river discharges upstream of the 3 river gauges and the baseflows obtained from hydrogram separation. When it was assumed that the springs were flowing to the rivers, the resulting simulated baseflow differed from the total observed baseflow by 4 % to 8 % (Table 2.5) depending on the method used for hydrogram separation.

The analysis of the spatial distribution of simulation errors indicates that no zones occur with systematically positive or negative errors. There are small simulation errors in the upper part of the Kryos subcatchment, while in the upper part of the Limnatis subcatchment high positive and negative errors often appear close to each other. This could be the result of either strong rock heterogeneity in the north-eastern part of the catchment, or it could indicate the depletion of a compartmentalised groundwater reservoir due to over pumping.

The uncertainties in spring locations, elevations and total discharges are large, and so, could be taken into account only approximately. Another problem in spring simulation occurred because hydraulic gradients controlling spring discharge are normally influenced by local parameters and hydrodynamic head distributions, which are neither resolved by the field information available nor by the model grid. In some springs, the actual discharge was unknown. In those cases, any computed discharge was counted as a satisfactory result. In the absence of other information, a large value of 10,000 m²/day was assigned to drain hydraulic conductance, and the elevation of drain was set to the value derived from the Digital Elevation Model. Usually the latter value had to be reduced by up to 100 m to obtain reasonable discharge rates.

Model results

The resulting estimates of the total water balance of the Kouris catchment for 1988/1989 for variants 1 and 2 are presented in Table 2.6. The simulated and observed heads agree with the satisfactory accuracy of the estimate (Fig. 2.13). Figure 2.14 shows the simulated piezometric
Waterbalance calculations for different regions show that the groundwater flow from the gabbros and sheeted dykes to the sediments through the pillow lavas is high: about 6-7 Mm$^3$/year. Later, most of this water is discharged as river leakage into the upper part of the sediments and pillow lavas, and to a lesser extent, through springs in the sedimentary rocks. Groundwater outflow to the Mediterranean Sea is the smallest discharge component. Variant 2 also gives satisfactory simulated heads, with the exception of 4 boreholes where the errors are slightly larger than 100 m. Thus, a calibration based only on heads would not be unique since two calibrated sets of parameters could result in completely different water management scenarios. However, Variant 2 can be discounted for other reasons: the predicted total river leakage is too high (Table 2.6), the ratios between leakage fluxes in the 3 rivers are incorrect (Table 2.6), the transmissivity of 30 m$^2$/day of Zone 2 is unreasonably high, and spring elevations must be lowered more than is acceptable in order to obtain the correct discharges. A third variant with recharge equal to 16 % of rainfall gives a reasonable calibrated transmissivity for Zone 2 of 20 m$^2$/day and an acceptable water balance. Thus, within the assumptions of this model, the annual actual recharge for 1988/89 is estimated to be between 12 % and 16 % of total rainfall.

The location of the groundwater catchment boundaries is uncertain, since they might not coincide with their surface expressions. This uncertainty is highest in the sediments in the south where the surface catchment becomes very narrow. In order to check the model sensitivity to this parameter, the model boundaries were extended significantly in the sediments to construct Variant 4 (Fig. 2.15). All other parameters were kept the same as in Variant 1. The resulting water balance, presented in Table 2.7, shows that the groundwater flow to the sea remains less than 2% of the total outflow and that the other components of the water balance are not changed considerably.

The impact of groundwater extraction in 1988/89, predicted by model Variant 1, is very limited. The average water levels are different from those of the model run without pumping by only 10 metres. A further modelling variant (Variant 5) assumes that all extraction rates were higher by a factor 5 (i.e. 8.77 Mm$^3$/year in total), which approaches the present level of water demand. Again, all other parameter values are kept identical to those in Variant 1. The results of this simulation are shown in Figure 2.16 and Table 2.7. Increasing groundwater extraction reduces mainly the baseflow in the upstream parts of the Kouris and Limnatis subcatchments from 25 to 18 Mm$^3$/year. This would represent a 7 Mm$^3$/year deficit of inflow for the Kouris dam. Of the 42 springs, the 14 closest to the pumping boreholes (33 %) dry up. The total spring discharge is reduced by 36 % compared to variant 1. Additional
drawdown occurs only in the upstream of the Kouris and Limnatis subcatchments over an area of about 50 km². In this area, drawdowns increased by 30 metres on average and reached 180 m around pumping boreholes.

2.4. DISCUSSION

Under the current state of knowledge of groundwater in the Kouris catchment, our model yields estimates for the recharge rate of between 12 and 16 % of annual rainfall for 1988/89. This is in agreement with the chloride mass balance data and is relatively insensitive to the geometry of the catchment boundaries in the sediments.

The model can be used to forecast possible impacts of increased exploitation. This was demonstrated in Variant 5 and could be extended to even higher exploitation rates. However, the model is not able to evaluate the time scale on which the baseflow will react to such increased exploitation. Such an investigation would require a transient model to be constructed. In addition, the model is not able to forecast the effect of potential variations in precipitation and temperature due to climatic change, as the recharge is not calculated from these meteorological data but calibrated. Furthermore, because of its steady-state assumptions, the model cannot be used to forecast some more subtle effects such as the shift of precipitation from one season to another, as observed for example by Kypris (1995) in Cyprus. Overcoming these limitations requires a more sophisticated transient model that couples groundwater and surface flow, and allows precipitation, rather than recharge, to be input directly.

The constraints applied in the present model are the head data, the baseflow estimated from hydrogram separations, the transmissivities estimated from pumping tests, and the spring discharges. The accuracy of the model is therefore directly linked to the accuracy of these data. The results obtained show that the most important constraint is the baseflow. Uncertainty in its value arises from errors in streamflow measurements and errors due to the estimation methods. The estimation methods that can be used include: tracer methods (Stichler et al. 1986), field methods requiring boreholes in the alluvial aquifer or seepage meter installations (Cruickshank et al. 1988; Holko 1995), and methods using only streamflow observations (Nathan and McMahon 1990). The curve-fitting techniques used for the Kouris catchment (Sloto and Crouse 1996) belongs to the last group of methods. The most
reliable hydrogram separations can be obtained by a combination of several methods. Unfortunately, at present only streamflow observations are available in the catchment. Thus, baseflow was calculated using only one type of method. Nevertheless, curve-fitting (in our case fixed-interval and sliding interval methods) is widely accepted by hydrologists for hydrograph separation even in mountainous regions (e.g. Lindsey et al. 1997; Sinclair and Pitz 1999). Certainly, since the absolute value of the baseflow is the most important constraint, environmental tracer methods should be applied in the future.

Improved estimates for spring locations and elevations would also be useful for constructing the piezometric head field. However, better measurements of spring discharge will not significantly improve total water balance calculations.

The calibrated transmissivity of the gabbro/sheeted dykes complex varies within a small range (16 m²/day – 20 m²/day) and coincides with the geometrical mean of all pumping test interpretation results. The modelled piezometric map is highly sensitive to the transmissivity of mantle rock, which according to the model calibration can vary over a very small range of 4.5 – 5.5 m²/day. These results are comparable, at least in terms of contrast between the hydraulic conductivities of different lithologies, with the pumping test estimates of Dewandel et al. (2002) for the ophiolitic rocks of Oman.

A striking feature of the model is the groundwater flow that is predicted to occur from the gabbros and sheeted dykes to the sediments through the pillow lavas. Additional investigations should be conducted in order to check the validity of this calculation. This could be achieved by model calibration with additional streamflow measurements that would give a more detailed river leakage distribution, and could be tested with a comparative study of isotopic signatures of groundwater on both sides of the contact.

As we have already discussed, the accuracy of the head values is very limited in our model. This is not a problem for the regional water balance, but it would not be acceptable for a groundwater management model. If such a model was constructed, it would obviously have to be focused on the gabbros and sheeted dykes area in the upper Kouris and Limnatis subcatchments where groundwater extraction is the highest.

2.5. CONCLUSIONS

The Kouris catchment is a Mediterranean and partially (in the southern part) semi-arid area with scarce water resources and where water conflicts are increasing dramatically. The area has suffered a sustained long-term reduction of surface/ground water storage. The area is
characterised by high heterogeneity of rock properties, major evapotranspiration losses during the dry season, and strong coupling of surface and ground water.

Groundwater flow modelling of the area for the year 1988/89 indicated that recharge was between 12% and 16% of the total annual rainfall, which amounted to 100-150 mm for that year. Specifying recharge rates of less than 12% led to insufficient water supply for the estimated baseflow, whereas recharge of 20% already produced unacceptable changes in the predicted ratio between flows of the 3 rivers of the catchment. The chloride method gives an estimated long-term average recharge equal to 17% of rainfall, which is in agreement with the model.

If the current water demands were solely satisfied by groundwater, very high drawdowns in some locations of the catchment would likely occur. It is also very probable that spring discharge would be severely reduced, perhaps by as much as a factor two, resulting in the drying up of many springs. Furthermore, the natural discharge of the aquifer into the rivers could drop from 25 to 18 Mm$^3$ per year, thereby reducing significantly the annual inflow into the Kouris dam.

Whilst the results have serious implications for groundwater in Cyprus, they must be considered provisional owing to the many uncertainties in the model. To improve the accuracy of the simulation at the regional scale, additional streamflow measurements combined with isotopes investigations should be conducted and the flow through pillow lavas should be estimated. The development of a transient model could allow the transition time between the natural and the perturbed situation to be assessed. Finally, substantially improving forecasts for groundwater management requires that the area in the upstream of Kouris and Limnatis Rivers be further investigated.

ACKNOWLEDGMENTS

The authors thank W. Kinzelbach, F. Leuenberger, C. Leduc, K. Evans and D. Porcelli for their comments and suggestions. We are also grateful to A. Moll, M. Steiner, U. Jorin and S. Oerli for essential contributions during their diploma studies. The work was financially supported by the European Commission: Research Project BBW 97.0621 and by the Swiss Federal Institute of Technology.
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Seite Leer / Blank leaf
<table>
<thead>
<tr>
<th>Rock type</th>
<th>Boreholes with available records (since 1976)</th>
<th>Successful</th>
<th>Unsuccessful</th>
<th>Percentage successful boreholes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Plutonic and intrusive sequence (gabbro and sheeted dykes)</td>
<td>83</td>
<td>77</td>
<td>6</td>
<td>93 %</td>
</tr>
<tr>
<td>Volcanic sequence (pillow lavas)</td>
<td>10</td>
<td>3 (basal group)</td>
<td>7 (6: upper and lower pillow lavas, 1: basal group)</td>
<td>30 %</td>
</tr>
<tr>
<td>Sedimentary rock (chalks, marls, calcarenites, alluvium)</td>
<td>73</td>
<td>22</td>
<td>51</td>
<td>30 %</td>
</tr>
<tr>
<td>Total within the area</td>
<td>166</td>
<td>102</td>
<td>64</td>
<td>61 %</td>
</tr>
</tbody>
</table>
Table 2.2. Results of pumping test interpretation. Coordinates are in UTM Grid 32, T = Transmissivity in m²/day, q indicates the quality of the data (see text for explanation), Bh N = serial number of a borehole.

<table>
<thead>
<tr>
<th>Bh N</th>
<th>North</th>
<th>East</th>
<th>T</th>
<th>q</th>
<th>Type of rock</th>
</tr>
</thead>
<tbody>
<tr>
<td>71/92</td>
<td>484986 3851071</td>
<td>250</td>
<td>1</td>
<td>alluvium/sediments</td>
<td></td>
</tr>
<tr>
<td>34/96</td>
<td>484910 3850941</td>
<td>32</td>
<td>3</td>
<td>alluvium/sediments</td>
<td></td>
</tr>
<tr>
<td>156/85</td>
<td>495210 3857640</td>
<td>50</td>
<td>3</td>
<td>dykes/gabbro</td>
<td></td>
</tr>
<tr>
<td>42/86</td>
<td>495468 3857706</td>
<td>10</td>
<td>4</td>
<td>Dykes</td>
<td></td>
</tr>
<tr>
<td>128/87</td>
<td>492097 3861336</td>
<td>153</td>
<td>2</td>
<td>dykes/gabbro</td>
<td></td>
</tr>
<tr>
<td>285/89</td>
<td>497520 3861525</td>
<td>7</td>
<td>1</td>
<td>dykes/gabbro</td>
<td></td>
</tr>
<tr>
<td>46/92</td>
<td>486130 3859354</td>
<td>20</td>
<td>1</td>
<td>dykes/gabbro</td>
<td></td>
</tr>
<tr>
<td>64/76</td>
<td>502711 3861596</td>
<td>46</td>
<td>2</td>
<td>dykes/gabbro</td>
<td></td>
</tr>
<tr>
<td>66/76</td>
<td>500603 3862240</td>
<td>27</td>
<td>0</td>
<td>dykes/gabbro</td>
<td></td>
</tr>
<tr>
<td>69/79/80</td>
<td>489675 3862076</td>
<td>100</td>
<td>0</td>
<td>dykes/gabbro</td>
<td></td>
</tr>
<tr>
<td>68/69</td>
<td>503406 3861816</td>
<td>4</td>
<td>4</td>
<td>dykes/gabbro</td>
<td></td>
</tr>
<tr>
<td>101/82</td>
<td>491870 3859785</td>
<td>25</td>
<td>3</td>
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<td></td>
</tr>
<tr>
<td>104/88</td>
<td>497503 3862897</td>
<td>7</td>
<td>0</td>
<td>gabbro</td>
<td></td>
</tr>
<tr>
<td>105/76</td>
<td>503661 3862663</td>
<td>70</td>
<td>2</td>
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<td></td>
</tr>
<tr>
<td>118/88</td>
<td>502481 3857830</td>
<td>25</td>
<td>4</td>
<td>gabbro</td>
<td></td>
</tr>
<tr>
<td>124/90</td>
<td>488164 3859812</td>
<td>10</td>
<td>4</td>
<td>gabbro</td>
<td></td>
</tr>
<tr>
<td>133/90</td>
<td>488180 3867000</td>
<td>10</td>
<td>1</td>
<td>gabbro</td>
<td></td>
</tr>
<tr>
<td>143/90</td>
<td>503284 3862413</td>
<td>5</td>
<td>3</td>
<td>gabbro</td>
<td></td>
</tr>
<tr>
<td>158/87</td>
<td>499123 3863412</td>
<td>90</td>
<td>3</td>
<td>gabbro</td>
<td></td>
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<tr>
<td>274/89</td>
<td>498155 3864600</td>
<td>160</td>
<td>0</td>
<td>gabbro</td>
<td></td>
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<tr>
<td>30/74</td>
<td>486886 3860430</td>
<td>215</td>
<td>0</td>
<td>gabbro</td>
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</tr>
<tr>
<td>32/76</td>
<td>496618 3862619</td>
<td>3</td>
<td>2</td>
<td>gabbro</td>
<td></td>
</tr>
<tr>
<td>32/96</td>
<td>494506 3863212</td>
<td>7</td>
<td>3</td>
<td>gabbro</td>
<td></td>
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<tr>
<td>33/97</td>
<td>494096 3862994</td>
<td>63</td>
<td>3</td>
<td>gabbro</td>
<td></td>
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<tr>
<td>40/96</td>
<td>493885 3868220</td>
<td>8</td>
<td>4</td>
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<td></td>
</tr>
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<td>41/95</td>
<td>490550 3860800</td>
<td>25</td>
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<td></td>
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<tr>
<td>46/97</td>
<td>494530 3863752</td>
<td>77</td>
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<td></td>
</tr>
<tr>
<td>53/76</td>
<td>497588 3860394</td>
<td>703</td>
<td>2</td>
<td>gabbro</td>
<td></td>
</tr>
<tr>
<td>53/77</td>
<td>495973 3856571</td>
<td>9</td>
<td>3</td>
<td>gabbro/serpentinite</td>
<td></td>
</tr>
<tr>
<td>55/96</td>
<td>499475 3864667</td>
<td>11</td>
<td>3</td>
<td>gabbro</td>
<td></td>
</tr>
<tr>
<td>56/97</td>
<td>495677 3867864</td>
<td>4</td>
<td>4</td>
<td>gabbro</td>
<td></td>
</tr>
<tr>
<td>57/76</td>
<td>498485 3864190</td>
<td>2</td>
<td>2</td>
<td>gabbro</td>
<td></td>
</tr>
<tr>
<td>58/93</td>
<td>498350 3862635</td>
<td>40</td>
<td>4</td>
<td>gabbro</td>
<td></td>
</tr>
<tr>
<td>58/96</td>
<td>501018 3864592</td>
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<td>gabbro</td>
<td></td>
</tr>
<tr>
<td>61/97</td>
<td>500718 3862327</td>
<td>9</td>
<td>4</td>
<td>gabbro</td>
<td></td>
</tr>
<tr>
<td>62/76</td>
<td>498240 3866501</td>
<td>10</td>
<td>1</td>
<td>gabbro</td>
<td></td>
</tr>
<tr>
<td>63/76</td>
<td>500943 3864350</td>
<td>100</td>
<td>0</td>
<td>gabbro</td>
<td></td>
</tr>
<tr>
<td>63/93</td>
<td>499790 3862830</td>
<td>36</td>
<td>3</td>
<td>gabbro</td>
<td></td>
</tr>
</tbody>
</table>
Table 2.3. Separated hydrographs for the Kouris catchment. Rates are in Mm$^3$ per year

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Streamflow</td>
<td>5.5</td>
<td>1.3</td>
<td>18.1</td>
<td>9.1</td>
</tr>
<tr>
<td>Baseflow from hydrograph separation for 1988/89</td>
<td>20.3</td>
<td>8.4</td>
<td>43.9</td>
<td>18.8</td>
</tr>
<tr>
<td>Fixed-interval method (% of streamflow)</td>
<td>4 (73%)</td>
<td>0.79 (61%)</td>
<td>12.9 (71%)</td>
<td>6.9 (76%)</td>
</tr>
<tr>
<td>Sliding-interval method (% of streamflow)</td>
<td>3.9 (71%)</td>
<td>0.77 (59%)</td>
<td>12.5 (69%)</td>
<td>6.8 (75%)</td>
</tr>
</tbody>
</table>

