3. Hydrogeological conditions at the southwestern slopes of the Mt Olympos–Ossa range in NE Thessaly

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3.1. Introduction

This study aims to determine the recent hydrogeological conditions at the southwestern slopes of Mt Olympos–Ossa range in northeastern Thessaly, where the study area (Sykourio and Elateia basins) are situated (Figure 3.1). The hydrogeological setup of the Sykourio and Elateia alluvial basins and the hydraulic connection with the Eastern Thessalian Plain, which is one of the largest Quaternary alluvial basins of the country, is examined for the late 20th c. AD. The geology of the region is determined in conjunction with the available borehole lithological sections. This way, the geometry of the alluvial system is delineated and its lithological characteristics defined. The main meteorological parameters (rainfall, temperature) are analysed and their patterns are presented.

The available piezometric data are processed and analysed, with conclusions being drawn regarding the hydraulic interaction between the two aquifer units as well as between the studied system and the surrounding formations.

Based on the conclusions drawn from aforementioned study, the main groundwater flow mechanisms of the system are identified.

3.1.1. The study area

Situated at the southwestern slopes of Mt Olympos–Ossa range in northeastern Thessaly is the study area (Sykourio and Elateia alluvial aquifers), presenting marginal alluvial basins in the northeastern part of the Eastern Thessalian Plain in the region of Thessaly.[1]

The plains of Thessaly are situated in Central Greece and are the largest Quaternary alluvial basins of the country. They are separated by the mid-Thessalian hills into the western, or Karditsa plain and into the eastern, or Larissa plain, as illustrated in Figure 3.1. The plains are bounded in the north by the mountains Olympos and Antichasia, in the west by the Pindos mountain range, in the south by Mt Orthys and in the east by the mountains Pilion, Mavrovouni and Ossa.

3.1.2. Hydrogeological units

The geological formations of the wider area present different hydrogeological behaviour, according to their composition and the tectonic impact.

Permeable formations

The main aquifer is developed in the flat parts of the basins, inside the Quaternary deposits. The limestone outcrops of the Chassambali karst aquifer cover a small extent of the area, however their hydrogeological importance is high.

Impermeable formations

The hydrogeological importance of these formations (gneisses, schists, amphibolites, marbles, and serpentinites: compare contribution by V. Melfos in this volume) is limited to a small contribution of recharge to the alluvial aquifer system, via the limited number of joints and the dissolution channels created in the calcite intercalations that exist in the bedrock.

Infiltration and recharge occur mainly through these joints and bedding planes. Low potential aquifers have developed inside the impermeable formations due to the intensive tectonic activity and the creation of secondary porosity. Water from small springs emerging along tectonic faults from the joints of these formations is known to have been used in the past to cover limited water needs.

The alluvial and karst aquifers developed at the southwestern slopes of the Mt Olympos–Ossa range in northeastern and eastern Thessaly include the following (Figure 3.2): the Sykourio alluvial basin with the sub-basins of Kalochori and Pournari, the Chassambali karst system, and the

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[1] Compare also Reingruber et al. in print; Toufexis and Reingruber 2020.

Publiziert in: Reingruber, Agathe; Toufexis, Giorgos (Hg.): Siedlungsdynamik in den Becken von Sykourio und Elateia, Nordost-Thessalien, in prähistorischer Zeit. Ergebnisse der archäologischen, geomorphologischen und geophysikalischen Untersuchungen, Heidelberg, Propylaeum, 2025 (Prähistorische Archäologie in Südosteuropa, Bd. 34). DOI https://doi.org/10.11588/propylaeum.1583