Table 2.4. Calibrated set of parameters for Variant 1 and Variant 2

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value for Variant 1</th>
<th>Value for Variant 2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Transmissivity of zone 1</td>
<td>4.5 m$^3$/day</td>
<td>6 m$^3$/day</td>
</tr>
<tr>
<td>Transmissivity of zone 2</td>
<td>16 m$^3$/day</td>
<td>30 m$^3$/day</td>
</tr>
<tr>
<td>Transmissivity of zone 3</td>
<td>100 m$^3$/day</td>
<td>250 m$^3$/day</td>
</tr>
<tr>
<td>Transmissivity of zone 4</td>
<td>1 m$^3$/day</td>
<td>1.5 m$^3$/day</td>
</tr>
<tr>
<td>Transmissivity of zone 5</td>
<td>20 m$^3$/day</td>
<td>40 m$^3$/day</td>
</tr>
<tr>
<td>Recharge of zone A</td>
<td>248 mm/year</td>
<td>372 mm/year</td>
</tr>
<tr>
<td>Recharge of zone B</td>
<td>168 mm/year</td>
<td>292 mm/year</td>
</tr>
<tr>
<td>Recharge of zone C</td>
<td>146 mm/year</td>
<td>219 mm/year</td>
</tr>
<tr>
<td>Recharge of zone D</td>
<td>11 mm/year</td>
<td>16 mm/year</td>
</tr>
<tr>
<td>Hydraulic Conductance of the Riverbed (most of the rivers)</td>
<td>25 m$^2$/day</td>
<td>50 m$^2$/day</td>
</tr>
<tr>
<td>Hydraulic Conductance of the Riverbed (small parts upstream of Kryos and Kouris containing possibly only water from springs)</td>
<td>0.02 m$^2$/day, 0.2 m$^2$/day, 2 m$^2$/day</td>
<td>0.02 m$^2$/day, 0.2 m$^2$/day, 2 m$^2$/day</td>
</tr>
</tbody>
</table>

Table 2.5. Estimated and simulated (Variant 1) components of outflow upstream of the Kouris dam, in Mm$^3$ per year

<table>
<thead>
<tr>
<th>River subcatchment upstream of Kouris Dam</th>
<th>Baseflow from hydrograph separation for 1988/89</th>
<th>Modelled Discharge</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kryos</td>
<td>Fixed-interval method 4.0</td>
<td>Sliding-interval method 3.9</td>
</tr>
<tr>
<td>Kouris</td>
<td>12.9</td>
<td>12.5</td>
</tr>
<tr>
<td>Limnatis</td>
<td>13.0</td>
<td>12.3</td>
</tr>
<tr>
<td>Total discharge upstream the Kouris dam</td>
<td>29.9</td>
<td>28.7</td>
</tr>
</tbody>
</table>
Table 2.6. Water balance for the Kouris catchment, with all values in Mm³ per year for 1988/1989; Variants 1 and 2

<table>
<thead>
<tr>
<th>Inflow</th>
<th>Outflow</th>
<th>Simulated</th>
<th>Reference*</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Variant 1</td>
<td>Variant 2</td>
<td>Variant 1</td>
</tr>
<tr>
<td>River Leakage</td>
<td>0.8</td>
<td>1.7</td>
<td>0.1</td>
</tr>
<tr>
<td>Recharge</td>
<td>30.2</td>
<td>48.0</td>
<td>4.2</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>1.7</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>1.6</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>10.9</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>13.1</td>
</tr>
<tr>
<td></td>
<td>Sub-total</td>
<td></td>
<td>25.0</td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td></td>
<td>31.0</td>
</tr>
</tbody>
</table>

*Reference from Table 3. Note that for the variant with big recharge 12 simulated springs from 42 turned out to be dry.
**NA = Not available

Table 2.7. Water balance of the Kouris catchment with extended boundaries [all values in Mm³ per year]; Variants 4 and 5

<table>
<thead>
<tr>
<th>Inflow</th>
<th>Outflow</th>
<th>Variant 4</th>
<th>Variant 5</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Variant 4</td>
<td>Variant 5</td>
<td>Variant 4</td>
</tr>
<tr>
<td>River Leakage</td>
<td>0.7</td>
<td>0.8</td>
<td>0.6</td>
</tr>
<tr>
<td>Recharge</td>
<td>32.0</td>
<td>30.2</td>
<td>4.4</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>26.0</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>1.7</td>
</tr>
<tr>
<td></td>
<td>Total</td>
<td></td>
<td>32.7</td>
</tr>
</tbody>
</table>
Fig. 2.1. Simplified geological map of Cyprus showing the location of the Kouris catchment.

Fig. 2.2. Digital Elevation Model of the Kouris catchment.
Fig. 2.3. Left: map of the Kouris catchment; circles – springs areas, squares – numbered climatological stations equipped for calculation of potential evapotranspiration; triangles – river gauges with recent measurements; Right: Annual rainfall (1988-89) interpolated by linear kriging from 9 meteorological stations (black circles with numbers).

Fig. 2.4. Progressive increase in the depth of the ground water table in the extraction boreholes with serial numbers 53/76, 31/76, 57/76, 21/82 and 69/79 during non-pumping seasons (October to March).
Fig. 2.5. Monthly (left) and annual (right) rainfall in mm at 9 meteorological stations; numbers of meteorological stations are presented in the legend.

Fig. 2.6. Calculated monthly (left) and annual (right) potential evapotranspiration (Penman equation); numbers of meteorological stations are presented in the legends.

Fig. 2.7. Annual streamflow in the River Kryos as a function of annual weighted depth area precipitation across the Kryos subcatchment; the diamonds show values for the period of 1976-1996; the solid line is a calculated trend.
Fig. 2.8. Simplified geological map of the Kouris catchment; black circles show locations of boreholes drilled for irrigation purposes (some of them were unsuccessful).

Fig. 2.9. Piezometric map constructed from interpolated piezometric heads for March 2000.
Fig. 2.10. Water demand for the irrigation and domestic needs of the Kouris catchment.

Fig. 2.11. Map showing the location of boundaries for the modelling of the Kouris catchment.
Fig. 2.12. Left: Recharge zonation depending on the surface elevation; Right: Transmissivity zonation (Zone 1 - harzburgites, serpentinites; Zone 2 - gabbro, sheeted dykes; Zone 3 - pillow lavas; Zone 4 - sediments (chalks, marls, calcarenites); Zone 5 - alluvium).

Fig. 2.13. Scatter Diagram for observed and simulated heads in 71 boreholes, Variant 1; calculated variance is 2550 m$^2$. 
Fig. 2.14. Map showing contours of piezometric heads for the simulation of Variant 1.

Fig. 2.15. Map showing the location of extended catchment boundaries for the sensitivity analysis.
Fig. 2.16. Results of simulation with 8.77 Mm$^3$ annual groundwater pumping: Left - piezometric map; Right - differences to the basic variant (colour scale is in metres).
Chapter 3

Study of stable isotopes in the Kouris catchment (Cyprus) for the description of the regional groundwater flow

Anastasia Boronina, Werner Balderer, Philippe Renard & Willibald Stichler

Submitted to "Journal of Hydrology"
ABSTRACT

The stable isotopes of oxygen and hydrogen in groundwater and precipitation were integrated for description of groundwater flow in the Kouris catchment (Cyprus). The catchment consists of an ophiolitic complex in the North and sediments in the South. It is characterized by strong heterogeneity of an underground media and steep slopes of the topography.

The regression line, constructed on the data from 70 rainfall samples is described by equation: $\delta D = 6.6^*\delta^{18}O + 10.9$, showing evaporation during precipitation. The altitude gradients in precipitation were estimated to be $-2.2 \%/100 \text{ m}$ for $\delta D$ and $-0.34 \%/100 \text{ m}$ for $\delta^{18}O$. The stable isotope analysis of 234 groundwater samples reflected fractionation due to evaporation. The origin of groundwaters in the catchment was described, based on the regression equation between surface elevations and $\delta D$ contents for 33 selected springs of the ophiolitic complex. Additionally, $\delta D$ data were used for the calibration of the recharge rates of a steady state groundwater flow and transport model. The resulting calibrated total steady state recharge rate is 100-130 mm per year.

KEYWORDS: stable isotopes, ophiolites, semi-arid climate, Cyprus, groundwater, recharge.
3.1. INTRODUCTION

Water conflicts in Cyprus have risen dramatically within the last thirty years due to growing water demands and a partially semi-arid climate. A description of the hydrogeology and the calculation of groundwater balance become very urgent. Additionally these tasks are still challenging from scientific point of view because of the variability in time and space of climatic conditions, and because of the high heterogeneity of underground properties.

Stable water isotopes have been used for decades all over the world as powerful tool in understanding groundwater flow systems, especially for highly heterogeneous regions with a high role of evaporation. In Cyprus, stable isotopes analyses were done in 1976-79 by Jacovides (1979), the study area included a part of the Kouris catchment. Other investigations (Verhagen et al., 1991) focused on the description of stable isotopes in the Lefkara aquifer (chalks, marls), approximately 30 km to the East from the Kouris catchment.

The analysis of stable isotopes in precipitation was the subject of a great number of scientific investigations (Araguas et al., 2000; Gat, 2000; Gat, 1996; Dansgaard, 1964; Craig, 1965), our research delivers new results for the precipitation in Cyprus.

Although the “stable isotopes altitude effect” can be easily observed in precipitation, it is often different from the altitude effect in recharged water because of evaporation prior to- or during infiltration (Allison et al., 1983, Gonfiantini, 1986). In the present study, we show changes in the groundwater isotopic content due to evaporation and describe the altitude effect in recharge on the base of observations of several springs.

This paper presents application of new and published (Jacovides, 1979) stable isotopes analysis for the description of precipitation and groundwaters in the Kouris catchment (Cyprus). For the precipitation, the regression between $\delta$D and $\delta^{18}$O is established and isotopic variations due to different seasons and altitudes are analyzed. For the groundwaters, their origin and genesis are discussed based on their isotopic contents. In parallel, some tritium and chloride data, collected in 1998-2001, were used for the
description. Additionally, the steady state recharge rate for the groundwater numerical model (Boronina et al., 2003) is calibrated by deuterium data.

3.2. STUDY AREA

The detailed description of the area is given by Boronina et al. (2003). The Kouris catchment is bounded on the South by the Mediterranean Sea and it covers 300 km² on the southern side of the Troodos Massif (Fig. 3.1). Elevations range from the sea level to 2000 m within a distance of 30 km with local slopes up to 70 %.

Annual precipitation amounts increase from 300 mm along the coast to nearly 1200 mm in the Troodos mountains. Annual rainfall correlates with surface elevation (average correlation coefficient for 1970 – 1994 was 0.93), and is highly variable over the years. The rain falling from November to March is usually about 80% of the total yearly amount.

The calculated mean annual potential evapotranspiration for the catchment for 1986 - 96 varied from 1060 to 1360 mm per year for the stations at different surface elevations.

The basin is divided in two main geological zones: an ophiolitic complex in the North and an overlying sedimentary complex in the South. The ophiolites include ultramafic rocks, gabbros, sheeted dykes and pillow lavas, while the sediments are mainly chalks, marls, calcarenites and limestones (Fig. 3.2).

The ophiolitic rocks are highly heterogeneous because of the presence of different lithological units, which are fractured and altered at different scales. These rocks contain the major groundwater resources of Cyprus; the water is stored in the fractured and altered zones of harzburgites, dunites, gabbros and diabase dykes. The transmissivities of the gabbros and the diabase dykes vary from 2 m²/day and 703 m²/day with a geometrical mean of 20 m²/day according to the results of 40 pumping tests (Boronina et al., 2003).

Groundwater in the ultramafic rocks, gabbros and diabases is generally of CaMg-HCO₃ and Na-HCO₃ types with low to moderate salinities (200-600 mg/l), cation and anion compositions vary considerably with the local lithology. The Pillow Lavas are considered to be generally less permeable, although local zones of high conductivity may exist there. The amount of TDS are higher in Pillow Lavas, than in ultramafic, plutonic and intrusive rocks (500 mg/l to 950 mg/l) and within the anion content SO₄⁻ - plays the major role.
The sedimentary part consists of chalks, marls, calcarenites and limestones in the major part of the Kouris catchment (estimated average transmissivity 3 m²/day), and it consists of river alluvium (sand, gravel) in the Kryos, Kouris and Limnatis valleys. From the boreholes, drilled in this part for irrigation purpose, 70% were unsuccessful. The alluvial aquifer is narrow (sometimes less than 100 meters) and discontinuous, although it contains major amount of water for the southern part of the Kouris catchment. Groundwaters in the hard sedimentary rocks are generally moderately mineralised to brackish, the salinity varies between 600 and 1600 mg/l. They are of Na-Ca-HCO₃-SO₄, Na-Ca-SO₄-HCO₃ or Ca-Na-HCO₃-Cl types. Groundwaters in the alluvium aquifer are less mineralised (500-600 mg/l), of Ca-Mg-HCO₃, Mg-Ca-HCO₃ or Mg-Na-HCO₃ types.

Springs of the area originate mainly from ultramafic, plutonic and intrusive rocks (harzburgites, gabbro, sheeted dykes) in the upper part of the catchment. Only a few springs are located in the sediments, mainly discharging water in the river valleys (if not captured). In total, 64 springs have been mapped within the catchment and there are some areas (mainly in plutonic rocks) where many small springs could not be mapped separately.

The Kouris catchment is drained by the rivers Kouris and Limnatis, which consist of spring water during dry seasons, and the river Kryos, which normally dries in summers (Fig. 3.2). Kouris is the largest river in Cyprus and has for the last 30 years an average annual streamflow of 36 Mm³/year.

3.3. ANALYTICAL METHODS AND SAMPLE LOCATIONS

For the present study, during 1998-2001 we collected and analyzed 224 groundwater and precipitation samples from 114 locations for stable isotopes and 13 ³H samples. The samples were taken from groundwater (springs and boreholes), and from precipitation in different seasons and years to investigate their temporal variability (see Table 3.1 and Fig. 3.2 for details). The samples were also analyzed for chemical components. All 13 samples for ³H were taken from the sedimentary complex and Pillow Lavas.

Additionally, 67 values for stable isotopes in the Kouris catchment, reported by Jacovides (1979), and 23 isotope analysis in precipitation from GNIP database (http://isohis.iaea.org) were used.
Hydrogen and oxygen isotope ratios are expressed by $\delta$D and $\delta^{18}$O respectively, where $\delta = [(R_{\text{sample}}/R_{\text{standard}})-1] \times 1000$ ($\%$), R is the ratio of D/H or $^{18}$O/$^{16}$O in sampled water ($R_{\text{sample}}$) or in Standard Mean Ocean Water ($R_{\text{standard}}$). All analyses were carried out in GSF, Munich; the analytical errors ($2\sigma$) were 0.1 $\%$o for $\delta^{18}$O, 1 $\%$o for $\delta$D and 0.7 – 2.0 TU (1 TU = 0.118 Bq/l) for $^3$H.

3.4. RESULTS

Stable isotopes in precipitation

Figure 3.3 presents $\delta$D versus $\delta^{18}$O for 70 precipitation samples collected in the Kouris catchment, showing that the deuterium excess, defined by Dansgaard (1964) as $d = \delta$D – 8*$\delta^{18}$O, is quite high and refers to the Mediterranean Meteoric Water Line ($d = 20$) rather than to the Global Meteoric Water Line ($d = 10$). The linear regression line:

$$\delta$D = 6.6*$\delta^{18}$O + 10.9

(3.1),

differs from the Global Meteoric Water Line ($\delta$D = 8* $\delta^{18}$O + 10, Dansgaard 1964, Craig 1965), indicating evaporation during precipitation. Equations with similar slopes (6.4±0.5) are published for the precipitation in the northern part of Epirus in Greece (Leonitiadis and Nikolaou, 1999), in Eastern Macedonia (Leonitiadis et al., 1996), and in Central Macedonia (Christodoulou et al., 1993).

Seasonal variations of stable isotopes content in precipitation reach 27$\%$o for $\delta$D and 4.0$\%$o for $\delta^{18}$O in 2000-2002 (see Fig. 3.4 for $\delta$D). Isotopically lightest water (for all three meteostations – Figure 3.4) precipitates during November – December and it gradually becomes heavier afterwards. General decrease of deuterium excess in precipitation in hot seasons indicates evaporation as an important reason of seasonal changes.

Single rainfall events might differ in isotopic content even more and their compositions are influenced by precipitation amount – Figure 3.5.

Averaging isotope values of rainfall samples for September-April 2000/2001 results in the following Table 3.2, suggesting, in agreement with Verhagen et al. (1991), general altitude gradients in precipitation –2.2 $\%$o / 100 m for $\delta$D and -0.34 $\%$o / 100 m for $\delta^{18}$O.
Deuterium excess decreases with decreasing altitudes (Table 3.2) which indicates enhancing evaporation. Thus, altitude effect in precipitation can not be separated from effect of changing evaporation rate with surface elevations.

**Stable isotopes in surface water**

Surface water (rivers, dams) shows large variability in the isotopic content (from -6.98 ‰ to -3.24 ‰ for δ¹⁸O and from -39.4 ‰ to -18.2 for δD). The isotopic compositions are affected by evaporation at different seasons and mixture of baseflows from the different altitudes with surface runoffs. The isotope analysis of surface water are available in the Engineering Geology library of ETH, Zuerich and will be published later.

**Stable isotopes in ground water**

The range of variations of isotopic contents of all sampled groundwaters is smaller, than in precipitation and in surface water: it extends from -7.66 ‰ to -4.61 ‰ for δ¹⁸O and from -41.00 ‰ to -24.75 ‰ for δD.