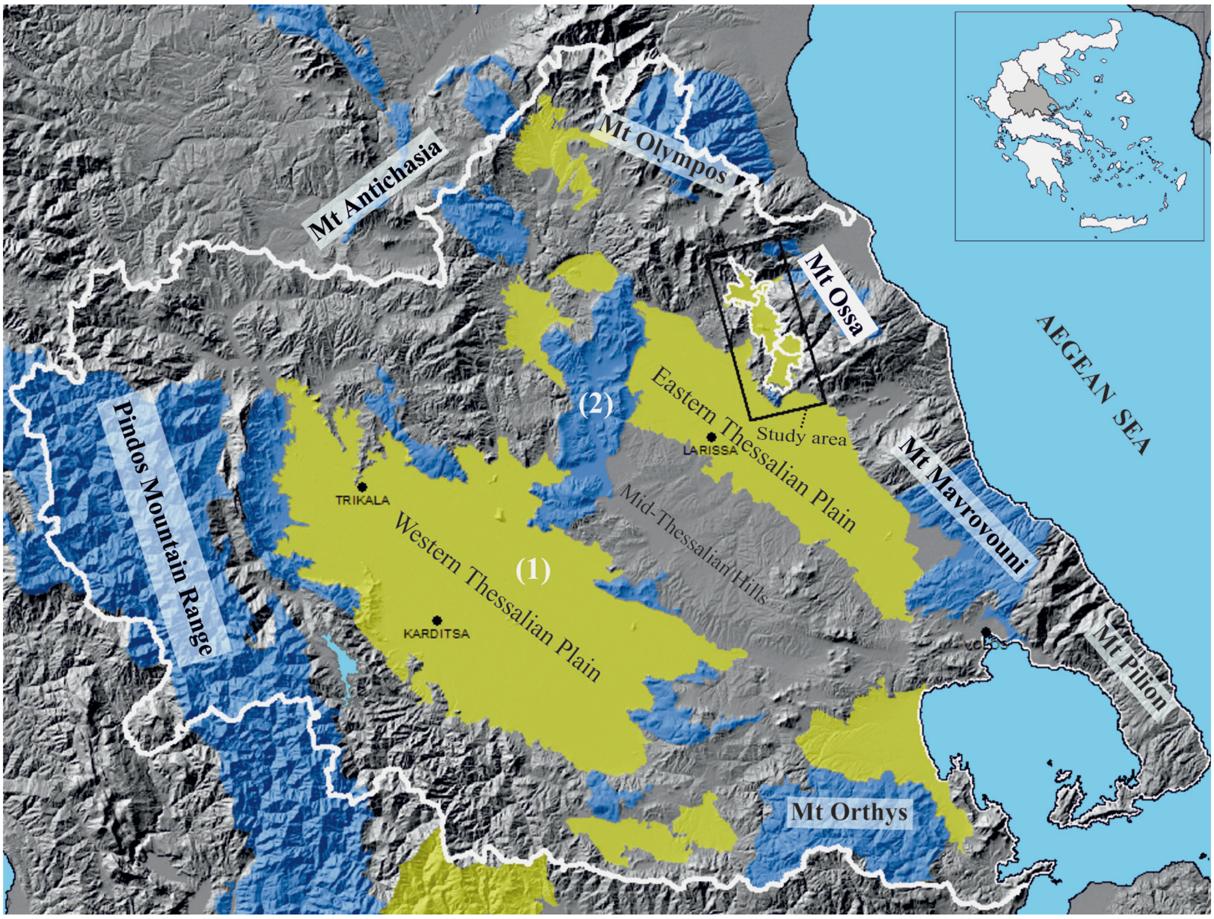

Fig. 3.1. Alluvial (1) and karst (2) aquifers in Thessaly and location of study area.

catchment of Pinios River with the alluvial basins of Elateia and Gonnoi.

The Sykourio and Elateia alluvial basins are intramountain basins found at the southwestern slopes of the Mt Olympos–Ossa range in northeastern Thessaly. In the basins exists an extensive aquifer system, the resources of which cover most of the irrigation demands for the agriculture practiced in the region. The available geological, hydrological and hydrogeological data are limited and non-standardized. However, this program makes an effort to identify and understand the main hydrological and hydrogeological characteristics of the aquifers in the area.

3.2. The Sykourio alluvial basin

The Sykourio alluvial basin develops at the southwestern slopes of Mt Ossa into a NE–SW direction with an area of approximately 114.8 km^2. The surface runoff allows us to distinguish two independent hydrological sub-basins: that of Kalochori and that of Pournari (Figure. 3.2).

The slope, landform and the configuration of the watershed have created a drainage network, which takes the surface flow towards a swampy area. The drainage pattern of the area is of sparse dendritic type. Ephemeral streams (creeks) flow out of Mt Ossa usually only in wet months of the year and discharge mainly through the Xerias and Kalamitsa streams onto the plain of Sykourio.

Irrigation demand in the whole Sykourio alluvial basin is covered by private and municipal boreholes and, to a minor extent, by water abstracted from the Xerias and Kalamitsa streams. Domestic water is supplied by groundwater. Systematic exploitation of the groundwater resources of the alluvial aquifer system was initiated in the early 1970s, as part of the Thessalian plains groundwater resources development project.[2]

[2] SOGREAH 1974.