Figure 3.6 shows δD versus δ¹⁸O for the 234 groundwater samples. Two types of groundwaters are clearly distinguished: sedimentary complex with isotopically enriched water and the ophiolitic complex with isotopically depleted groundwater. In total, 52% of groundwaters in the ophiolites and 10% in the sediments fall on the local regression line of precipitation, at the range of uncertainties of measurements (filled symbols in Figure 3.6). The rest of the samples are displaced below the local regression line around a line of a slightly smaller slope ($m = 5.9$, Fig. 3.6). There might be combination of, at least, two reasons to explain this shift. One – is the evaporation prior to- or during infiltration, leading to the observed smaller slopes of a regression line. This factor may have big influence also because of reinfiltration of isotopically enriched and fractionated water remaining after irrigation in summers. Another probable reason is described by Allison et al. (1983) showing, that the samples in the δD-δ¹⁸O diagram are displaced toward a line parallel and below the local regression line because of partial evaporation from soils and dilution by subsequent recharge.

Figure 3.7 shows deuterium excess ($d = \delta D - 8 \times \delta ^{18}O$) in groundwaters of some springs and boreholes in different seasons. No clear tendency in springs is observed: most
of the changes are in the range of measurement errors. Increase of d in dry seasons, observed in deep artesian boreholes of the ophiolitic complex (57/76, 32/76, 31/76, 53/76, v2/94, 146/90 – Figure 3.7), might mean simply evaporation from the borehole space rather than the case of evaporation from the deep aquifer.

The high salinities in the groundwaters from sediments and pillow lavas cannot be explained by concentration because of evaporation alone. Deuterium excess (Table 3.3) shows no correlation with salinities. For example wells 67/01, 68/01 and 183/01 in the sedimentary complex (Table 3.3), located at the same surface elevations, with 7 meters difference of penetration depths (in fact, 67/01 and 68/01 were in the distance of 20 meters from each other), show nearly the same isotope ratios and deuterium excess, but very different chloride contents and total salinities. Similar situation occurred in the boreholes 9/5, 5/5, 10/5 (Table 3.3), which had the same altitudes of locations, δ¹⁸O and δD ratios (which most probably indicates the same degree of fractionation due to evaporation), but very different salinities. The differences in TDS in those boreholes were caused by SO₄⁻, while Cl⁻ contents were not so different; this fact also confirmed that high salinity in Pillow Lavas was caused by rock dissolution, rather than evaporation or transpiration.

From the ³H content, the ground water in the sediments and Pillow Lavas must have had the residence time higher, at least, than 45 years, since it is not influenced by high ³H concentrations of the 1960-th (Table 3.3). This fact favors also the explanation of the origin of the brackish water from rock dilution rather than through concentration due to evaporation.

Figure 3.6 shows that isotope compositions of groundwaters are generally different from those in precipitation because of evaporation during infiltration. Thus, the altitude effect in the precipitation can not be used for the description of groundwater origin (recharge conditions) in the Kouris catchment. Therefore, the following methodology was applied to obtain the regression between stable isotopes content of recharge and topographic elevations. In a first step, the regression equation between deuterium contents and surface elevations was established, based on the deuterium data from 33 ophiolitic springs and their average catchment altitudes, obtained from the Digital Elevation Model (Hall, 1998). Samples of rainy seasons were used, in an assumption that obtained values of average surface catchments represent average altitudes of recharge. As enrichment due to
evaporation generally less affects the $\delta^D$ than the $\delta^{18}O$ values, the $\delta^D$ relationship with altitudes was used. The resulting linear regression is shown in Figure 3.8 and described by equation:

$$Z[m] = (-2034 \pm 253) - (90 \pm 7) \times \delta^D \quad (3.2)$$

From Equation 3.2 the altitudes of the recharge areas for the remaining 85 samples (we excluded samples with clear evidence of evaporation) were calculated and plotted in Figure 3.9 against the altitudes of the sampling points for three groups of samples (1 - ophiolitic complex, excluding pillow lavas, 2 - alluvium in the southern part of the catchment, 3 - sedimentary rocks and pillow lavas). Although the data in Figure 3.8 were highly scattered, most of the samples from ophiolites and sedimentary rocks fall between two lines: $Y = X + 253$ m and $Y = X - 253$ m (Fig. 3.9), where $Y$ and $X$ refer to altitudes of recharge and altitudes of sampling points respectively. This fact means, that, within the range of accuracy of Equation 3.2, the groundwaters sampled within the ophiolites and the sediments, originates from the altitudes not far from the altitudes of the sampling points. An exception from this rule is the Phylagria spring, discharging water from higher altitudes. Other exceptions are few groundwater samples from the ophiolites, displaced below the line $Y = X - 253$ m, probably due to evaporation. On the contrary, the samples from the alluvium aquifer seem to contain systematically water from higher altitudes (the altitudes of the ophiolitic complex). This means that this aquifer is either fed by water from the rivers or by groundwater, flowing from higher altitudes via deep fractures zones in the sedimentary aquifer. The $^3H$ and chemical data support the first hypotheses (river origin of the groundwater in the alluvium aquifer), because the $^3H$ content of water in boreholes 34/96, 10B-st, 71/92, 21/00, 22/00, 23/00, 24/00, drilled in the alluvium aquifer, varied from 3.2 TU to 10.2 TU, similar to the range of $^3H$ variations, observed in the rivers of the catchment. Additionally the groundwaters in the alluvium aquifer are of Ca-Mg-HCO$_3$, Mg-Ca-HCO$_3$ or Mg-Na-HCO$_3$ types, always contain considerable amounts of Mg$^{2+}$ in their cation composition. As Mg$^{2+}$ is one of the leading cations of the groundwater within the ophiolitic aquifer, it rather confirms the hypothesis that water infiltrates from the rivers, fed by ophiolitic springs, but does not flow via the sediments.
Calibration of the steady state groundwater model with the stable isotopes data

A steady state 2-dimensional groundwater model was developed during a previous stage of the research (Boronina et al., 2003). It is based on the MODFLOW code (Harbaugh et al., 1996). The mesh, the external and internal boundaries (drains, rivers, constant head, impermeable boundaries) are presented in Figure 3.10. The recharge is imposed on the top of the layer and zonated according to the surface elevation and rock types. The recharge rates and the transmissivities are shown in Figure 3.11 while the resulting piezometric surface is presented in Figure 3.12. In spite of a good fit of observed and simulated piezometric heads, the calibration of the model only based on piezometric observations resulted in a reliable relation “recharge rates-transmissivities” rather than in absolute values of the parameters. Further model calibration by river leakages allowed us to obtain the absolute values of the total recharge, although the quality of the data used for calibration (baseflows, separated from rivers hygrograms by the sliding interval method) forced us to look for additional proofs of the conclusions by Boronina et al. (2003).

Thus, in the present study, our aim was to calibrate the absolute values of the recharge by stable isotopes ($\delta^2$D) data. During the calibration procedure one basic variant (the resulting model from Boronina et al. (2003)) and six variants with different absolute values of recharge were created. The ratios between the recharges of different zones in every modeling variant were kept equal to those in the basic variant, while the absolute values of recharge were changed comparatively to the basic variant by factors 0.6, 0.75, 0.9, 1.2, 1.5, and 2. Then the transmissivities of every modeled variant were changed respectively in order to obtain the best fit of the simulated piezometric heads to observed ones. Thus, seven models, calibrated only by piezometric heads, with different absolute values of recharge, were created. To develop the transport model using the code MT3D (Zheng, 1990), we had to give porosity and dispersivity values of different lithologies, and deuterium concentrations of recharge. The transport was modelled for a time long enough to reach a steady state deuterium distribution in the aquifer, which was assumed to be in the reality. Though some general values of porosities were input, they did not make a difference for a steady state transport model. The longitudinal and transverse dispersivities were assumed to be 60 and 6 meters respectively and reflected macrodispersion in an average regional scale of 10 km (Gelhar et al., 1992). Hybrid MOC method was used as a
solution scheme to avoid numerical dispersion. The altitude effect of the deuterium content in the recharge was approximated by 8 zones according to surface elevation, the δD concentrations of the recharge for every zone were calculated by Equation 3.2. The simulated concentrations of δD were then compared with the observed ones for all modeled variants. Figure 3.13 shows a scatter diagram between the 94 observed and simulated deuterium concentrations in the aquifer for the basic variant, while Figure 3.14 presents the spatial distribution of simulation errors and map of observation points. A high scatter of the points is mainly caused by the uncertainties of input concentrations, propagated from the uncertainties of Equation 3.2; an additional scatter below the line Y=X is probably caused by fractionation due to evaporation, which is not possible to model with MT3D. Figure 3.15 presents the mean square root of the square deviations between observed and simulated δD values, averaged for seven locations in the Kouris catchment, and plotted against the steady state recharge (in mm). The function in Figure 3.15 has a minimum between 100 mm and 130 mm, suggesting this range as the optimal recharge rate from the calibration with stable isotopes.

3.5. DISCUSSION

In this part of the paper we will discuss reliability of the conclusions of stable isotopes studies. All conclusions were made only concerning precipitation and groundwater of the Kouris catchment. Despite the big amount of surface water analysis, we did not describe them in details, since we consider this as a topic for our future investigations, where other approaches and types of models will be applied.

Description of stable isotopes in precipitation

In spite of rather large amount of precipitation samples used for this study, we are still limited only to general observations about the stable isotopes in precipitation in Cyprus. For example, our data can not confirm, whether the seasonal changes in the stable isotope contents in 2000 - 2002 (Fig. 3.4) were common for longer period or not. Also, an accurate description of the altitude effect would require more points to estimate the uncertainties.
Nevertheless, we still consider the present studies as useful, since it is, to our knowledge, one of the first published descriptions of stable isotopes in precipitation in Cyprus.

**Origin of groundwater**

The uncertainty related to Equation 3.2 is high, most probably because of the high uncertainty in determination of the average surface elevations of springs catchments. At the present stage, one couldn’t reduce this uncertainty. It then propagates through all estimates based on Equation 3.2 and results in a large scatter in Figure 3.9. That is why we could draw only general and comparative conclusions for the different water origins in the Kouris catchment. It seems, for example, that the groundwaters in the sedimentary complex come from the local recharge, rather than from the ophiolites at higher altitudes. On the contrary, alluvium aquifer is fed by water from the ophiolites via the rivers, and this conclusion is supported by chemical, $^3$H data and results of numerical modelling. These results might be further uncertain because of evaporation, which theoretically decreases the deuterium excess and makes the calculated altitudes of recharge even smaller. We tried to avoid this by eliminating the samples with pronounced evaporation, but it was not sufficient. In that case, real altitudes of recharge might be slightly higher in the sedimentary complex and in the alluvium aquifer.

**Calibration of the numerical model**

The uncertainties related to Equation 3.2 propagate through the transport model. We consider that the deuterium concentrations imposed in the recharge have an uncertainty of ±3 per mille which results in the scatter visible in Figure 3.13, and this inaccuracy can not be reduced. The mean square root of the square deviations between 94 observed and calculated deuterium concentrations is 1.7 per mille, which still looked satisfactory, although there were 10 points, where the simulated values exceeded the observed ones by 3 – 4 per mille (more, than the range of input uncertainty). The evaporation might have been the cause of these big discrepancies; another possible reason is the local heterogeneities which are not included in a regional deterministic model. Thus, in Figure 3.15, we would rather look not at the absolute values, which were mostly near the range of input errors, but at the type of
3.6. CONCLUSIONS

The study of oxygen and hydrogen isotopes in precipitation and in groundwater revealed some characteristic features of the Kouris catchment. The precipitation is affected by evaporation during rainfall events. The calculated local regression line of precipitation is: 
\[ \delta D = 6.5 \delta^{18}O + 10.6. \]
The altitude gradients in precipitation are estimated to be \(-2.2 \% / 100\) m for \(\delta D\) and \(-0.34 \% / 100\) m for \(\delta^{18}O\). The analysis of groundwater samples showed, that during- or after infiltration evaporation occurs as well, so that it was not possible to use the altitude effect of precipitation for the description of groundwater origin. Thus, the altitude gradients of stable isotopes in the aquifer were obtained from the data of 33 springs in the ophiolitic complex. From the regression between deuterium data and surface elevations, the altitude of recharge for all samples were calculated and compared to the altitudes of the sampling points. We can conclude that the ophiolites, except Phylagria spring, and the sedimentary complex contain water from the local recharges at high and low altitudes respectively. The high salinity of the ground water in the sediments is caused by rock dissolution, rather than evaporation. On the contrary, the groundwater in the alluvium aquifer originates from the high altitudes in the ophiolitic complex. \(^3\)H and chemistry data further supported this hypothesis. Additionally, deuterium data were used for the calibration of the recharge rates of a steady state groundwater flow and transport model. The resulting calibrated total steady state recharge rate is 100-130 mm per year, which is in agreement with the recharge, obtained by the chloride mass-balance method and by the previous calibration of a numerical model with the river leakages (Boronina et al., 2003).

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<thead>
<tr>
<th></th>
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</thead>
<tbody>
<tr>
<td></td>
<td>Quantity of samples</td>
<td>Quantity of sampling points</td>
</tr>
<tr>
<td>Rainfall samples</td>
<td>23</td>
<td>2</td>
</tr>
<tr>
<td>Spring samples:</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ophiolitic complex (ultramafic rocks, gabbros, diabase dykes)</td>
<td>27</td>
<td>15</td>
</tr>
<tr>
<td>Alluvium aquifer</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Borehole samples:</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ophiolitic complex (ultramafic rocks, gabbros, diabase dykes)</td>
<td>36</td>
<td>17</td>
</tr>
<tr>
<td>Ophiolitic complex (Pillow Lavas)</td>
<td>4</td>
<td>3</td>
</tr>
<tr>
<td>Sedimentary complex</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Alluvium aquifer</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Total</td>
<td>90</td>
<td>37</td>
</tr>
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Table 3.2. Isotopic contents in rainfalls, averaged for September-April 2000/2001, at the meteostations in Agros (1013 m a.s.l.) and at the Kouris Dam (242 m a.s.l.)

<table>
<thead>
<tr>
<th></th>
<th>$\delta^{18}$O, $%$</th>
<th>$\delta D$, $%$</th>
<th>Deuterium excess</th>
</tr>
</thead>
<tbody>
<tr>
<td>Agros</td>
<td>-7.00</td>
<td>-36.28</td>
<td>19.7</td>
</tr>
<tr>
<td>Kouris Dam</td>
<td>-4.38</td>
<td>-19.40</td>
<td>15.6</td>
</tr>
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</table>
Table 3.3. The examples of TDS, Cl', δ¹⁸O, δD contents and deuterium excess in the groundwaters of Pillow Lavas and sedimentary rocks

<table>
<thead>
<tr>
<th>Well N</th>
<th>Altitude, m</th>
<th>Rock</th>
<th>TDS, mg/l</th>
<th>Cl', mg/l</th>
<th>δ¹⁸O, per mille</th>
<th>δD, per mille</th>
<th>Deuterium excess</th>
<th>²H, TU</th>
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</thead>
<tbody>
<tr>
<td>67/01</td>
<td>318</td>
<td>Sedimentary rocks</td>
<td>1288</td>
<td>152</td>
<td>-5.11</td>
<td>-24.8</td>
<td>16.1</td>
<td>&lt;0.8</td>
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<tr>
<td>68/01</td>
<td>318</td>
<td>Sedimentary rocks</td>
<td>1605</td>
<td>220</td>
<td>-5.18</td>
<td>-25.4</td>
<td>16</td>
<td></td>
</tr>
<tr>
<td>183/01</td>
<td>318</td>
<td>Sedimentary rocks</td>
<td>1133</td>
<td>160</td>
<td>-5.05</td>
<td>-25.3</td>
<td>15.1</td>
<td></td>
</tr>
<tr>
<td>8/01</td>
<td>378</td>
<td>Pillow Lavas</td>
<td>522</td>
<td>82.8</td>
<td>-5.14</td>
<td>-27.1</td>
<td>14</td>
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<td>273/01</td>
<td>640</td>
<td>Pillow Lavas</td>
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<td>-27.8</td>
<td>14.9</td>
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<tr>
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<td>457</td>
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<td>87</td>
<td>-5.20</td>
<td>-28.0</td>
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<td>272/01</td>
<td>525</td>
<td>Pillow Lavas</td>
<td>515</td>
<td>127</td>
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<td>-28.6</td>
<td>15.4</td>
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<td>9/5</td>
<td>495</td>
<td>Pillow Lavas</td>
<td>518</td>
<td>23.6</td>
<td>-5.27</td>
<td>-26.2</td>
<td>16</td>
<td></td>
</tr>
<tr>
<td>5/5</td>
<td>495</td>
<td>Pillow Lavas</td>
<td>651</td>
<td>39.5</td>
<td>-5.25</td>
<td>-26.8</td>
<td>15.2</td>
<td>&lt;1.8</td>
</tr>
<tr>
<td>10/5</td>
<td>495</td>
<td>Pillow Lavas</td>
<td>908</td>
<td>39.4</td>
<td>-5.21</td>
<td>-27.6</td>
<td>14.1</td>
<td></td>
</tr>
</tbody>
</table>
Figure 3.1. Simplified geological map of Cyprus showing the location of the Kouris catchment.

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Chapter 4

Application of tritium in precipitation and in groundwater of the Kouris catchment (Cyprus) for description of the regional groundwater flow

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ABSTRACT

$^3$H concentrations in groundwater and precipitation were studied to describe groundwater flow in the Kouris catchment (Cyprus). The catchment consists of an ophiolitic complex in the North and overlying sedimentary rocks in the South. It is characterized by strong heterogeneity of the underground and steep slopes of the topography.