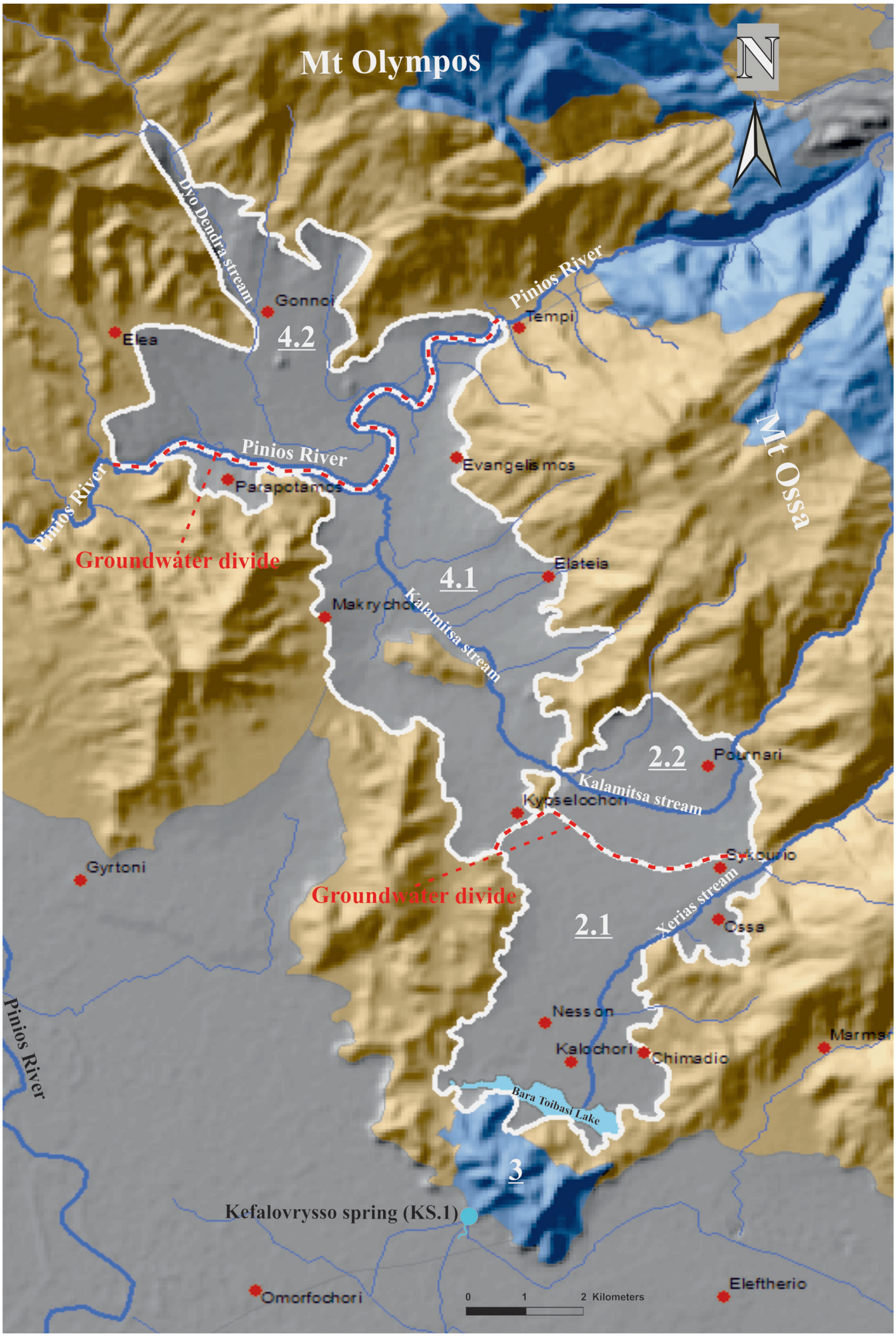

Fig. 3.2. Alluvial and karst aquifers at the southwestern slopes of the Mt Olympos–Ossa range in north-eastern Thessaly (2.1 Kalochori sub-basin, 2.2 Pournari sub-basin, 3 Chassambali karst aquifer, 4.1 Elateia basin, 4.2 Gonnoi basin).

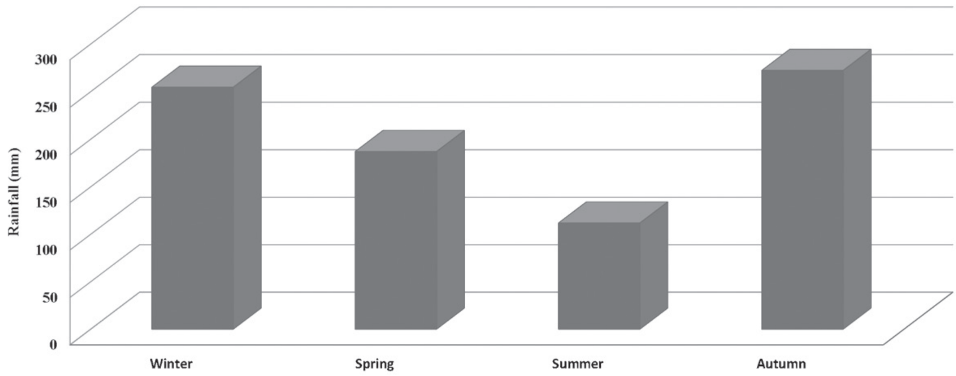

Fig. 3.3. Mean seasonal rainfall for the period 1974–1999 at Spilia station.

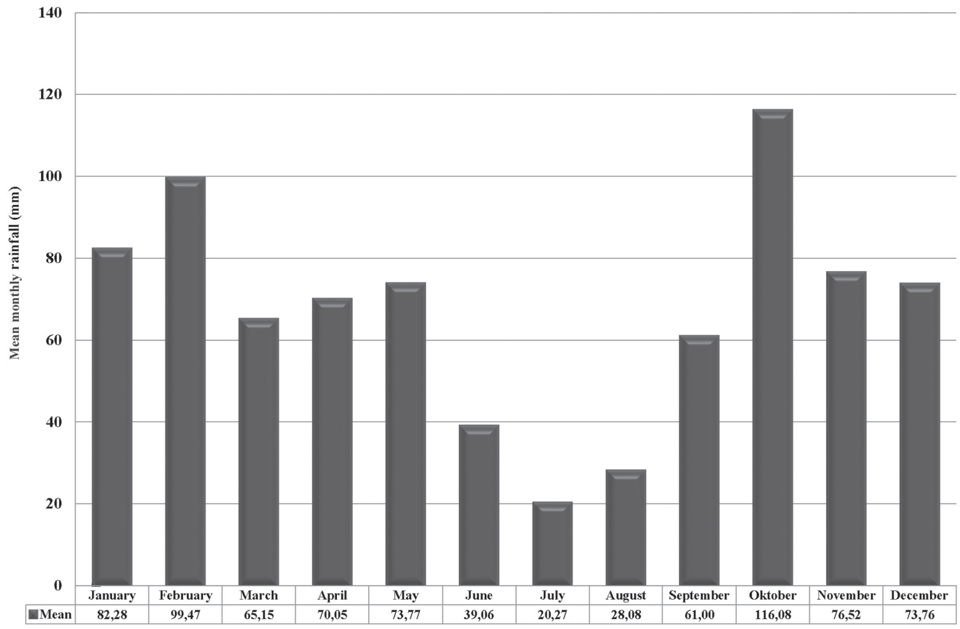

Fig. 3.4. Mean monthly rainfall pattern at Spilia station for the period 1974–1999.

Rainfall

The wide area of Sykourio is characterized by cool winters and hot summers. Rainfall is distributed mainly during the autumn and winter months, with some storm events between June and September.

The watershed of Sykourio is covered by one rain gauge only, located in Spilia at an altitude of 813.00 m above sea level (masl). The rainfall data available for the 26 years between 1974 and 1999 are used to calculate the mean annual precipitation for this period. It was calculated to be 810.46 mm. Figures 3.3 and 3.4 present the mean seasonal and mean monthly rainfall for the period 1974–1999 at the Spilia station.

Temperature

Temperature data in the whole of Thessaly was available from only nine stations, and it was provided in the form of mean monthly values. These values are understood to be derived from observations made three times each day. The available temperature records of the stations cover a time series of 18 years.

The mean annual temperatures for the period 1974–1999 were plotted against their respective altitude, and a linear regression trend line was drawn (Figure 3.5). The derived equation is

$$Y=-0.0048X+16.64$$

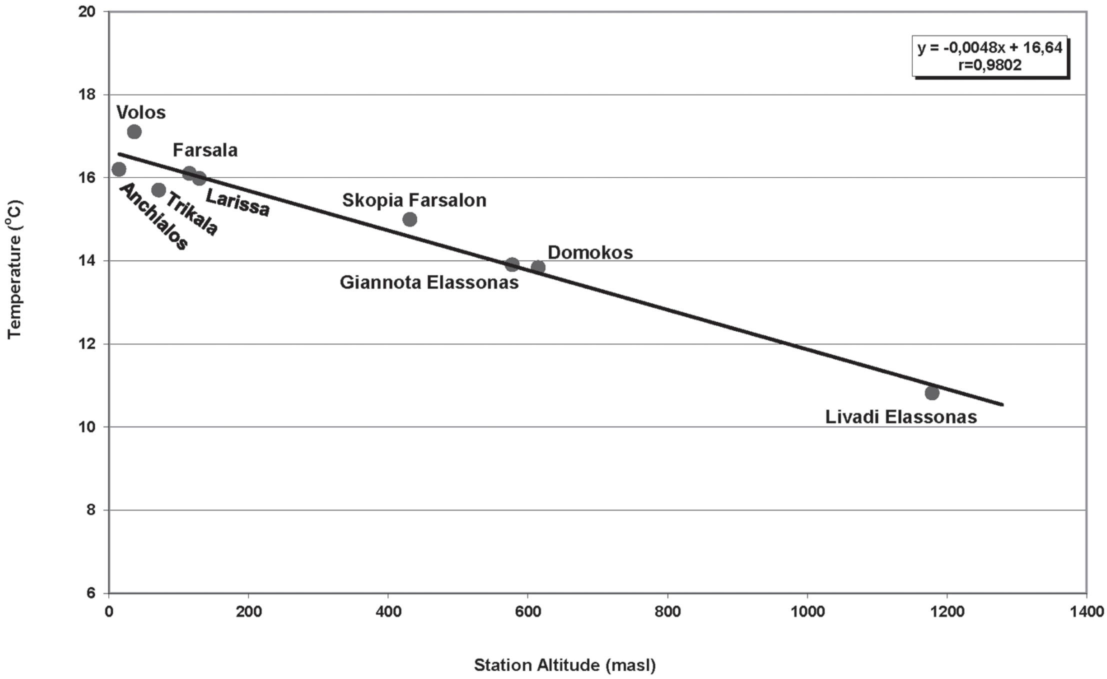

Fig. 3.5. Linear regression trend and derived equation between temperature and altitude of the stations for the period 1974–1999 in Thessaly (Manakos 2010).

r=0.9802 and is valid for altitudes within the range of 15–1179 masl.[3] The derived expression provides an accurate estimate of temperature at altitudes within the above-mentioned range. Using the above derived equation in the wider area of Sykourio the mean annual temperature was calculated to be 13.63°C.

3.2.1. The Kalochori alluvial sub-basin

The alluvial sub-basin of Kalochori forms the central and southwestern part of the Sykourio plain and contains important quantities of groundwater. The watershed of Kalochori hydrological sub-basin has an elongated NE–SW direction and occupies an area of 81.7 km^2. Steep to very steep slopes, of the Ossa mountainous area, which cover 59.5% of the total area of the watershed, characterize the northeastern part of the catchment.

Flat to gentle slopes, which cover 27.5% of the total area of the watershed and have a slope ranging 1.3–2.0%, form the Kalochori alluvial sub-basin and characterize the remaining parts of the watershed.