An input function of $^3$H in precipitation in Cyprus is obtained as a result of a linear regression between data in Cyprus and in neighbouring meteorological stations. This function is applied to calculate groundwater residence times in different aquifers by lumped-parameter and numerical models.

The paper presents the input data, calibration and conclusions of a transient three-dimensional groundwater flow model of the Kouris catchment. It also describes the simulation of $^3$H transport with the MODFLOW-PMPATH codes and presents distributions of residence times for springs of the ophiolitic complex. Average residence times vary between 14 and 30 years that corresponds to the effective porosity of the ophiolites equal to 0.04-0.06.

KEYWORDS: tritium, ophiolites, semi-arid climate, Cyprus, groundwater, recharge.
4.1. INTRODUCTION

The Kouris catchment in Cyprus contains the major groundwater resources of the island. Consisting of fractured ophiolites in the North and low-permeable sediments in the South, the aquifer is complex; thus, a quantitative description of groundwater flow remains a challenging scientific problem.

$^3$H concentrations in groundwaters have often been applied to validate conceptual hydrogeological models. In most of the studies, $^3$H has been used as a dating tool for groundwaters (Maréchal & Etcheverry, 2003; Stimson et al, 1996; Plummer et al., 2001; Zoellmann et al., 2001). Commonly, residence times of groundwaters are estimated by converting transient input function of $^3$H in precipitation into simulated concentrations in groundwater through numerical and analytical (lumped-parameter) models.

The lumped-parameter, or box-models (Maloszewski & Zuber, 1982; Zuber, 1986, Maloszewski, 1996; Etcheverry, 2000) imply several distribution functions (Piston-flow, exponential, dispersion – the most common) for transformations of $^3$H contents in precipitation into observed concentrations in groundwaters. The main advantage of this type of models is the minimum information required for their calibration. On the other hand, it can be also the disadvantage since it is not possible to incorporate quantitatively all available information about a study area. Another disadvantage of these models is that a type of a transfer function must be known a priori. Additionally, for a transfer function of the dispersion type (describing the most common behavior of hydrogeological systems) a dispersion coefficient is required, which brings a large uncertainty in residence times calculations.

An alternative approach for the interpretation of $^3$H data is numerical groundwater flow and transport modelling (see Konikow, 1996 for a review). Although, these models require a few additional parameters beside input and output concentrations, they result in a more quantitative description of a hydrogeological system. These methods allow not only to obtain the mean residence times of groundwater (which in most cases are useful to forecasts groundwater contamination but not for other water resources problems), but also to validate the main concepts of numerical models and, in an ideal case, to calibrate some model input parameters, for example, effective porosities. Disadvantages of these models are related to their
requirements of larger (in comparison with box-models) amount of input data and more sophisticated studies required for their development.

Finally, both numerical and box-models have uncertainties, propagated from uncertainties of input function of $^3$H in precipitation and they have non-uniqueness due to a bell-shape type of the input function.

The aim of our studies is to use the information related to $^3$H contents in precipitation and groundwaters of the Kouris catchment in Cyprus for validation of a regional hydrogeological concepts and for calibration of a transient numerical groundwater flow model.

There were several attempts to use $^3$H data for the description of Cyprus hydrogeology. In 1972-78 research contracts on environmental isotope survey of Cyprus were sponsored and conducted by IAEA in collaboration with the Water Development Department (WDD) of Cyprus. The research carried out consisted of sampling for environmental isotope contents of springs, boreholes and baseflow, originating from the Troodos Mountain. Additionally, samples of snow and rainfall were analysed for $^3$H. Despite a great number of analysis (more than 200 samples), the conclusions were quite uncertain due to the high heterogeneity of the underground media and a big area of coverage. Thus, the results of the survey were never published although it was integrated in a technical report (Jacovides 1979). In the 1980s, isotope survey was conducted on the sedimentary complex located west of Larnaka - approximately 30 km to the east of the Kouris catchment (Verhagen et al. 1991). Other studies were concentrated on the unsaturated zone of the sedimentary complex in the southeast of Cyprus in the Mesaoria area; those were related to the estimation of recharge from precipitation (Kitching et al. 1980).

In this paper, we integrate new and reported (Jacovides 1979) $^3$H data in precipitation and groundwater for the description of the hydrogeology of the Kouris catchment.

In the first part, $^3$H input function in precipitation and its uncertainties are obtained by linear regression between the data in Cyprus and in neighbouring meteorological stations.

Then, different aquifers of the Kouris catchment are compared relatively to their $^3$H groundwater contents and roughly estimated residence times.
To simulate the $^3$H transport in the Kouris aquifer, the groundwater flow model of the Kouris catchment (Boronina et al., 2003) is further modified and its calibration and conclusions about aquifer storativities are discussed in the next part of the paper.

In the last part, the paper describes simulation of $^3$H transport with the MODFLOW-PMPATH codes and presents distributions of residence times for springs of the ophiolitic complex. Finally, the $^3$H simulations result in conclusions about effective porosities, types of transfer function and sensitivities of $^3$H concentrations to residence time along a single flow path.

4.2. HYDROGEOLOGY OF THE KOURIS CATCHMENT

The hydrogeology of the area is described in details by Boronina et al. (2003a), Jacovides (1979), Afrodisis et al. (1986); in the present paper, we provide only the most essential information.

The Kouris catchment is bounded to the South by the Mediterranean Sea and covers 300 km$^2$ on the southern side of the Troodos Massif (Fig. 4.1). Elevations range from the sea level to 2000 m within a distance of 30 km with local slopes up to 70 %.

Annual precipitation amounts increase from 300 mm along the coast to nearly 1200 mm in the Troodos mountains, while mean annual potential évapotranspiration for the catchment is 1210 mm.

The Kouris catchment is drained by the perennial rivers Kouris, Kryos and Limnatis (Fig. 4.2), consisting of spring water during dry seasons. Kouris is the largest river in Cyprus and has had for the last 30 years an average annual streamflow of 36 Mm$^3$.

The basin is divided in two main geological zones: an ophiolitic complex in the North and unconformably overlying it a sedimentary complex in the South.

The ophiolitic complex
The ophiolites include ultramafic rocks, gabbros, sheeted dykes and pillow lavas (Fig. 4.2); the rocks are highly heterogeneous due to the presence of different lithological units, which are fractured and altered at different scales. The transmissivities of the Gabbros and the Diabase Dykes vary from 2 m$^2$/day and 703 m$^2$/day with a geometrical mean of 20 m$^2$/day according to the results of 40 pumping tests (Boronina et al., 2003).
The Pillow Lavas are considered to be generally less permeable, although local zones of high conductivity may exist there.

There are 62 mapped springs in the plutonic and intrusive rocks (no springs were observed in the Pillow Lavas); moreover, in some areas (for example the North-East of the Limnatis valley) many small springs are found, which have not been mapped individually.

These crystalline rocks contain the major groundwater resources of Cyprus; the water is stored in the fractured and altered zones of harzburgites, dunites, gabbros and diabases. The big springs are used for drinking/irrigation water supply of several mountain villages (springs Archolochania - an average discharge is 31 l/s, Loumata “B”, “Loumata “C”) and tourist centre “Troodos” (springs Troodos “A”, “B”, “C”).

The sedimentary complex
The sedimentary part consists of chalks, marls, calcarenites and limestones (Fig. 4.2) with an estimated average transmissivity of 3 m²/day; it also consists of river alluvium (sand, gravel) in the Kryos, Kouris and Limnatis valleys. From the boreholes, drilled in the sedimentary complex for the irrigation purpose, 70 % were unsuccessful.

The alluvial aquifer (gravel, sand) is narrow, sometimes less than 50 meters, and discontinuous, although it contains major amount of water within the sedimentary complex.

There are few big springs (five of them are mapped), associated with the alluvium aquifer, discharging water in the river valleys (if not captured). Other springs in the river valleys were observed in the outcrops of sedimentary rocks: they were usually highly mineralised (up to 1600 mg/l) and had small discharges (less, than 1 l/sec).

4.3 ANALYTICAL METHODS AND SAMPLE LOCATIONS

For the present study, 176 groundwater, surface water and precipitation samples from 95 locations were collected and analyzed for $^3$H and major ions (a part of the analysis were done during diploma studies - Jorin, 2001; Moll, 2000; Oertli, 2002; Steiner, 2000). Waters were sampled in different seasons during the years 1998-2001 to investigate their temporal variability (see Table 4.1 and Figure 4.2 for details).
For the $^3$H content in precipitation in Cyprus, 76 measurements were used. Among them, 1) twelve new samples were collected in 2000-2001 in Agros (N: 3863800, E: 501300, altitude: 1015 m), at the Kouris Dam (N: 3842500, E: 493000, altitude: 220 m), and in Saittas (N: 3858300, E: 492000, altitude: 640 m); 2) and 64 analysis originate from the IAEA network (http://isohis.iaea.org) sampled in 1960-1974 in Prodromos (N: 3868300, E: 484200, altitude: 1380 m).

For the $^3$H distribution in the groundwater, 77 measurements, reported by Jacovides (1979) were used in addition to sampling for the present studies.

All samples were analyzed in GSF, Munich; the analytical errors (2σ) were 0.7 – 2.0 TU (1 TU = 0.118 Bq/l).

4.4. RESULTS

4.4.1. Construction of the tritium input function in precipitation for Cyprus

The meteorological station in Prodromos, located at an altitude of 1380 m, in 2 km north-east from the Kouris catchment, was chosen as a representative point for the description of $^3$H concentrations in the regional precipitation. This station has monthly records of $^3$H contents for the period of 1960-1974 (http://isohis.iaea.org), the concentrations had a high positive linear correlation (correlation coefficient is 0.99) with the data at the Nicosia meteorological station.

Table 4.2 presents results of linear correlation analysis between data from Prodromos and from the five meteorological stations, located at latitudes, near the latitude of Prodromos. The meteorological station in Ottawa (Canada) was included in the analysis, because only at that station $^3$H contents in precipitation were measured before the year 1960. A comparative analysis of linear correlations was based on 27 monthly values of $^3$H in precipitation for December 1963 – March 1974, since only for that period the data were recorded in all meteorological stations (Table 4.2).

From the results of Table 4.2, the optimal input function for $^3$H in precipitation at Prodromos was constructed on the linear regression with the data in Ottawa for 1953-1960 (64 values, correlation coefficient 0.95), in Athens - for 1960-1962 and 1976-91 (45 values, correlation coefficient 0.92) and in Ankara - for 1992-1999 (63 values, correlation coefficient 0.87). We used $^3$H concentrations in the arithmetic form for 1950s, while for later periods logarithms of the concentrations were applied in order to
smooth the influence of big values in 1960s on the predicted $^{3}$H concentrations in 1980s-90s.

Thus, calculations of $^{3}$H contents in precipitation at Prodromos were based on the following equations:

\[
C_p = -1.393 + 0.587C_O \quad \text{for 1953-1960} \tag{4.1},
\]

\[
\ln(C_p) = 0.83 + 0.82\ln(C_{Ath}) \quad \text{for 1960-1962 and 1976-91} \tag{4.2},
\]

\[
\ln(C_p) = 0.20 + 0.82\ln(C_{Ank}) \quad \text{for 1992-1999} \tag{4.3},
\]

where $C_p$ and $C_O$, $C_{Ath}$ and $C_{Ank}$ – $^{3}$H contents in precipitations at Prodromos, Ottawa, Athens and Ankara respectively. An average relative uncertainty of the approximation was calculated as:

\[
\sigma = \frac{\sum (C_{p,obs} - C_{p,i})^2}{N} \tag{4.4},
\]

where $C_{p,obs}$ and $C_{p,i}$ – observed and calculated $^{3}$H concentrations in the Prodromos precipitation for the time period $i$.

Uncertainties of the input function of $^{3}$H in monthly precipitation in Prodromos, estimated from Equation 4.4 were: $\sigma = 0.5$ TU - for 1953-1960; $\sigma = 0.3$ TU – for 1960-1962 and 1976-91; $\sigma = 0.3$ TU - for 1992-1999.

An input $^{3}$H function of annual precipitation at Prodromos was built on calculated (Equations 4.1, 4.2, 4.3) and measured monthly $^{3}$H concentrations, averaged for wet seasons (October – March) – Figure 4.3. Dashed lines in Figure 4.3 represent the range of yearly averaged uncertainty according to calculated $\sigma$ (Eq. 4.4). For the years 1963-1976 the data, measured directly in Prodromos, were applied with $\sigma = 0.1$ (mean calculated relative error of the measurements in a laboratory). Additionally the values obtained in the Kouris catchment (Agros village) in November-December 2000 were added to the input function with the uncertainty of measurements equal to 0.7 TU.

4.4.2. $^{3}$H distribution in the waters of the Kouris catchment

Surface water

The $^{3}$H contents in the rivers (Table 4.1) were spatially quite uniform: from 31.8 to 37.4 TU in 1978 and from 2.7 to 8.6 in dry seasons of 1998-2002. These values were similar to average $^{3}$H concentrations in springs of the ophiolitic complex. During rainfalls, $^{3}$H
contents of the rivers were changing due to surface runoff. However, separation of hydrogramms with $^3$H data was not possible, because most of variations were at the range of measurement errors.

**Groundwater**

The ranges of $^3$H concentrations in different years in groundwaters of the Kouris catchment are presented in Table 4.1.

*Ophiolitic complex*

Average concentrations from Table 4.1 indicate, that groundwater, discharging in springs of ophiolitic complex, contains anthropogenic tritium (residence times up to 35 years). In rainy season springs are normally discharging considerable amount of surface water from recent rainfalls (Fig. 4.4) which coincides with the maximum flow rates. Those springs might be in danger because of surface pollution. Some springs in Agros and Agious Theodorous villages are already not suitable for drinking because of petrol and sewage contamination.

$^3$H concentrations in boreholes were highly variable, and calculated residence times, in Piston assumptions (Zoellmann et al., 2001), varied from less than 10 years (30-60 TU in the years 1976-78 for boreholes 19/76, 57/76, 30/76) to more than 45 years (less than 1.5 TU in 2000 for borholes v2/94, w215/90, 146/90). The inter-annual changes were not systematic, since they were dependent on the regime of pumping, rather than on season. Boreholes, which were pumped heavily during dry seasons (typically for irrigation), sometimes eventually yielded water, that was older than 45 years, indicating extraction from old reservoirs. Such water cannot be considered renewable in the short term. An evidence of several boreholes of the catchment, turned to be dry after some period of pumping, supported this idea.

*Volcanogenic rocks*

The aquifer in volcanogenic rocks (Pillow Lavas) is confined and located at the depth of more than 100 meters; no springs were observed. The $^3$H data from 1976-78 and 1998-2002 show, that the aquifer contains the groundwater, infiltrated before 1953 (Table 4.1).

*Alluvium aquifer*

$^3$H contents of the springs in the alluvium were similar to those in the streamwaters (Table 4.1): in dry seasons it was a mixture of the waters from springs of the ophiolites, in wet seasons the water was originated from surface runoff. These facts
further supported the idea of Boronina et al. (2003b), that the alluvium aquifer was fed by the streamwater, both in dry and wet seasons, rather than by the groundwater of the sedimentary aquifer.

Sedimentary aquifer

From seven boreholes drilled in the sedimentary rocks, three had $^3$H concentrations less than 0.8 TU in 2002, corresponding to residence times, higher than 48 years; the remaining four contain tritium in an amount of 2.3-3.1 TU, indicating probably mixing with recent water. Comparatively high amounts of NO$_3$ (more than 50 mg/l in two boreholes), usually accompanying the crop fertilization, also supported the idea of “young age” of the water, although, this might be due to reinfiltration of water for irrigation in the borehole itself. Existence of “old” water (less than 1 TU in 1984 - Verhagen et al. 1991) in the same kind of rocks in 30 km to the east from the Kouris catchment also supported the idea of higher groundwater residence times in sedimentary rocks.

The groundwater $^3$H contents of the sedimentary rocks (0.8±0.7 TU) differed from those in streamwater (3-4 TU) even in the vicinity of the rivers, which suggested groundwater leakage from the aquifer to the rivers, rather than opposite.

4.4.3. Modelling of transient groundwater flow in the Kouris catchment

A transient 3-Dimensional groundwater flow model was developed on the base of a regional steady state model (Boronina et al., 2003a), which was converted to a transient regime and further discretized (vertically and areally). The MODFLOW code (Harbaugh and McDonald, 1996; Chiang and Kinzelbach, 2001) was used for simulations. The model, on one hand, was expected to reproduce temporal baseflow and piezometric variation with input transient recharge and, on the other hand, to be appropriate for simulation of a tritium transport in the aquifer.

Time-independent parameters

A regular model grid was refined to the size of 125 X 125 meters; while vertically the aquifer was divided into four layers. Two upper layers were associated with an unsaturated zone in order to model the tritium delay; two lower layers represented the fractured aquifer, with rivers and springs located in the third layer from the top (Fig.
All layers, including two upper ones, were modelled as confined in order to avoid drying up upper layers while simulating a tritium transport.

The total transmissivities were kept equal to the transmissivities of the steady state model (Fig. 4.5) and were distributed vertically between 2 lower layers, representing the aquifer. For two upper layers a vertical hydraulic conductivity was input equal to 0.1 m/day for the first upper layer (soils) and 0.5 m/day for the second layer (fractured ophiolites and sediments) while horizontal hydraulic conductivity was assumed to be negligibly small in order to model only vertical transport in an unsaturated zone.

External and internal boundaries, input parameters for River and Drain packages were transferred from the steady state model (Fig. 4.5) and slightly corrected during a calibration procedure.

**Time-dependent parameters**

The years 1978-1999 were modelled by 21 stress-period (each stress-period had a length of one year) with transient recharge and evapotranspiration. The most complete input parameters and calibration data existed for the years 1984-99, consequently the main model constraints were based on that period.