This physiographic unit includes a swampy area (Bara Toibasi Lake) that lies in the southern part of the watershed. The slope of the watershed rises slowly in S–N direction away from the swampy area. Ephemeral streams flow out of Mt Ossa usually only in wet months of the year and discharge onto the plain of Sykourio. The Xerias (Megalo Rema) is the major stream. It has a torrential intermittent character and flows at steep slopes, before entering the Kalochori alluvial sub-basin.

Large water amounts of the Xerias discharge into the Kalochori alluvial sub-basin for a short distance before dissipating by evaporation and infiltration into permeable sediments of the Kalochori alluvial sub-basin. There is no data on runoff drainage, and only a minor amount of the Xerias surface water reaches and recharges Bara Toibasi Lake. Because of the scarcity of surface water resources, there is consequently a heavy reliance upon groundwater for stock, domestic, irrigation and villages water supplies.

The margins of the Kalochori alluvial sub-basin contain a thick sequence of Quaternary alluvial deposits, and are formed in the northern, eastern and western direction by Paleozoic gneisses, schists and amphibolites (compare Chapter 5). The southern boundary is defined by the tectonically

[3] Manakos 2010.

uplifted ophiolites (serpentinites) and Jurassic-Cretaceous limestone outcrops of the Chassambali. This uplifted block is 397.0 m high and 3.5 km wide. It has created a morphological-hydrological barrier in surface water of the Xerias stream to a southward flow. Due to the lack of surface runoff in a final surface river recipient, the surface waters formed and filled the Bara Toibasi Lake.

3.2.1.1. Structure

Agriculture based upon irrigation from groundwater forms one of the main elements of the Greek economy. Most of the agricultural activity is concentrated in the Quaternary alluvial basins in Greece. Due to the socio-economic development of the country in the beginning, coupled with changes in climatic conditions since the late 1980s, the exploitation of the groundwater resources of these basins have been significantly increased resulting in excessive piezometric surface decline. Consequently, concern has been raised nationally regarding the future sustainability of much of the groundwater.

As noted above, the Sykourio and Elateia alluvial basins are intra-mountain basins found at the southwest slopes of the Mt Olympos–Ossa range in northeastern Thessaly. The Sykourio basin contains alluvial deposits of mixed fluvial, lacustrine and terrestrial origin.

In the lowland part of the Kalochori hydrologic sub-basin, and in particular from Sykourio in the north to the hill of Chassambali in the south, the alluvial aquifer of Kalochori was formed. All Quaternary sediments (aquifers-aquitards) are deposited on the crystalline rocks of the Pelagonian zone. Their total thickness exceeds 170 m in places, and within these sediments formed the main aquifer system of the Kalochori sub-basin.

Lithological data of drilling columns, which were performed to cover irrigation and water needs by the Ministry of Agriculture (Land Reclamation Services) and the communities of the area, were used to delineate and to determine the geometrical elements of the Kalochori aquifer. Depths of identified boreholes constructed in the Kalochori aquifer vary from less than 80 m to more than 140 m. Cross-sections portray the lithological sequence from drill holes along a generally NE–SW profile through the basin (Figures 3.6 and 3.7). Four typical sections of the basin are illustrated, but a fifth borehole (P16L) has not been documented lithologically, since here only the water level measurements were taken (compare Chapter 3.2.1.3).

The borehole SY-1 was drilled near the northern margin of the sub-basin in the year 1992, at an elevation of 143.0 masl with a total depth of 115.0 m below land surface (mbls). It penetrates the filled sediments of the basin and also intersects harder unweathered crystalline bedrock (schists) in 92.0 mbls depth. The static water level was 55.30 m (1992) and at a constant rate of 80 m^3/h.

The borehole 35L was drilled 500 m southwest of the borehole SY-1 in the year 1976, at an elevation of 139.0 masl with a total depth of 82.0 m. It penetrates only the filled sediments, but does not reach the crystalline bedrock. The static water level was 30.73 mbls (20.07.1976). The pumping test was conducted for 72 hours at a constant rate of 138 m^3/h. After a continuous 72 hours pumping, a volume of 9,936 m^3 water was pumped out and a drawdown of 6.51 m was measured.

The borehole 34L was drilled in the central part of the sub-basin in the year 1975, at an elevation of 102.0 masl with a total depth of 121.0 mbls, penetrating the sediments and the crystalline bedrock (schists). The static water level was 28.65 mbls (20.07.1975). The pumping test was conducted for 72 hours at a constant rate of 144 m^3/h. After a continuous 72 hours pumping, a volume of 10,368 m^3 water was pumped out and a drawdown of 5.80 m was measured.

The borehole 47L is located near the southern margin of the basin in the year 1976, at an elevation of 89.0 masl with a total depth of 140.0 mbls, penetrating the sediments and the crystalline bedrock (schists). The static water level was 13.82 mbls (20.08.1976). The pumping test was conducted for 72 hours at a constant rate of 130 m^3/h. After a continuous 72 hours pumping, a volume of 9,360 m^3 water was pumped out and a drawdown of 8.57 m was measured.

The four bore logs (SY-1, 35L, 34L and 47L) show significant lithological variations, and the examination of their sections indicated a high degree of heterogeneity along with abrupt lateral variations.

Examination of the boreholes' lithological sections indicates the presence of a complex multilayered

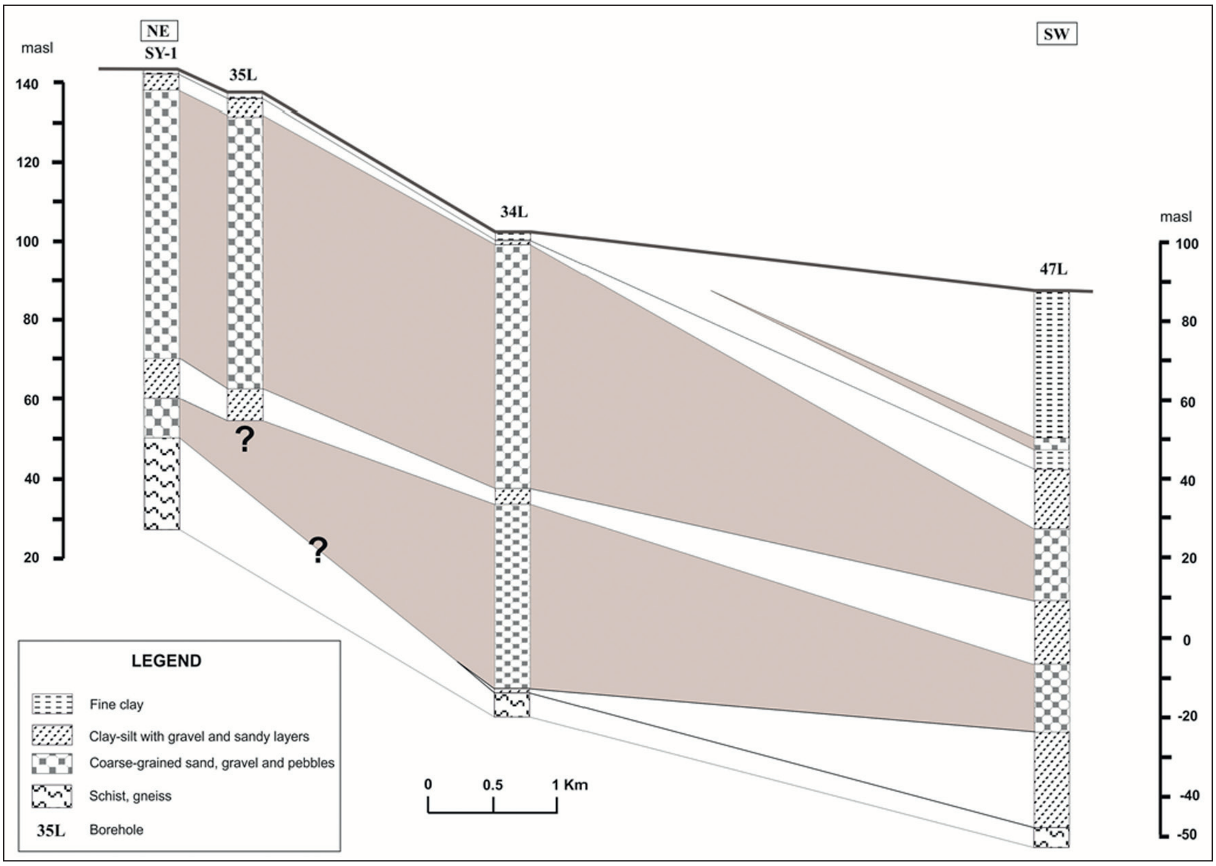

Fig. 3.6. Typical lithological sections of Kalochori sub-basin showing the structure of the aquifers. The exact direction of the section and the location of boreholes are indicated in the map on Figure 3.7.

structure deposited during the Quaternary. Boreholes SY-1 and 35L showed that the alluvium composition is generally a yellow-coloured fine clay in its upper parts (1.0 and 1.5 m thick) and clay-silt with gravel and sandy layers in the lower parts (ca. 4.0 m and 4.5 m thick). A strong lithological contrast between the last sequence and the sediments below is exhibited. Observed below these parts of the clay-silt with gravel and sandy layers was an extensive sequence consisting of alternating bands of coarse-grained sand, gravel and pebbles (67.0 m and 68.0 m thick) and clay-silt with gravel and sandy layers in the lower part (ca. 10.0 m and 8.0 m thick respectively). Below the last section, at the base of the unconsolidated succession and, other than in 35L, in SY-1 appears again the section of coarse-grained sand, gravel and pebbles layers (10 m thick) in a depth between 72.0 and 82.0 mbls (60–70 masl) and finally the schists of the crystalline bedrock in a depth of 92.0 mbls.

In the central part (borehole 34L) both the fine clay (ca. 2.0 m) and the clay-silt with gravel and sandy layers (ca. 1.0 m thick) show similar thickness to the boreholes SY-1 and 35L. Below these parts of fine clay and clay-silt with gravel and sandy layers the extensive sequence of alternating bands of coarse-grained sand, gravel and pebbles decreases in thickness (ca. 61.0 m), compared to boreholes S-1 and 35L. Below, once again, appears the same lithological transition from sequences of clay-silt with gravel and sandy layers (ca. 4.0 m thick) to the alternating bands of coarse-grained sand, gravel and pebbles (ca. 46.0 m thick).The borehole penetrates the entire sediment infill (114.0 m) of the basin before encountering the crystalline basement in a depth of 114.0 mbls.