The input Recharge for every year from 1984-99 was calculated by a simple water balance equation:

\[
Recharge = Rainfall - \text{Actual Evapotranspiration} - \text{Surface runoff} \pm \pm \text{Changes in Groundwater Storage} \tag{4.5}
\]

The Average Actual Evapotranspiration \((\text{AvAcEv})\) for 1984-99 was estimated as:

\[
\text{AvAcEv} = \left( \sum_{15 \text{ years}} \text{Rainfall} - \sum_{15 \text{ years}} \text{Streamflow} - \sum_{15 \text{ years}} \text{Water demands} \right) /15 \text{ years} \tag{4.6},
\]

where: \(\text{AvAcEv} = 184.6 \text{ Mcum.m/year}\)

Actual Evapotranspiration was assumed to be uniform for all years and equal to Average Actual Evapotranspiration (Eq. 4.6). This assumption was based on the fact, that potential evapotranspiration, estimated by the modified Penmann equation, stayed almost constant through the years (Boronina et al., 2003). We admit, although, that this assumption has serious limitations in the semi-arid climate, where actual evapotranspiration is highly dependent not only on climate, but also on the amount of water available in the ground.
Surface runoffs were estimated by hydrogramms separations with fixed- and sliding interval methods (Pettyjohn and Henning 1979; Sloto and Crouse 1996).

The values of all the variables for Equation 4.5 are presented in Table 4.3 (columns II, III, IV). In the column IV ("Recharge-Changes in Groundwater Storage"), positive values indicate occurrence of recharge while negative ones are related to a decrease of groundwater storage due to excess of total evapotranspiration.

Columns V and VI show total baseflows, separated from river hydrogramms and total water demands (Boronina et al., 2003a).

Averaging recharge values (positive changes in the column IV) of Table 4.3 results in an annual amount of 38.4 Mm³, which is equal to 17 % of the average annual rainfall (223.0 Mm³). This percentage is in agreement with the results of chloride mass-balance method (Boronina et al., 2003a).

For the years 1978-83, the estimates of the input recharge and evapotranspiration were more uncertain because of lack of data. The calculations were based on the relationship between rainfall and recharge (evapotranspiration), empirically derived for the period 1984-1999 (Fig. 4.6):

\[
\text{Recharge (Evapotranspiration)} = 0.92 \times \text{Rainfall} - 184.2
\]

Recharge zonation was kept equal to that of the steady state model with a gradual increase of recharge with altitude and comparatively low recharge in the sedimentary complex (Fig. 4.7). For the years when evapotranspiration exceeded recharge, evapotranspiration in the river valleys was input (Fig. 4.7).

The model was calibrated on transient piezometric heads observations, on river baseflows and springs discharges. The main calibrated parameter was a regional storativity. The model showed that only quite a high value of that parameter led to a satisfactory agreement between observed and simulated piezometric heads. The total storage was distributed in four layers in the following way: the first layer (from the top) – 0.0025; the second – 0.0015, the third – 0.005 and the fourth – 0.01. A high input storativity of the upper layer could be related to the partially unconfined regime, while a high storativity of the fourth layer was due to its probable large (more than 500 meters) thickness. The initial storage coefficient of the third layer was obtained from a pumping test, conducted in gabbro.

Figure 4.8 shows the scatter diagram between 210 observed and simulated transient piezometric heads in 29 boreholes with the variance 8612. To obtain final results, recharges and evapotranspirations of extreme (wet and dry) years were slightly
smoothed. Discrepancies between measured and simulated values did not exceed 100 meters, that was an a priori expected accuracy for the Kouris catchment (Boronina et al. 2003a). Figure 4.9 presents the examples of comparison of observed and simulated transient piezometric heads for two boreholes, showing similar amplitudes of simulated and measured heads. The comparison of modelled baseflows and those, derived from hydrogramms, is presented in Figure 4.10. The final calibrated storage coefficient did not allow to simulate amplitudes of annual baseflows, equal to ones in Table 4.3; simulated transient baseflows were smoother. Although by decreasing of a storativity twice, we could “sharpen” simulated baseflow (Fig. 4.10), but a variance between observed and simulated piezometric heads would increase to 9370. Doubling storativities would improve slightly (variance 8417) piezometric heads fitting, but further smooth peaks of simulated baseflows (Fig. 4.10).

4.4.4. Modelling tritium transport in the aquifer

The aim of modelling was to reproduce $^3$H concentrations in the groundwaters, discharging from springs in the ophiolites in dry seasons of 1976-77, and 1998-99. All analysis of rainy seasons were eliminated from the calibration procedure in order to avoid modelling of coupled surface/groundwater flow. We did not use either analysis from the alluvium aquifer, or any data from boreholes because the first ones were mixtures of streamwaters and the second ones were associated with different and unknown depths and pumping histories. Thus, the $^3$H transport in the aquifer was simulated up to a certain depth (approximately 100 meters) related to the zone of active water exchange of springs.

The PMPATH code (Chiang and Kinzelbach 2001) was used to model particle movement backward from spring cells to recharge areas. In every spring cell, 30 particles were injected; those were put on circles with radiiuses of 50 meters at two levels: 3 and 6 meters from the top of the 3-rd layer. Further increase of particle amounts did not change the final distributions of flowpaths.

Residence times along different flowpaths for all observed springs were calculated under transient groundwater velocities. The calibrated parameters included porosities of the aquifer and the unsaturated zone, and their thicknesses.

$^3$H transient concentrations in the aquifer were calculated from MODPATH residence times and from the $^3$H input function, corrected for decay for the years 1976-77 and the years 1998-99. Consequently, the uncertainties of the input function
propagated in the uncertainties of simulated $^3$H concentrations. Those were for some years rather high (for example, 20 TU for groundwaters with the residence time of 10 years, sampled in 1976).

A scatter diagram of observed and simulated $^3$H concentrations for the best fit is presented in Figure 4.11, there for simulated points we took the values, closest to the observed ones within the range of uncertainty of simulations. The list of calibrated parameters for the best fit is shown in Table 4.4, while the locations of springs, used for calibration are presented in Figure 4.12. The resulting residence times of water discharging from the springs sampled in 1998/99 are shown in Table 4.5; the average residence times were sensitive to a period: in the year 1998 after several dry years they were higher, than in the year 1976. The transfer functions for the springs were of the Piston or dispersion types with different macrodispersivity values (see Figure 4.13 as an example of a dispersion distribution function for Archolochania spring).

The resulting concentrations were rather sensitive to $^3$H content of rain water, arriving via a single flowpath, while this did not influence the average residence time. Thus, 19-20 years of average residence time could be associated with the whole range of concentrations from 30 TU to 130 TU in 1976-77. For samples taken in 1998, it was enough to have two flowpaths with residence times of 36 year, which would not change the average residence times (19-20 years), but bring the simulated concentration in springs from 7 TU to 13 TU. This fact means, that numerical modelling of $^3$H concentrations in the aquifer is a difficult task due to heterogeneity, though even quite big differences between simulated and observed single values (in case they are normally distributed) result in the correct integral residence times.

The resulting concentrations and the residence times were very sensitive to porosities of the ophiolitic aquifer and unsaturated zone. The optimal porosities, derived from model calibration, were between 0.04 and 0.06 (Table 4.4, Figure 4.11). Increasing or reducing porosities by a factor 3 resulted in unacceptable simulated $^3$H concentrations (Fig. 4.11) with residence times 29-40 years and 10-13 years for higher and lower porosities, respectively.

Figure 4.14 shows the delineation of the catchments of some springs of the ophiolitic complex. The catchments were rather narrow and had lengths up to five kilometres originating in areas of higher surface elevations. This fact must be taken into account in case of planning protections of springs as drinking water supplies.
4.5. DISCUSSION

Modelling transient groundwater flow
The storativity calibration, based on boreholes observations, contradicted to that, resulted from hydrogramm separation.

One reason of the contradiction could be an over-estimation of baseflows obtained from hydrogram separations for extremely wet years. In Cyprus, big amounts of rainfall are often caused by intensive rains rather than big quantity of rainy days. In that case water comes to soils and drains slowly to the rivers, still not infiltrating in a deep aquifer. The assumption of complete ceasing of surface runoff in two days (Sloto and Crouse, 1996) can be wrong for big rains and high soil capacity. Thus, baseflows, for wet years, simulated by the numerical model might be more realistic than those, obtained from hydrograms separation. Note, that baseflows for dry years and the periods with average rains, obtained from hydrogramms separation, were pretty similar to those from numerical modelling.

Another reason for discrepancies might be a difference between storage parameters, controlling baseflows and those, responsible for piezometry variations. In the Kouris catchment, boreholes were drilled in the most potentially productive zones and often worked under unconfined conditions; thus, piezometry variations represent storage of local unconfined zones around boreholes. On the contrary, baseflows mainly originate from ophiolitic springs of the upper part of the catchment and, consequently, represent regional storage parameters of a fractured aquifer.

Modelling of $^3$H transport in the aquifer
The presented transient model describes regional distribution of tritium in the aquifer, thus, the discrepancies between observed and simulated tritium concentrations for some springs (Fig. 4.11) are expected results of heterogeneity of the rock properties. For example, in September 1976, the springs Loumata “A” and Loumata “C” had tritium concentrations 30.3 and 61.5 TU respectively. In the model these springs were located in neighbour model cells and, consequently, had to have similar simulated tritium concentrations. However, attempts to represent the local heterogeneities at the regional scale of available information will result in further over-parametrisation of the model. Thus we still prefer to keep results at the regional scale rather than for every single spring.
The narrow shapes of the catchments for the majority of the springs were related to the regional piezometry field strongly controlled by the rivers. However, unfortunately, we do not have any field evidence to prove or reject this hypothesis.

4.6. CONCLUSIONS

Seventy seven $^3$H analysis from Jacovides (1979) and 176 new ones in the Kouris catchment allowed us to construct an input function of $^3$H in precipitation and to obtain essential information about the aquifer.

The groundwaters in the sedimentary aquifer and pillow lawas had the mean residence times of more than 48 years; thus they were considerably "older" than the groundwaters of the ophiolitic complex due to lower flow velocities.

Even in the vicinity of the rivers, the sedimentary complex contained water with low $^3$H content, indicating that the sedimentary aquifer was discharging to the rivers rather than the other way around.

The groundwaters in the alluvium had the same $^3$H contents as river waters, and consequently different $^3$H contents from the groundwaters of the surrounding sediments.

Low $^3$H contents (less than 1 TU) were also observed in groundwaters of deep boreholes of the ophiolitic complex, exploited for irrigation in dry seasons. Consequently, those boreholes were pumping water having a long period of renewal.

The transient groundwater flow model showed contradictions between storage coefficients, calibrated on the baseflows, and those calibrated on the piezometry data from boreholes. The final storativity (0.02) was obtained from the model calibration and was interpreted as representing the storage of the deep fractured aquifer.

Simulation of $^3$H transport in the aquifer with PMPATH transient model allowed to obtain the distributions of the groundwater residence times and to reproduce the observed tritium concentrations in the aquifer. The mean residence times of spring water in the ophiolitic complex were estimated to be 14-30 years, although all springs contained considerable amount of surface water in wet seasons. The biggest springs in the Southern part of the catchment situated in river valleys were discharging groundwaters with the same $^3$H contents as those in the rivers and in the alluvium aquifer. The estimated porosities of the aquifer and the unsaturated zone varied between 0.05 and 0.11 with the thickness of 100 meters for the active water exchange zone of the
springs. The PMPATH model was also applied for delineation spring catchments, which were represented by quite narrow zones of lengths of the first kilometers.

AKNOWLEDGEMENTS

The authors thank prof. W. Kinzelbach for helpful discussions and Fanny Leuenberger for correcting draft of this paper. We are especially grateful to colleagues from the Water Development Department of Cyprus for their support in field work. The research was financially supported by ETH, Zuerich (Internal Research Project TH-22./01-1).

REFERENCES


<table>
<thead>
<tr>
<th></th>
<th>1972-78 (ground- and surface water), 1960-74 (precipitation)</th>
<th>1998-2002</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Quantity of samples</td>
<td>Quantity of sampling points</td>
</tr>
<tr>
<td>Rainfall samples</td>
<td>64*</td>
<td>1</td>
</tr>
<tr>
<td>Rivers</td>
<td>3</td>
<td>3</td>
</tr>
<tr>
<td>Spring samples:</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ophiolitic complex</td>
<td>26</td>
<td>14</td>
</tr>
<tr>
<td>(ultramafic rocks, gabbrros, diabase dykes)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Alluvium aquifer</td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>Borehole samples:</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ophiolitic complex</td>
<td>43</td>
<td>21</td>
</tr>
<tr>
<td>(ultramafic rocks, gabbrros, diabase dykes)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ophiolitic complex</td>
<td>4</td>
<td>3</td>
</tr>
<tr>
<td>(Pillow Lavas)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sedimentary complex</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Alluvium aquifer</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Total</td>
<td>141</td>
<td>43</td>
</tr>
</tbody>
</table>

*64 rainfall samples are collected in 1960-74 for the Prodromos meteostation by IAEA in a collaboration with WDD, analysis are included in the GNIP database ([http://isohis.iaea.org](http://isohis.iaea.org)).
Table 4.2. Correlation between $^3$H concentrations in monthly mixed rainfall samples in some meteostations

<table>
<thead>
<tr>
<th>Location</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Periods of measurements</th>
<th>Koeficient of linear correlation with the $^3$H content in Prodromos</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>In arithmetic form</td>
</tr>
<tr>
<td>Ottawa (Canada)</td>
<td>45°19'12&quot;</td>
<td>75°40'12&quot;</td>
<td>1953-1999</td>
<td>0.86</td>
</tr>
<tr>
<td>Antalya (Turkey)</td>
<td>36°52'48&quot;</td>
<td>30°42'0&quot;</td>
<td>1963-1999</td>
<td>0.73</td>
</tr>
<tr>
<td>Ankara (Turkey)</td>
<td>39°57'00&quot;</td>
<td>32°52'48&quot;</td>
<td>1963-1999</td>
<td>0.98</td>
</tr>
<tr>
<td>Bet Dagan (Israel)</td>
<td>32°00'00&quot;</td>
<td>34°49'12&quot;</td>
<td>1960-1999</td>
<td>0.86</td>
</tr>
<tr>
<td>Alexandria (Egypt)</td>
<td>31°11'00&quot;</td>
<td>29°57'00&quot;</td>
<td>1961-1989</td>
<td>0.98</td>
</tr>
<tr>
<td>Athens (Greece)</td>
<td>37°54'00&quot;</td>
<td>23°43'48&quot;</td>
<td>1960-1991</td>
<td>0.99</td>
</tr>
</tbody>
</table>
Table 4.3. Components of water balance for the Kouris catchment, used for transient groundwater modelling, Mm³ per hydrological year

<table>
<thead>
<tr>
<th>Hydrological year</th>
<th>Rainfall</th>
<th>Surface Runoff</th>
<th>Recharge-Changes in Groundwater storage</th>
<th>Water demand</th>
<th>Baseflow</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td>II</td>
<td>III</td>
<td>IV</td>
<td>V</td>
<td>VI</td>
</tr>
<tr>
<td>984/85</td>
<td>234.0</td>
<td>9.6</td>
<td>39.8</td>
<td>8.1</td>
<td>29.9</td>
</tr>
<tr>
<td>1985/86</td>
<td>174.8</td>
<td>3.8</td>
<td>-13.7</td>
<td>8.3</td>
<td>11.7</td>
</tr>
<tr>
<td>1986/87</td>
<td>293.8</td>
<td>11.2</td>
<td>98.0</td>
<td>8.3</td>
<td>39.8</td>
</tr>
<tr>
<td>1987/88</td>
<td>286.2</td>
<td>17.4</td>
<td>84.1</td>
<td>8.6</td>
<td>52.4</td>
</tr>
<tr>
<td>1988/89</td>
<td>242.2</td>
<td>16.9</td>
<td>40.7</td>
<td>8.9</td>
<td>29.4</td>
</tr>
<tr>
<td>1989/90</td>
<td>149.0</td>
<td>3.2</td>
<td>-38.8</td>
<td>9.3</td>
<td>9.2</td>
</tr>
<tr>
<td>1990/91</td>
<td>131.0</td>
<td>1.6</td>
<td>-55.2</td>
<td>9.6</td>
<td>4.3</td>
</tr>
<tr>
<td>1991/92</td>
<td>329.7</td>
<td>10.5</td>
<td>134.7</td>
<td>9.8</td>
<td>28.4</td>
</tr>
<tr>
<td>1992/93</td>
<td>259.9</td>
<td>9.1</td>
<td>66.1</td>
<td>9.8</td>
<td>31.5</td>
</tr>
<tr>
<td>1993/94</td>
<td>218.8</td>
<td>7.5</td>
<td>26.7</td>
<td>10.0</td>
<td>13.5</td>
</tr>
<tr>
<td>1994/95</td>
<td>234.5</td>
<td>14.1</td>
<td>35.8</td>
<td>10.1</td>
<td>22.6</td>
</tr>
<tr>
<td>1995/96</td>
<td>186.2</td>
<td>3.1</td>
<td>-1.6</td>
<td>10.2</td>
<td>8.7</td>
</tr>
<tr>
<td>1996/97</td>
<td>169.8</td>
<td>3.3</td>
<td>-18.2</td>
<td>10.3</td>
<td>7.5</td>
</tr>
<tr>
<td>1997/98</td>
<td>185.4</td>
<td>2.6</td>
<td>-1.9</td>
<td>10.3</td>
<td>6.7</td>
</tr>
<tr>
<td>1998/99</td>
<td>249.9</td>
<td>15.0</td>
<td>50.3</td>
<td>10.8</td>
<td>11.0</td>
</tr>
<tr>
<td>Average for 15 years</td>
<td>223.0</td>
<td>8.6</td>
<td>29.8</td>
<td>9.5</td>
<td>20.4</td>
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</table>
Table 4.4. Input parameters for PMPATH model for the ophiolitic complex as a result of model calibration

<table>
<thead>
<tr>
<th>Calibrated parameter</th>
<th>Min value for ophiolites</th>
<th>Max value for ophiolites</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Transmissivity, m²/day</strong></td>
<td>Lower layer of the aquifer</td>
<td>3.6</td>
</tr>
<tr>
<td></td>
<td>Upper layer of the aquifer</td>
<td>0.9</td>
</tr>
<tr>
<td><strong>Vertical hydraulic conductivity of the unsaturated zone, m/day</strong></td>
<td>0.5</td>
<td>0.5</td>
</tr>
<tr>
<td><strong>Porosity</strong></td>
<td>Aquifer (both layers)</td>
<td>0.05</td>
</tr>
<tr>
<td></td>
<td>Lower layer of the unsaturated zone</td>
<td>0.05</td>
</tr>
<tr>
<td></td>
<td>Upper layer of the unsaturated zone</td>
<td>0.09</td>
</tr>
<tr>
<td><strong>Thickness, m</strong></td>
<td>Lower layer of the aquifer, m</td>
<td>40</td>
</tr>
<tr>
<td></td>
<td>Upper layer of the aquifer, m</td>
<td>10</td>
</tr>
<tr>
<td></td>
<td>Lower layer of the unsaturated zone, m</td>
<td>20</td>
</tr>
<tr>
<td></td>
<td>Upper layer of the unsaturated zone, m</td>
<td>10</td>
</tr>
</tbody>
</table>

Table 4.5. Simulated residence times for some springs of ophiolitic complex

<table>
<thead>
<tr>
<th>Spring name</th>
<th>Min residence time</th>
<th>Max residence time</th>
<th>Average residence time of distribution</th>
</tr>
</thead>
<tbody>
<tr>
<td>Loumata “A”</td>
<td>18</td>
<td>31</td>
<td>23.4</td>
</tr>
<tr>
<td>Loumata “B”</td>
<td>18</td>
<td>23</td>
<td>20.0</td>
</tr>
<tr>
<td>Loumata “C”</td>
<td>18</td>
<td>27</td>
<td>21.2</td>
</tr>
<tr>
<td>Loumata “E”</td>
<td>17</td>
<td>18</td>
<td>17.9</td>
</tr>
<tr>
<td>Troodos (Spring 21A,B,C)</td>
<td>18</td>
<td>41</td>
<td>22.5</td>
</tr>
<tr>
<td>Seven Sisters</td>
<td>18</td>
<td>23</td>
<td>19.8</td>
</tr>
<tr>
<td>Kalidhonia</td>
<td>18</td>
<td>20</td>
<td>18.9</td>
</tr>
<tr>
<td>Yerokamina</td>
<td>18</td>
<td>20</td>
<td>19.4</td>
</tr>
<tr>
<td>Kephalovrysos (at P.Platres)</td>
<td>17</td>
<td>23</td>
<td>20.4</td>
</tr>
<tr>
<td>Spring 02</td>
<td>17</td>
<td>20</td>
<td>18.2</td>
</tr>
<tr>
<td>Spring 24</td>
<td>17</td>
<td>22</td>
<td>19.1</td>
</tr>
<tr>
<td>Spring 27</td>
<td>18</td>
<td>18</td>
<td>18.0</td>
</tr>
<tr>
<td>Archolochania</td>
<td>18</td>
<td>30</td>
<td>21.4</td>
</tr>
<tr>
<td>Mozoras</td>
<td>20</td>
<td>31</td>
<td>25.2</td>
</tr>
<tr>
<td>Spring 15</td>
<td>18</td>
<td>31</td>
<td>21.9</td>
</tr>
<tr>
<td>Ali-Tsaousi</td>
<td>18</td>
<td>27</td>
<td>21.1</td>
</tr>
<tr>
<td>Spring 14</td>
<td>17</td>
<td>20</td>
<td>19.0</td>
</tr>
<tr>
<td>Spring 26</td>
<td>30</td>
<td>31</td>
<td>30.5</td>
</tr>
<tr>
<td>Vasiliki</td>
<td>13</td>
<td>17</td>
<td>14.2</td>
</tr>
<tr>
<td>Spring 04, Spring 05</td>
<td>17</td>
<td>26</td>
<td>21.3</td>
</tr>
<tr>
<td>Springs at Agios Thodoros</td>
<td>13</td>
<td>16</td>
<td>14.1</td>
</tr>
<tr>
<td>Anastasia</td>
<td>13</td>
<td>32</td>
<td>22.0</td>
</tr>
<tr>
<td>Kaoras</td>
<td>13</td>
<td>17</td>
<td>14.3</td>
</tr>
<tr>
<td>Spring 09</td>
<td>17</td>
<td>27</td>
<td>21.8</td>
</tr>
</tbody>
</table>
Figure 4.1. Simplified geological map of Cyprus showing the location of the Kouris catchment.

Figure 4.2. Geological map of the Kouris catchment with sampling locations: crosses – precipitation, circles – boreholes, triangles - springs, squares – rivers.
Figure 4.3. Input function for $^3$H content in the precipitation over the Kouris catchment; dashed lines limit the range of the uncertainty.

Figure 4.4. Tritium contents in groundwater, discharging from some springs of the ophiolitic complex in 1998-99: empty bars – in dry seasons; grey bars – in wet seasons.
Areal boundary between ophiolites and sediments

stations with stream flow measurements
- spring with a number
- borehole used for irrigation and domestic needs

Constant head boundary (H=0)
Mediterranean Sea

Figure 4.5. Plan view of time-independent input parameters for the transient groundwater flow model. Left – boundary conditions (springs, rivers and boreholes are associated with the third layer); right – areal transmissivity zonation (absolute values refer to the sum of transmissivities of the third and the fourth layers).

Figure 4.6. Regression between annual rainfall and variations in groundwater storage totally for the Kouris catchment for 1984-99.
Figure 4.7. Time-dependent input parameters for the transient groundwater flow model. Left – Recharge; Right – Evapotranspiration (Zone 2 refers to no evapotranspiration for all time periods).

Figure 4.8. Scatter diagram of 210 observed and simulated transient piezometric heads for the basic variant.
Figure 4.9. Examples of comparison between observed (solid lines) and simulated (dashed lines) transient piezometric heads for boreholes 67/76 and 145/90.

Figure 4.10. Comparison of transient baseflows, separated from hydrograms, with those simulated for variants with different storativities.
Figure 4.11. Scatter diagram between observed and simulated $^3$H concentrations (at the uncertainty of the input function).

Figure 4.12. Location of springs with simulated residence times (Table 4.5).
Figure 4.13. Distribution of residence times for Archolochania spring in September 1976 with the concentration 27.5 TU and the average residence time 21.9 years.

Figure 4.14. Delineation of some spring catchments in the ophiolitic complex with PMWIN.
Chapter 5

Estimation of actual evapotranspiration from an alluvial aquifer of the Kouris catchment (Cyprus) using continuous streamflow records

Anastasia Boronina, Philippe Renard, Sergey Golubev & Werner Balderer

Submitted to “Hydrological Processes”
ABSTRACT
The study applies continuous streamflow records in the River Limnatis (Kouris catchment, Cyprus) for calculation of actual evapotranspiration in dry seasons from a alluvial aquifer. Actual evapotranspiration from the alluvial aquifer was reflected by automatic and manual streamflow observations, thus it was possible to estimate this water balance component on a daily basis and extrapolate estimated values over the whole year. The paper contributes in quantification of a regional water balance of the Kouris catchment, which is currently experiencing water scarcity problems due to its partially semi-arid climate and over-exploitation of water resources.

KEY WORDS: evapotranspiration, Cyprus, alluvium, streamflow observations, water balance
5.1. INTRODUCTION

The Kouris catchment is currently experiencing water scarcity problems due to its partially semiarid climate, growing population and a lack of quantitative knowledge about the area.

The catchment encompasses an area of 300 km² on the Southern side of the Troodos Mountains and it is bounded on the South by the Mediterranean sea (Fig. 5.1). The topography is typical for mountains regions: an altitude is changing from the sea level to 2000 meters within a distance of 30 kilometres; local slopes can reach 80 % and a relief is controlled by river valleys. There are three perennial rivers in the catchment: Kryos, Kouris and Limnatis (Fig. 5.2), containing spring water in dry seasons. The River Kouris is the largest in Cyprus and has had an average annual streamflow of 36 Mm³/year during the last 30 years; its downstream reservoir (Kouris Dam) has the highest storage capacity in Cyprus (110 Mm³) and it was built for drinking water supply of the whole island.

The area includes two main geological parts: an ophiolitic complex in the North and overlying sediments in the South (Fig. 5.2). The highly fractured ophiolitic sequence is ranging from Pillow Lavas at the stratigraphic top through a Sheeted Dyke Complex, Gabbros and Dunites to tectonised Harzburgites and Serpentinites at the North-Western part of the catchment. Plutonic and Sheeted dykes complexes contain the major groundwater resources of Cyprus; the water is stored in the fractured and altered zones of harzburgites, dunites, gabbros and the diabase dykes.

The Sedimentary complex (Fig. 5.2) consists of chalks, marls, calcarenites of the Lefkara and gypsum sediments of the Pakhna formations. These rocks contain comparatively small groundwater resources due to high evaporation and low both permeability and rainfall.

The alluvium aquifer (sands, gravels) constrains river beds of Kryos, Kouris and Limnatis in the southern part of the Kouris catchment. It is discontinuous, narrow (the width up to 50 meters) and has thickness up to 10-15 meters. It contains large amount of water, comparatively to the whole sedimentary complex.
Despite the fact that the quantitative knowledge of water balance of the Kouris catchment is essential for Cyprus, only few attempts were made in estimation of its water balance components.

At the stage of design of the Kouris dam, transient streamflow rates of the Kouris catchment were calculated by Jacovides et al. (1982) with MERO model, which has been used in Cyprus since 1967 (UNDP/FAO, 1970). That was a simple reservoir model with input rainfall, potential evaporation, conceptual characteristics of the watershed and output streamflows. Despite quite a good fit of observed and simulated streamflows, the model had certain disadvantages, of which, in our opinion, the biggest were: rough spatial discretization, assumption of no evaporation during dry seasons, absence of data for discharge calibration from different reservoirs, assumption of no human influence.

Recent attempts in water balance calculations were made by Boronina et al. (2003a). In that study recharge was estimated, at first, as a result of groundwater flow model calibration by piezometrical heads and baseflows; later it was additionally calibrated with deuterium concentrations in the aquifer (Boronina et al., 2003b). The steady state groundwater flow model, together with hydrogramm separation methods, resulted in the estimates of the regional water balance, presented in Table 5.1. In a next step, the authors reproduced temporal baseflow and piezometry variations by a transient groundwater flow model and obtained a storage coefficient as a result of a model calibration. However, these calculations had all limitations related to numerical groundwater flow modelling, where recharge and evapotranspiration were input data and had to be estimated by the means of other methods.

The present studies aim to calculate actual evapotranspiration from river valleys of the Kouris catchment for a dry season and, thus, make a contribution in quantifying an important water balance component of the region.

Estimation of actual evapotranspiration under semi-arid climate has been for decades a challenging scientific task. Often in mathematical models, these estimates are based on potential evapotranspiration values, measured at meteorological stations (see Singh, 1995 for a review). However, these methods can cause serious errors in simulations of semi-arid hydrology.

Figure 5.3 illustrates a danger of direct application of potential Pan A evaporation values for the Kouris catchment. It presents both potential Pan A evaporation and rainfall, averaged for the Kouris catchment for the hydrological year.
The assumption, that evapotranspiration takes place only in the case of sufficient soil water content and during the rainy season, results in "0" actual evapotranspiration for the period April-September. This error (since evapotranspiration actually occurs in dry seasons from groundwater table) often leads to wrong water balance calculation, particularly, to an overestimation of recharge.

Figure 5.4 shows (in black) locations of areas with Normalized Difference Vegetation Index (NDVI) of more than 0.3, calculated from a LANDSAT 5 image (Bands 3 and 4) from July 1987. Generally NDVI is an indicator for the presence and conditions of green vegetation. Thus, a high NDVI in the North is related to areas heavily irrigated in July, however, narrow areas with high NDVI along the rivers correspond to riparian vegetation in areas covered by alluvium. We suggest, that these are areas, where evapotranspiration occurs during dry seasons. The trees (Populus nigra, Plataneus Orientalis, Alnus Orientalis,) can uptake each from 10 to 100 litres per day (informal discussion with Dr.Lambs, UMR Laboratoire Dynamique de la Biodiversity, Touluse). Additionally, in these areas, the water level is shallow enough (2-3 meters) to allow direct evaporation from a water table.

There are number of methods for estimation of actual evapotranspiration in dry seasons. Lysimeter studies allowed the quantification of evapotranspiration directly, however they can give only small scale values, often not representative. For example, the estimates of recharge in the sedimentary complex of Cyprus with lysimeters resulted in, most probably, wrong numbers (Kitching et al., 1980). Transpiration by a single tree can be calculated by sap flow measurements (Cermak and Prax, 2001; Granier et al., 1996). Other methods (Bowen ratio – Bowen (1926), Eddy-correlation - Swinbank (1951)) require expensive equipment, though, still estimate evapotranspiration only in one location. Algorithms, applying remote sensing (Bastiaanssen et al., 1998; Roerink et al., 2000) on the contrary, resulted in spatial distribution of real evapotranspiration, however its absolute values, especially in mountain areas, have unsatisfactory uncertainty.

The aim of this study is to use continuous streamflow records in the River Limnatis in order to calculate actual evapotranspiration in dry seasons from the alluvial aquifer. We would like to estimate the integral effect of transpiration and direct evaporation from the water table for the whole catchment. The idea of calculating evapotranspiration from streamflow and piezometry records has often appeared in scientific publications. Some studies (White, 1932; Meyboom, 1967; Bauer et al., 2003)
used continuous records of piezometric heads in boreholes, located in phreatic aquifers, to estimate daily evapotranspiration. In other studies (Tschinkel, 1963; Rorabaugh, 1964; Daniel, 1976) daily streamflow records were used to calculate evapotranspiration from a recession curve. Requirements of storativities and transmissivities, integrated over the basin, however, brought additional uncertainties in those methods. Often researchers (see Wittenberg and Sivapalan (1999), Griffiths and Clausen (1997) and Tallaksen (1995) as examples) tried to find actual evapotranspiration among other water balance components through a calibration of hydrological models in daily or monthly time steps. Nyholm et al. (2003) showed the effect of evapotranspiration in continuous inter-daily streamflow records in one alluvium catchment in Denmark; however the authors were not interested in absolute values and only took them as a component of uncertainty of daily streamflow.

The first part of our paper deals with comparison of different instruments, used for streamflow measurements in the River Limnatis, and calculation of a calibration curve for an electronic equipment.

In the second part, after a discussion of the origin of streamflow variations, a daily actual evapotranspiration for several days is estimated.

In the last part, a linear correlation between estimated values and daily potential Pan A evaporation ratios are used to calculate total actual evapotranspiration for the Kouris catchment during a dry season (April-September) of the year 1989.

5.2. MATERIALS AND METHODS

In this study, we used continuous measurements from the flow-gauging station N 1967-654-70 (Water Development Department observation network), constructed downstream the River Limnatis (location R4 in Figure 5.2) and equipped with a mechanical water level recorder. Data for the period of studies (01.10.00-30.11.01) were digitized at an interval of a half an hour. Additionally, at the location R4 two pressure and temperature loggers were set up for measurements at an interval of 10 minutes for the period 10:00 30.10.2001 – 9:30 26.11.2001. One of the loggers was used to record atmospheric pressure and temperature, while the other was measuring water pressure in the river. Additionally, some manual measurements of streamflow were conducted at the locations R1, R2, R3, R4 (Fig. 5.2) with the universal anemometer. Manual
measurements were conducted irregularly up to three times a day to calibrate data logger records and to get an idea about spatial streamflow distribution.

The accuracy of manual measurements was generally quite low (approximately 20%), due to the complex relief of a river bed and high streamflow variations at river cross-sections, but also because measurement sites were not specially prepared (except the location R4). However, temporal variations of streamflow rates obtained in all locations had higher accuracy than absolute values of streamflow, because temporal variations in a single location were not influenced by spatial heterogeneities.

Continuous measurements by datalogger and by a mechanical recorder gave the higher accuracy than manual measurements, because, first, they were based on water level in a well, connected to the river channel, thus, reflecting an average effect of streamflow changes; second, the measurement site for continuous measurement was specially prepared and covered by concrete.

To have an idea about the distribution of rainfall and potential evaporation over the Kouris catchment (Fig. 5.3), we used daily rainfall and potential Pan A evaporation records from the meteorological stations N 320 (Saittas, altitude 640 m a.s.l.) and N 377 (Agros, altitude 1015 m a.s.l.) – Figure 5.2. The potential evaporation data from the two stations were also applied to obtain a linear regression between Pan A evaporation and estimated actual evapotranspiration (the period 30.10.01-5.11.01) and to calculate annual evapotranspiration for April-September of the year 1989. This year was selected for calculations because some water balance components for the Kouris catchment of that period were already known (Table 5.1, Boronina et al., 2003a; Boronina et al., 2003b).

5.3. RESULTS

Water levels at the river flow-gauging station were calculated by the following formula from logger pressure data:

$$ h = \frac{100000 \times (P_{\text{water}} - P_{\text{atm}})}{\rho \times g}, $$

(5.1)

$h$ [mm] – water levels; $P_{\text{water}}$ [mBar] and $P_{\text{atm}}$ [mBar] – measured water and atmospheric pressures, respectively; $g$ – gravitation constant, $g = 9.81 \text{ m/s}^2$; $\rho$ – density of water, $\rho = 1000 \text{ kg/m}^3$. 
Comparison of the records for the period 30.10 – 05.11.2001 of the datalogger and the digitized data from the mechanical recorder are presented in Figure 5.5. Both records show obvious daily cycles of water levels in the river, the same tendencies and average absolute values, however the mechanical records look smoother than the electronic ones. High-frequency pressure variations, visible from digital data, were filtered out by a mechanical device. This fact might have a simple explanation, that a mechanical gauge was not as sensitive to small and rapid variations, as electronic equipment. Therefore, for the analysis of comparatively small pressure changes, for example due to evapotranspiration, only electronic datalogger was appropriate. Finally, the comparison proved that daily cycles observed in electronic records were not artificial effect of the logger itself.