In the southern part of the sub-basin (borehole 47L) both the fine clay in the sequence shows significant thickness (ca. 42.0 m, interrupted by a 3 m-thin lenticular layer of sand and gravel at ca. 50 masl), as does the clay-silt with gravel and sandy layers (ca. 15.0 m). Below this we observe an extensive sequence of alternating bands of coarse-grained sand, gravel and pebbles that decrease in thickness

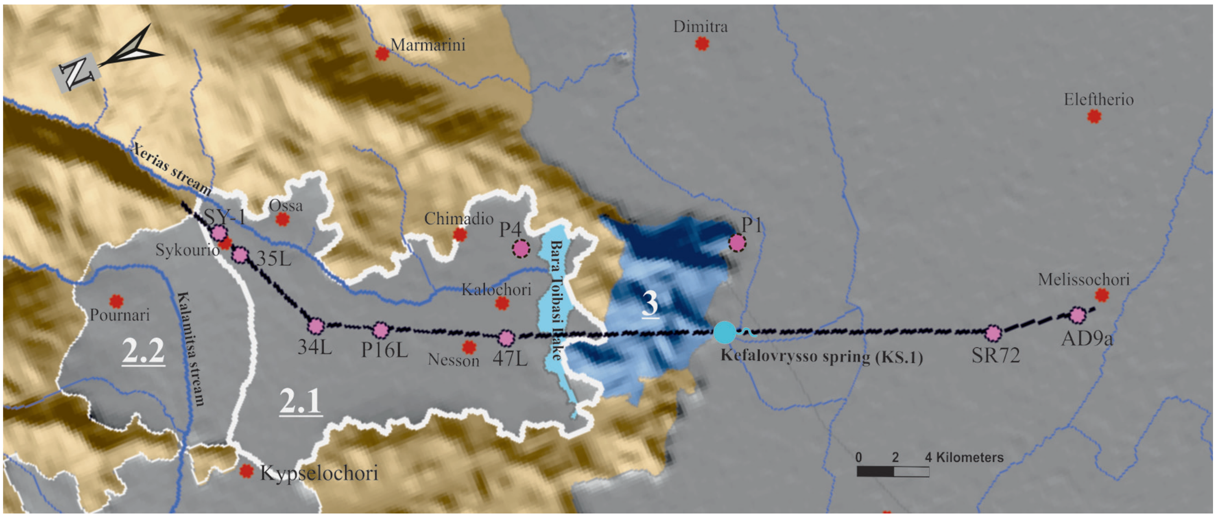

Fig. 3.7. Kalochori (2.1) and Pournari (2.2) alluvial sub-basins, the Chassambali karst aquifer (3) and the adjacent Thessalian Plain with the location of boreholes and the direction of the geological and hydrogeological sections.

(ca. 18.0 m). The same pattern of lithological transition, namely from sequences of clay-silt with gravel and sandy layers (thickness of ca. 16.0 m and 24.0 m respectively) to alternating bands of coarse-grained sand, gravel and pebbles (thickness ca. 18.0 m and 17.0 m respectively) can be identified in the deeper section of the borehole. The borehole penetrates the entire sediment infill of the basin before encountering the crystalline basement in a depth of 135.0 mbls.

Presented in Figure 3.6 is a lithological cross-section through the Kalochori sub-basin. It shows a general progression from coarse, gravelly alluvium in its northeastern part to finer grained sediments (fine clay and clay-silt with gravel and sandy layers) in the southwestern part of the sub-basin. This progressive refinement is represented by an increase in the proportion of clay-silt and shows the effects of progressive mechanical abrasion, decreasing depositional energy, and perhaps a greater contribution of fine-grained detritus weathering from bedrocks present in the northeastern, eastern and western part of the sub-basin.

Based on the compiled geological map and the examination of the lithology of the alluvium, a geological section was constructed across the Kalochori sub-basin (Figure 3.8). Basing on the above factors we delineated the extent and calculated the thicknesses of the exploitable aquifers. Specifically, the area of the aquifer is approximately 20.4 km^2; the aquifer has an average thickness of 72.5 m, while the average total volume of the aquifer is 4,457 hm^3.

3.2.1.2. Hydrogeology

Examination of the lithological sections in the upper parts of the Kalochori sub-basin showed a lithological alternation of aquitard-semi aquitard layers (of clay-silt) with water bearing layers (of coarse-grained sand, gravel and pebbles), each of few meters only. This sequence was identified in the two typical lithological sections of the studied boreholes to a depth of ca. 40.0 mbls. In the deeper parts of the basin these bands of the water-bearing layers increase in thickness (50.0–57.0 m), whereas the thickness of the aquitard-semi aquitard layers remains the same as in the upper part.