The mechanical records, digitized during one year (Fig. 5.6), showed the following tendencies, which should be the same for the datalogger records.

- The cycles of streamflow were periodic, with a period of 24 hours, minimums at 22:00-02:00 and maximums at 11:00-15:00; that showed that they were not induced by the Moon and Earth tides.
- The daily variations of water levels in absence of rainfall existed only during dry seasons, which indicated their relation to climatic conditions. In wet months, when there is always water available in soils and in atmosphere, alluvial aquifer is not affected by evapotranspiration.
- Those streamflow cycles were independent on days of a week and appeared even after irrigation stopped; this fact suggested, that they were not results of pumping. Additionally, irrigation system of the River Limnatis was based on diverting of stream water which could not cause periodic, but rather continuous influences.

To summarise information about daily cycles presented in Figure 5.6, one can conclude, that they were caused by evapotranspiration from the alluvial aquifer, rather than by any other factors.

In order to calculate daily amount of water lost from the river due to evapotranspiration, water levels at the gauging station had to be transferred to streamflow rates. For this aim we constructed a calibration curve (Fig. 5.7), based on 33 manual measurements by the anemometer for the period 30.10 – 26.11.2001. The curve can be described by a regression equation:
\[ Q = 0.0003 \times h^{2.81} \]  

(5.2)

with R-squared value equal to 0.97, where \( Q \) represents streamflow rate \([m^3/\text{hour}]\) at R4 and \( h \) \([\text{mm}]\) is the water level, calculated from the logger measurements by Equation 5.1.

For the calculation of daily actual evapotranspiration, hourly averaged pressure data were used in order to filter out high-frequency variations. The estimates were made by the simple formula:

\[ E_{ \text{daily} } = \frac{2Ah_{\text{hours}}}{1200} (Q_{\text{max}} - Q_i) \]  

(5.3)

where \( E_{ \text{daily} } \) represents the daily amount of water, lost from the River Limnatis due to evapotranspiration \([m^3]\); \( Q_{\text{max}} [m^3/\text{hour}] \) is the daily maximum flow in the river (in all days it was between 12:00 and 15:00; \( Q_i [m^3/\text{hour}] \) – averaged flow for every hour of the day.

The estimated evapotranspiration amounts have positive linear correlation \((k=0.8)\) with potential Pan A evaporation rates, averaged for the meteorological stations NN 320, 377 – Figure 5.8. This fact further supports the idea that evapotranspiration was the main cause of daily streamflow variations. The linear regression for daily values is described by the equation:

\[ E_{ \text{daily} }^{\text{alluv}} = (0.52 \pm 0.11)E_{\text{pot}}^{\text{alluv}} + (0.33 \pm 0.75) \]  

(5.4)

there \( E_{ \text{daily} }^{\text{alluv}} \) - real evapotranspiration from the alluvium aquifer of the Limnatis subcatchment, 1000 x \( m^3 \); \( E_{\text{pot}}^{\text{alluv}} \) - potential Pan A evaporation, averaged for the stations N320 and N377.

Equation 5.4 was used to calculate the actual evapotranspiration from the alluvium aquifer during the whole dry season of the hydrological year 1988/89 (April-September, 1989). In these estimates we assumed, that the alluvium aquifer always contained water available for evapotranspiration; this assumption did not contradict to field observations. Actual evapotranspiration from the alluvium of the Limnatis subcatchment for the dry season of the year 1988/89 calculated by Equation 5.4, resulted in the range from 550,000 \( m^3 \) to 1,111,000 \( m^3 \) with 710,000 \( m^3 \) as the best estimate. Under the assumption of approximately equal evapotranspiration rates for Kryos, Kouris and Limnatis subcatchments, the actual evapotranspiration from the
alluvium aquifer of the Kouris catchment for the selected period was at the range 1.7-3.3 Mm³.

According to the field observations and Figure 5.4, the width of the area of the alluvial aquifer, where evapotranspiration occurred, was 50 meters in average, although a length of the evapotranspiration zone remained unclear.

Streamflow records at the location R4 (Figures 5.5, 5.6) show approximately 12-hour shift with usual daily evapotranspiration cycle (starting about half an hour after the sunrise in the morning, reaching a maximum at solar noon and continuing a little after the sunset). On the other hand, at the locations R1, R2, R3, times of streamflow extremums look similar to a daily evapotranspiration cycle (Fig. 5.9). Assuming that (according to anemometer measurements) an average stream velocity was 0.5 m/sec and evapotranspiration mostly took place between R1 and R3, at a distance of 5-8 km from R4, a 9-13 hour shift of streamflow extremums at R4 relative to a daily evapotranspiration cycle could be explained. On the other hand this shift can also be partly caused by river bed filtration properties at the location R4.

Assuming 500,000 m² (50 m x 10,000 m) as an area of evapotranspiration from the alluvium of the Limnatis subcatchment, one obtains daily rates of evapotranspiration between 1 and 5 mm depending on the daily weather for the period of observations.

5.4. DISCUSSION

Actual evapotranspiration from an alluvial aquifer under semi-arid climate is an important term for water balance required by all kind of hydrological/hydrogeological models. However, estimation of this term is rather difficult, and very often even an accuracy 200%-300% is desirable.

For the Kouris catchment it was possible to estimate the actual evapotranspiration from the alluvial aquifer by streamflow observations. Uncertainties of estimates resulted from the following factors.

1. **Uncertainties of the calibration formula (Equation 5.2) relating stream fluxes and water levels in the gauging station.** These uncertainties can reach 30 % (Figure 5.7) and they are the results of non-uniformity and local turbulences of streamflow. These uncertainties can be reduced by more detailed (in time and in a stream crosssection) manual measurements of streamflow.
2. **Neglecting daily amounts of discharge losses from the river.** The occurrence of discharge from the river to the alluvial aquifer is supported by isotope and chemical data (Boronina et al., 2003b; Boronina et al., 2003c); however, it most probably has a constant rate, independent on the time of a day. Thus, it doesn’t influence inter-daily variations of streamflow, and doesn’t need to be taken into account for the calculation of daily evapotranspiration.

3. **Neglecting streamflow variations resulted from recession of streamflow.** The recession of the streamflow is caused, on one hand, by changing the amounts of surface water component in streamflow after rainfall, while on the other hand, it is a result of transient regime of groundwater discharge due to the storage of the ophiolitic aquifer. This recession is clearly visible in Figure 5.6 – it is the most steep in January and the least – in April; thus, for those months this term must be taken into account for the calculations of the actual evapotranspiration. However, the recession was not observed for the period of measurements (Figure 5.5) because of rather small influences of last rainfalls and aquifer storage at this period of the year (the end of dry season). Thus, for this period, we suggest, that neglecting of streamflow recession does not bring uncertainties in the estimates of the actual evapotranspiration.

4. **Uncertainties of the regression equation (Equation 5.4) for extrapolation of the evapotranspiration from the period of observation to the whole dry season.** Uncertainties of Equation 5.4 were propagated in the final range of uncertainty of the actual evapotranspiration (see RESULTS section). Those resulted from the fact that local Pan evaporation data do not necessary have high linear correlation with the regional actual evapotranspiration. Although the calculation results, in our opinion, are essential, it would be more precise to continue measurements of streamflow for a longer period of observations and to use them directly for the calculations of actual evapotranspiration.

The method described can be used for the estimation of regional amounts of evaporated water in the Kouris catchment and similar semi-arid areas. However, it doesn’t allow the construction of spatial pattern of evapotranspiration and it gives only approximate ideas, where evapotranspiration takes place. Also, the assumption, that the amount of water evaporated from all three river valleys were equal, works only under certain limitations. To have more precise information, one will have to observe the rivers Kryos and Kouris as well.
We do not make a difference between water uptaken by trees and water directly evaporated from a shallow water table – both of these processes affect streamflow variations.

Finally, the estimated value for the Kouris catchment (1.7 - 3.3 Mm³) falls within the range of previously estimated uncertainty of recharge (Boronina et al., 2003a). It is rather small in a comparison with other water balance components and could be even neglected at the present stage of modelling of the Kouris catchment. However, simply knowing an order of magnitude for this value is important for understanding the regional water balance; it can be used later in more detailed studies of the area.

5.5. CONCLUSIONS

Inter-daily variations of streamflow in the River Limnatis (the Kouris catchment of Cyprus) were caused by transpiration of water by trees in the river valley and direct evaporation from a water table of the alluvial aquifer. Continuous streamflow records were used to estimate daily actual evapotranspiration rates from the alluvial aquifer. These values had a high positive linear correlation with average daily Pan A potential evaporation rates (correlation coefficient is 0.8), that allowed us to calculate an annual actual evapotranspiration from the alluvial aquifer (the hydrological year 1988/89 was taken as an example). The calculations of a regional (for the whole Kouris catchment) actual evapotranspiration from the alluvial aquifer resulted in 1.7 ± 3.3 Mm³ for April-September of the year 1989. The method presented here could be applied to estimate regional daily evapotranspiration in the semi-arid areas, similar to the Kouris catchment.

ACKNOWLEDGMENTS

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REFERENCES


Table 5.1. Water balance for the Kouris catchment, with all values in Mm\(^3\) per year for 1988/1989

<table>
<thead>
<tr>
<th>Inflow</th>
<th>Outflow</th>
</tr>
</thead>
<tbody>
<tr>
<td>River</td>
<td>Groundwater outflow into the Mediterranean Sea</td>
</tr>
<tr>
<td>Leakage</td>
<td>Mediterranean Sea</td>
</tr>
<tr>
<td>Recharge</td>
<td>Springs Discharge</td>
</tr>
<tr>
<td></td>
<td>Groundwater extraction</td>
</tr>
<tr>
<td></td>
<td>Discharge in the River Kryos</td>
</tr>
<tr>
<td></td>
<td>Discharge in the River Kouris</td>
</tr>
<tr>
<td></td>
<td>Discharge in the River Limnatis</td>
</tr>
<tr>
<td></td>
<td>Sub-total all rivers</td>
</tr>
<tr>
<td>Total</td>
<td>Total</td>
</tr>
</tbody>
</table>
Figure 5.1. Simplified geological map of Cyprus showing the location of the Kouris catchment.

Figure 5.2. Simplified geological map of the Kouris catchment with locations of flow, rainfall and evaporation measurements; the main evaporation losses occur in the river valleys in the sedimentary part of the catchment.
Figure 5.3. Monthly rainfall (stars) and potential Pan A evaporation (triangles) in mm, averaged for the meteostations N377 and N 320 for the hydrological year 1988/89.

Figure 5.4. NDVI map of the Kouris catchment, derived from a Landsat 5 satellite image, taken in July, 1987; black colour represent areas with NDVI more than 0.3.
Figure 5.5. Comparison of records from a mechanical gauge and datalogger for the period 30.10.01-05.11.01 at location R4.

Figure 5.6. Continuous records from the mechanical device (location R4) for seven days in January, March and April, digitised every hour.
Figure 5.7. Calibration curve for calculation of river flow in the River Limnatis (location R4) from the logger data.

Figure 5.8. Regression between daily evapotranspiration from the alluvium of the Limnatis subcatchment and potential evaporation of the Kouris catchment
Figure 5.9. Manual measurements of streamflow at the locations R1 (circles), R2 (triangles) and R3 (stars); dashed lines represent approximated daily cycles of streamflow for all locations.
Chapter 6

Conclusions
In the Kouris catchment, water conflicts are rising dramatically due to partially semi-arid climate and water over-exploitation. The region has suffered a sustained long-term reduction of surface/ground water storage. The hydrogeology of the catchment is rather complex; it is characterised by high heterogeneity of rock properties, major evapotranspiration losses during the dry season, and strong coupling of surface and ground water.

In this chapter, we present a synthesis of all results obtained during the PhD project; we also discuss the limitations of our studies, reliability of conclusions and give recommendations for further investigations in the Kouris catchment.

6.1. STEADY STATE AND TRANSIENT WATER BALANCE OF THE KOURIS CATCHMENT

According to the numerical modelling and isotope results, under the steady state assumptions for the hydrological year 1988/89 (01.10.88-30.09.89), all recharged water from precipitation (30.2 Mm$^3$) is discharging via springs (4.1 Mm$^3$) and via Kouris (10.4 Mm$^3$) Kryos (1.3 Mm$^3$) and Limnatis (12.7 Mm$^3$) baseflows. Groundwater outflow to the Mediterranean sea is negligible. Groundwater extraction for that year was equal to 1.7 Mm$^3$, however water demands have increased in the last years (Chapter 2, Figure 2.10).

The estimated range of the steady state recharge for the year 1988/89 is between 12% and 16% of the total annual rainfall, which amounted to 100-150 mm for that year. These values are the results of numerical simulations. Specifying recharge rates of less than 12 % led to insufficient water supply for the estimated baseflow, whereas recharge of 20 % already produced unacceptable changes in the predicted ratio between flows of the 3 rivers of the catchment. The calibration of the steady state recharge rate with deuterium (Chapter 3) resulted in an optimal recharge rate, equal to 90-140 mm, which is in agreement with the previous numerical flow model.

Most of the regional recharge (90 %) occurs in the Plutonic and Sheeted Dykes complexes; this is a result on one hand, of high fracturing of these rocks and, on the other hand, of large effective precipitation at high surface elevations.

The evapotranspiration takes places during the dry season over many places of the Kouris catchment. The evaporated water is extracted from groundwater storage, thus must be taken into account in the groundwater balance calculations. Even several months after
the last rainfall, there is a large area covered by green vegetation (Fig. 6.1), which includes irrigated crops, pine forest and riparian vegetation. For the year 1988/89, evaporation from the irrigated lands (see indications in Figure 6.1) was estimated to be not more than 8 Mm$^3$ (amount of water demand for irrigation in the Kouris catchment – Chapter 2). For that year, evaporation during dry season from the groundwater table and transpiration by riparian vegetation was calculated to be between 1.7 and 3.3 Mm$^3$ (Chapter 5).

![Figure 6.1. NDVI map of the Kouris catchment, derived from Bands 3 and 4 of a LANDSAT 5 satellite image, taken in July, 1987.]

Transient water balance components for the years 1984-1999 obtained by numerical modelling (Chapter 4) are presented in Figure 6.2. They demonstrate the high buffering role of the groundwater storage.
6.2. ORIGIN AND RESIDENCE TIMES OF GROUNDWATERS IN DIFFERENT AQUIFERS

The ophiolitic aquifer can be vertically divided in two major zones: the first 50-100 meters from the surface - the zone of active water exchange and to the depth of several hundreds of meters the zone of slow water exchange. In the zone of an active water exchange, most of the groundwaters do not travel long distances. Normally the waters, recharged in the ophiolites, are discharging at distances not more than the first kilometres in the rivers and in the springs. The reason for this is, on one hand, a high density of fractures resulting in a large number of springs, while, on the other hand, the shape of the catchment, steep river valleys and dense and wide-spread river network influencing the piezometry.

The groundwaters of the upper zone have average residence times of 1-25 years, where “young” waters appear in springs in wet seasons; in some areas they are already contaminated from the surface. Springs of ultramafic rocks are usually discharging groundwater of high residence times, due to smaller conductivity, than those in the Gabbros and the Sheeted Dykes.

The groundwaters of the slow water exchange zone are not considerably discharging to rivers and springs, but they are often tapped by deep (more than 100 meters) boreholes in dry seasons. These waters have residence times more than 48 years with uncertain recharge areas and travel distances. In case of pumping water from those deep reservoirs, groundwater resources are only very slowly renewable.
The groundwaters in the Pillow Lavas have high residence times and are probably originating from the plutonic and diabase rocks. The fact that the aquifer is strongly confined support the idea of their remote origin at high altitudes.

Concerning the sedimentary aquifer (the Lefkara and the Pakhna formations), stable isotope studies (Chapter 3) indicated, that the groundwaters are originated from local recharge. Small recharge rates in combination with small rock conductivities resulted in high residence times (more, than 48 years). Of course, we can not completely exclude the possibility of deep ground water flow from the ophiolites to the sediments. However, most of recharged water in the ophiolites still participates in an active water exchange (and discharges in river and springs of the ophiolitic complex), rather than in slow water exchange.

The alluvium aquifer in the southern part of basin contains water, originated from the Troodos Mountains and infiltrated in river beds, rather than from the surrounding sediments. This hypothesis is supported by estimated residence times, stable isotope and chemical macro-component studies.

6.3. CHEMICAL COMPOSITION OF GROUNDWATERS AND FACTORS INFLUENCING IT

Groundwater in the ultramafic rocks, the Gabbros and the Sheeted Dykes is generally of CaMg-HCO₃ and Na-HCO₃ types with low to moderate salinities (200-600 mg/l), cation and anion compositions vary considerably with the local lithology. The Pillow Lavas contain the water with higher TDS (500 mg/l to 950 mg/l) and, within the anion content, the SO₄⁻ ion plays the major role. Groundwaters in the consolidated sedimentary rocks are generally moderately mineralised to brackish, the salinity varies between 600 and 1600 mg/l. They are of Na-Ca-HCO₃-SO₄, Na-Ca-SO₄-HCO₃ or Ca-Na-HCO₃-Cl types. Stable isotope studies showed (Chapter 3) that the high mineralisation of the groundwaters in the sedimentary complex and in the Pillow Lavas are the result of water/rock interaction, rather than evaporation. Groundwaters of the alluvial aquifer are less mineralised (500-600 mg/l), of Ca-Mg-HCO₃, Mg-Ca-HCO₃ or Mg-Na-HCO₃ types. Their salt content is caused by a mixture of waters of different origins from the ophiolitic aquifer, that have been partially evaporated and transpirated by vegetation. Evaporation is clearly reflected by stable isotope content of these groundwaters.
6.4. HYDROGEOLOGICAL PROPERTIES OF THE AREA

Transmissivities
To get an initial idea about transmissivity distribution within the Kouris catchment, 123 single pumping tests were interpreted (Table 2.2, Chapter 2). These tests were conducted only in four types of rocks: the Gabbros and the Sheeted Dykes (66 boreholes), the sedimentary aquifer (8 boreholes), the alluvial aquifer (2 boreholes). Thus, the Gabbros and the Sheeted Dykes, which were considered as one combined aquifer were supported by information at most. Experimental variogram indicates no spatial correlation in transmissivity data. Thus, to obtain preliminary values of transmissivity for these rock types, their geometrical mean values were multiplied by the ratio of successful boreholes to all boreholes (Table 2.1, Chapter 2). The values, resulted from this preliminary analysis are presented in Table 6.1.

The calibration of the numerical model allowed obtaining a set of transmissivities (Table 6.1), which were similar for the Gabbros, Sheeted Dykes and the sedimentary rocks to those, estimated from the field studies. The transmissivity of the ultramafic rocks was not possible to check, because there were no boreholes in that area. However, from the model sensitivity analysis, the transmissivity of these rocks can vary in a very little range: 4.2-4.8 m²/day. The ultramafic rocks have the highest degree of heterogeneity in the whole catchment: there are few big springs related to faults, but the surrounding rocks are almost impermeable. However, they have little importance for water management, since they cover only around 5 % of the area on the top of the catchment, no boreholes are drilled there and most of recharged water is discharging in local springs. These results are comparable, at least in terms of contrast between the hydraulic conductivities of different lithologies, with the pumping test estimates of Dewandel et al. (2002) for the ophiolitic rocks of Oman.
Table 6.1. Comparison of transmissivities estimated from pumping tests and from calibration of the numerical model.

<table>
<thead>
<tr>
<th>Type of rocks</th>
<th>Transmissivity resulted from pumping test interpretation, m²/day</th>
<th>Transmissivity resulted from numerical model calibration, m²/day</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ultramafic rocks (harzburgites, dunites, serpentinites)</td>
<td>-</td>
<td>4.5</td>
</tr>
<tr>
<td>Gabbro and Sheeted Dykes</td>
<td>18</td>
<td>16</td>
</tr>
<tr>
<td>Pillow Lavas</td>
<td>-</td>
<td>100</td>
</tr>
<tr>
<td>Sediments (chalks, marls, calcarenites, limestones)</td>
<td>3</td>
<td>1</td>
</tr>
<tr>
<td>Alluvium</td>
<td>30-250</td>
<td>25</td>
</tr>
</tbody>
</table>

The transmissivity of the Pillow Lavas as a result of calibration of the numerical model (Table 6.1) seems to be overestimated, even though the zones of high conductivity might exist there. The reason of overestimation is probably too high input recharge in that area. In our model, we did not have the possibility to estimate the input recharge for every zone; we only estimated its total amount and distributed it in four zones, bearing in mind increasing of recharge rate with altitude and low recharge over the sediments due to their small permeability. At the present stage, after 3.5 years of study of the Kouris catchment, we think, that recharge in the Pillow Lavas was overestimated as well as transmissivity, but we do not have information to check this idea. This will not influence, however, a regional water balance, since the area, covered by the Pillow Lavas is very small in comparison with the catchment and there is only few groundwater flowing via these rocks because of the shape of the catchment and the low transmissivity of the sediments.

The transmissivity of the alluvium seems to be slightly underestimated (informal communication with colleagues from the Hydrology Division of WDD of Cyprus). The reason of underestimation might be the coarse spatial discretisation of the model (250m X 250m), while the actual width of the alluvial aquifer hardly extends 50 meters. However in the regional water balance calculations, underestimation of transmissivity of the alluvium is compensated by a larger extent of the modelled area covered by the alluvium; thus this doesn’t bring main errors in the simulations.

**Storage coefficient**

We have only 2 experimental values (0.004 and 0.001) of storativity in the Kouris catchment, which were derived from the interpretation of two pumping tests in the Gabbro, in the north-eastern part of the Kouris catchment. These values can’t be considered as
representative for the whole catchment, neither even for the Gabbros, however, we consider that at least an order of magnitude is correct.

The model calibrations by piezometry in boreholes and by baseflow fluxes resulted in very different storage parameters, with both calibrations, sensitive to storativities. To obtain fluxes, compatible with observed (separated from streamflow) baseflows, we had to input storativity, distributed in the four layers in following manner (from the top): 0.0025, 0.0015, 0.005 and 0.01. To obtain a better fit of the transient piezometry, it would be necessary to increase the storativities at least twice, which would worsen considerably the fit of the baseflows.

Storage parameters controlling transient piezometry and baseflows might be simply different from each other. Temporal variations of baseflows are caused by temporal changes of springs discharges in the ophiolitic complex. Thus, baseflows in rivers rather represent integral storage of the regional confined aquifer. On the contrary, boreholes probably exploit the most productive parts of the aquifer with a long history of pumping (which means that the aquifer may not be confined in the vicinity of the boreholes). Logically, one can expect bigger storativity from boreholes observations, than from baseflows. In our opinion, baseflows reflect a more representative storage for the catchment, while storativity, estimated from piezometry observations is rather the result of local heterogeneities and artificial disturbances.

**Effective porosity**

![Figure 6.3. Scatter diagram between observed and simulated $^3$H concentrations.](image)
Absolute values of effective porosities were obtained from the calibration of the transient 3-Dimensional MODPATH model with tritium data and they are presented in Table 6.2. These values are associated with the first from the surface zone of approximately 100 meters, related to the zone of an active water exchange. Increasing or decreasing input porosities by factor “3” will result in wrong output tritium concentrations (Fig. 6.3); thus for the fractured aquifer we were able to achieve results, varying in a quite a narrow range. However, it was not possible to obtain any field information about distribution of porosities at depth; neither there are constraints to model a groundwater flow/transport below the zone of an active exchange. The lower part of the aquifer will definitely influence tritium transport in the upper zone, but, at this stage we can’t estimate, how strong is this influence.

Table 6.2. Input parameters for PMPATH model for the ophiolitic complex as a result of model calibration (Chapter 4).

<table>
<thead>
<tr>
<th>Calibrated parameter</th>
<th>Min value for ophiolites</th>
<th>Max value for ophiolites</th>
</tr>
</thead>
<tbody>
<tr>
<td>Transmissivity, m²/day</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lower layer of the aquifer</td>
<td>3.6</td>
<td>12.9</td>
</tr>
<tr>
<td>Upper layer of the aquifer</td>
<td>0.9</td>
<td>3.2</td>
</tr>
<tr>
<td>Vertical hydraulic conductivity of the unsaturated zone, m/day</td>
<td>0.5</td>
<td>0.5</td>
</tr>
<tr>
<td>Porosity</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Aquifer (both layers)</td>
<td>0.05</td>
<td>0.06</td>
</tr>
<tr>
<td>Lower layer of the unsaturated zone</td>
<td>0.05</td>
<td>0.05</td>
</tr>
<tr>
<td>Upper layer of the unsaturated zone</td>
<td>0.09</td>
<td>0.11</td>
</tr>
<tr>
<td>Thickness, m</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lower layer of the aquifer, m</td>
<td>40</td>
<td>40</td>
</tr>
<tr>
<td>Upper layer of the aquifer, m</td>
<td>10</td>
<td>10</td>
</tr>
<tr>
<td>Lower layer of the unsaturated zone, m</td>
<td>20</td>
<td>60</td>
</tr>
<tr>
<td>Upper layer of the unsaturated zone, m</td>
<td>10</td>
<td>10</td>
</tr>
</tbody>
</table>

6.5. DISCUSSION OF ISOTOPE METHODS
APPLIED IN THE KOURIS CATCHMENT

A large amount of isotope analysis was done in the Kouris catchment. Only during the current PhD project 176 water samples were analyzed for $^3$H and 224 - for stable isotopes (part of the samples were collected by diploma students of ETH Ulrich Jörin, Andreas Moll, Simon Oertli and Mathias Steiner). Furthermore, 167 water isotope results for the Kouris catchment were available from Jacovides (1979); additionally we used some isotope data from surrounding areas (Verhagen et al., 1991). The main questions, that one can ask: looking back, what would be the most optimal way (concerning new knowledge
obtained, efforts made, time and money spent) for isotope investigations in these types of 
area?

**Tritium**
The last years groundwater sampling for \(^3\)H in the plutonic and intrusive rocks did not 
provide much new information. It rather proved the fact, that the Troodos ophiolites are 
highly heterogeneous, but it was known before, from the data of 1970's (Jakovides, 1979). 
Furthermore, because of decreasing tritium content in atmosphere and tritium decay, 
results from the last years were even more uncertain, than before.

On the contrary, sampling other types of rock (the sediments, the Pillow Lavas, the 
alluvium) supported our hypothesis about origin of water in the alluvial aquifer and about 
low actual velocities in the sedimentary aquifer and in the Pillow Lavas. Actually, 
determination of residence times in the sediments and in the Pillow Lavas was the only 
way to support these concepts, and, at this stage, the conclusions were valuable. However, 
to go further, one has to use other tracers for determination of higher residence times (\(^{14}\)C, 
for example) and sample perhaps 5 more locations in the sedimentary and 5 – in the Pillow 
Lavas.

Sampling rainfall for \(^3\)H was not only interesting for the Kouris catchment, but also 
in a broader sense – to know transient \(^3\)H distribution in atmosphere over different regions.

**Stable isotopes**
In our opinion, stable isotope analysis provided good knowledge about hydrology of the 
Kouris catchment, and some constraints could only be obtained by means of stable isotopes 
(for example, origin of saline waters in the sedimentary aquifer). Large amount of 
sampling is usually desired (and it was fortunately fulfilled for the present PhD project), 
since in this case one has the possibility to make statistics and to estimate uncertainties of 
conclusions. We think, that the more groundwater analysis for stable isotopes are done – 
the better. On the other hand, since knowledge about isotope content of precipitation 
doesn’t contribute at all in understanding of hydrology (Chapter 3), quantity of 
precipitation samples could be considerably smaller. It, however, depends on research 
priorities, because studies of stable isotopes in precipitation is itself an interesting scientific 
subject (IAEA – Research Project on isotopes in precipitation).
6.6. GENERAL ACHIEVEMENTS AND LIMITATIONS
OF THE CURRENT STUDY

The PhD thesis presents the first attempt of formulated concepts and quantifications of the regional hydrogeology of the Kouris catchment. To our knowledge, no hydrogeological concepts concerning occurrence of ground water in the ophiolites existed before. Thus, the present studies are interesting as one of the first attempts to understand the hydrogeology of ophiolites in general and of the Kouris catchment in particular, however most of the conclusions must be further validated.

Regional transmissivities, in our opinion, are quite reliable results of the project, since they were more or less independently derived from the field studies and from the numerical model calibration. However, we were not able to model a real 3-Dimensional media since we had no information concerning the distribution of hydrogeological properties with depth. Additionally we took a concept of a confined aquifer, which was not true for the whole region. Finally, one can say, that in the framework of assumptions of a 2-Dimensional confined aquifer, we think, that the estimated transmissivities are reliable, but these assumptions must be further validated.

Regional porosities (Table 6.2) seem to be correct and vary in a quite narrow range as a result of the MODPATH model calibration (Chapter 4). However, the fact, that we model only the zone of active water exchange, without taking into account deep groundwater circulation, brings large uncertainty in the simulation results. The MODPATH model simulates both tritium transport in the unsaturated zone and in the aquifer, which means that it requires porosities and vertical hydraulic conductivities of the unsaturated zone. Although these parameters were also, in our opinion, satisfactory corrected by calibration procedure (Table 2), unfortunately the final model included too many parameters that were not possible to measure in the field.

The concepts about water origin in different types of rock are supported by isotope studies, additionally they can be implemented in the transient numerical model without any contradictions. In our opinion, the conclusions of the Chapters 3 and 4, are reliable. However these concepts are far from quantitative hydrogeological description, where one must not tell “mostly water comes from the ophiolitic complex”, but rather give absolute values with the range of uncertainty. Unfortunately, this task is rather difficult and can be solved only by more sophisticated and refined modelling with appropriate input and calibration data. This must be the next stage of modelling of the Kouris catchment.
Estimation of the steady state recharge by chloride mass-balance method (Chapter 2) is correct only in the ophiolitic part of the catchment. For the sedimentary complex the results are wrong since stable isotope data show that chloride contents in that type of rocks are caused not only by evaporation, but also by rock dissolution.

Calculation of the actual evapotranspiration from the alluvial aquifer (Chapter 5) gives useful values, however this is only one type of actual evapotranspiration occurring in dry seasons in the area. To get more precise ideas about regional evapotranspiration one has to examine transpiration by the pine forest and by riparian vegetation around springs in the upper part of the catchment.

Estimates of steady state water balance components seem to be rather reasonable, as a result of sensitivity analysis, but also because all input data and implemented concepts were compatible. However all model constrains are based on absolute values of baseflows, consequently a water balance is as reliable as estimated baseflow. In the present project we applied only one type of methods of calculation of baseflow (sliding-interval and fixed-interval methods by Sloto and Crouse (1996)), which doesn’t suggest any uncertainty estimation. Thus, we think, that baseflow estimation only by one method, validity of which can’t be checked, is not good at all, since all model conclusions are strongly dependent on this parameter.

Simulation of transient water balance (Figure 6.2) is a pure modelling exercise, which needs to be further validated by baseflow measurements (see §6.7 for further studies).

One simplification of steady state and transient modelling is an assumption of confined regime everywhere in the catchment while it is most probably unconfined in some areas, obviously in the river valleys. In a way, the unconfined behaviour of the aquifer was taken into account by implying high input storativity (§ 6.4.). However, we did not have any field data to derive aquifer thicknesses and hydraulic conductivities in order to model a true unconfined aquifer. In our opinion, this simplifications doesn’t bring considerable errors in final result at the regional scale due to a rough discretization of the model and also because the aquifer is still confined in most of the region according to field data.
Future studies of the hydrogeology of the Kouris catchment should aim to further quantify transient water balance components, to estimate uncertainties of results and to reduce these uncertainties.

The modelling should be done under transient conditions in a coupled regime of surface- and groundwater. The “key” issue of the regional water balance is the separation of transient streamflow in several flow components (surface water, soil water, ophiolitic aquifer, sedimentary aquifer, etc.). Even separating streamflow simply in groundwater and surface-soil water components can considerably improve the reliability of the results of the present study. Also it will be very important to determine more precisely the spatial distribution of baseflow/streamflow. One of the conclusions from the PhD study is that spatially distributed transient data of baseflow/streamflow is the best information to calibrate a regional numerical model of the Kouris catchment. Calibration based only on piezometric heads, on the contrary, may lead to non-uniqueness (Chapter 2) or on even to wrong estimates of storativity (Chapter 4).

Pumping tests in the ophiolitic complex would be helpful more for local water management rather than for the description of regional hydrogeology; however the new pumping tests should, if possible, include one observation borehole at least to estimate storativity (see §6.8). Several new pumping tests are required for the description of the hydrogeology of the sedimentary complex and the Pillow Lavas.

The results obtained from water isotopes were quite useful (see §6.5), however we don’t see the needs for further water sampling for stable isotopes and \(^3\)H. It would be interesting to sample groundwater of the Pillow Lavas and the sediments for other isotopes (\(^{14}\)C, \(^{36}\)Cl, etc.), which can be applied for estimation of higher residence times.

The monitoring of springs discharges, water levels in boreholes and streamflows conducted by the Water Development Department of Cyprus stays very important for further understanding of the regional hydrogeology. It certainly has to be continued. If possible, gauging stations for streamflow monitoring have to be installed also in two or three additional locations of the catchment upstream the existing stations, the best – in the area of the contact between the ophiolites and the consolidated sediments.
6.8. RECOMMENDATIONS FOR DEVELOPMENT
SUSTAINABLE WATER MANAGEMENT

There are several problems in the Kouris catchment, which can be solved by development of sustainable water management. Gradual reduction of groundwater storage resulted in drying up several springs and dying riparian trees in the upper part of the Limnatis subcatchment. However, presently the influence of groundwater pumping on the surrounding environment is not estimated. In our opinion, permission to drill a borehole and to extract certain amount of groundwater, in such a sensitive environment must be based on results of pumping test with calculated storage coefficient and on simplified numerical model estimating influence of the borehole on the water levels and springs in the area of the borehole influence.

Farmers extracting groundwater from the sedimentary complex of the Kouris catchment often have problems with its shortage and high mineralization. Some boreholes can work only for half an hour of a day, while others pump groundwater more steadily. Despite severe water problems appearing in that area, the groundwater resources have never been studied systematically. Our first attempt to investigate the sedimentary rocks of the Kouris catchment has to be further developed.

The absence of integrated water management can worsen the situation even in the areas, which now do not have problems with water. The water balance of the Kouris Dam, that is drinking water reservoir for the whole country, is highly dependent on the regional ground/surface water balance of the area. Ninety five percents of water discharging at the Kouris Dam in dry season, is coming from the ophiolitic aquifer as baseflow. At present, there is always water available in the Dam lake, though its quantity for the last 15 years has had a negative trend. In our opinion, prediction of water levels at the Dam under conditions of increasing groundwater extraction in the ophiolitic complex and a global climate change is an essential water management problem, which can be satisfactory solved only by further quantifying the regional water balance of the Kouris catchment (see §6.7) and developing coupled surface/groundwater models of the area.
REFERENCES


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