Examination of the lithological sections of the upper deposits of the basin showed a fine clay sequence (Figure 3.8). Outcrop exposures of these formations extend from the northeastern basin border near Sykourio southwards through the central part as far as the southern basin border. Their thickness varies from 1.0 m to 2.0 m in the northern (borehole SY-1) and central (borehole 34L) part of the basin and 42.0 m (borehole 47L) in the southern part. This clay layer presents a total aquitard hydrogeological sequence and creates confined aquifer conditions mainly in the southern part of the basin. Due to the minor thickness of this fine clay layer, the largest recharge rate from Xerias stream into the

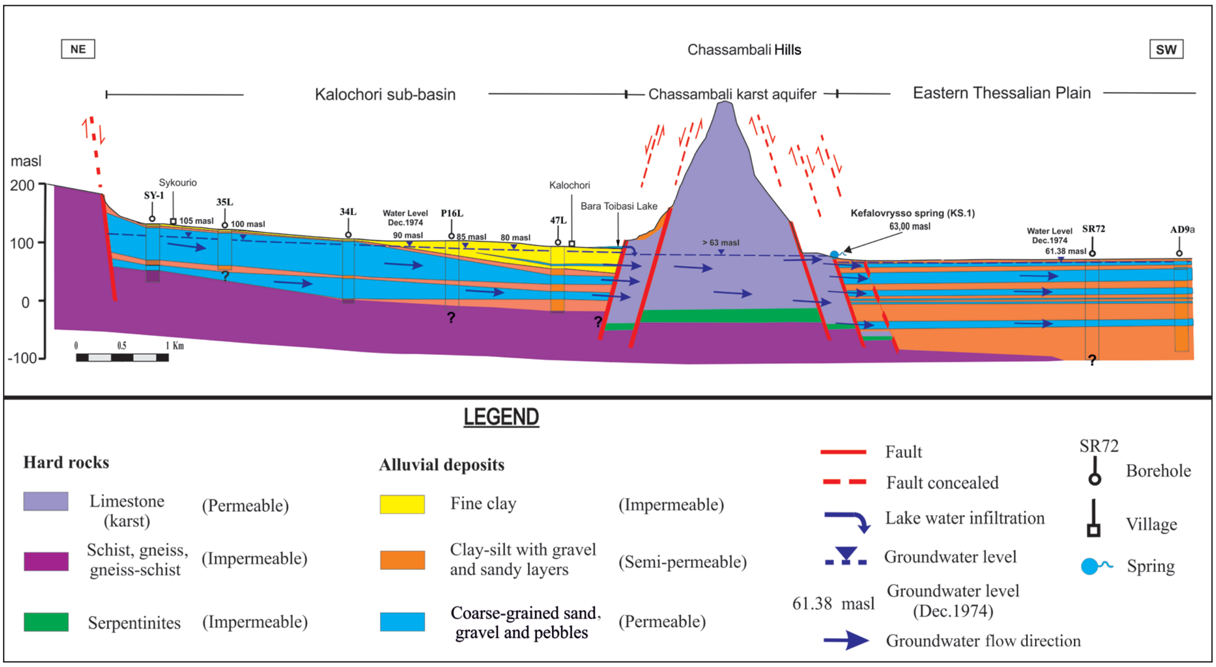

Fig. 3.8. Hydrogeological section across the alluvial aquifers of Kalochori, Eastern Thessalian Plain and Chassambali karst aquifer. The exact direction of the section and the location of boreholes are indicated in the map on Figure 3.7.

permeable upper coarse alluvium of the Kalochori aquifers (coarse-grained sand, gravel and pebbles) occurs in the northern part of the basin. The thickness varies from 67.0 m, to 61.0 m and 18.0 m in the northern (borehole SY-1), central (borehole 34L) and southern (borehole 47L) parts of the sub-basin respectively and is the main aquifer of the basin.

The semi-aquitard intermediate layer of clay-silt with gravel and sand underlies this formation. Its thickness varies from 12.0 m to 4.0 m and 16.0 m in the northern, central and southern part of the sub-basin respectively. It is followed by the water bearing deposits that are dominated by coarse-grained sand, gravel and pebbles. Its thickness varies from 10.0 m to 46.0 m and 17.0 m in the northern, the central and the southern parts of the sub-basin respectively.

The semi-aquitard layer with clay-silt, gravel and sand underlies this formation only in the central and southern part of the sub-basin. Its thickness shows southwards an enormous increase varying from 1.0 m in the central to 24.0 m in the southern part of the sub-basin. In three of the four boreholes the bedrock was reached, which consists of impermeable rocks (schist and gneiss).

Except for the upper fine clay sequence, the different deposits of coarse-grained sand, gravel and pebbles and clay-silt with gravel and sandy layers act as one aquifer with direct hydraulic contact; i.e., there is no hydraulic separation into separate aquifers.

Based on the above factors (data, elements), we could delineate the extent and calculate the thicknesses of the exploitable aquifers. Specifically, the area of the aquifer is ca. 20.4 km^2. The aquifer has an average thickness of 73.0 m, while the average total volume of the aquifer is 4457 hm^3.

3.2.1.3. Groundwater fluctuation

Water-level measurements from observations on boreholes are the principal source of information about hydrological stresses acting on aquifers and how these stresses affect groundwater recharge, storage and discharge. They reveal how changes in water stored in the aquifer can vary from place to place depending upon soil type, irrigation practices, recharge from precipitation, and the areal extent and magnitude of water withdrawals. The depth to groundwater generally decreases from north to south at a rate of about 0.46 m/km. The deepest water levels are near Sykourio in the northeastern area of the basin, and the shallowest in the southern area near Kalochori and Bara Toibasi Lake (Figure 3.8).

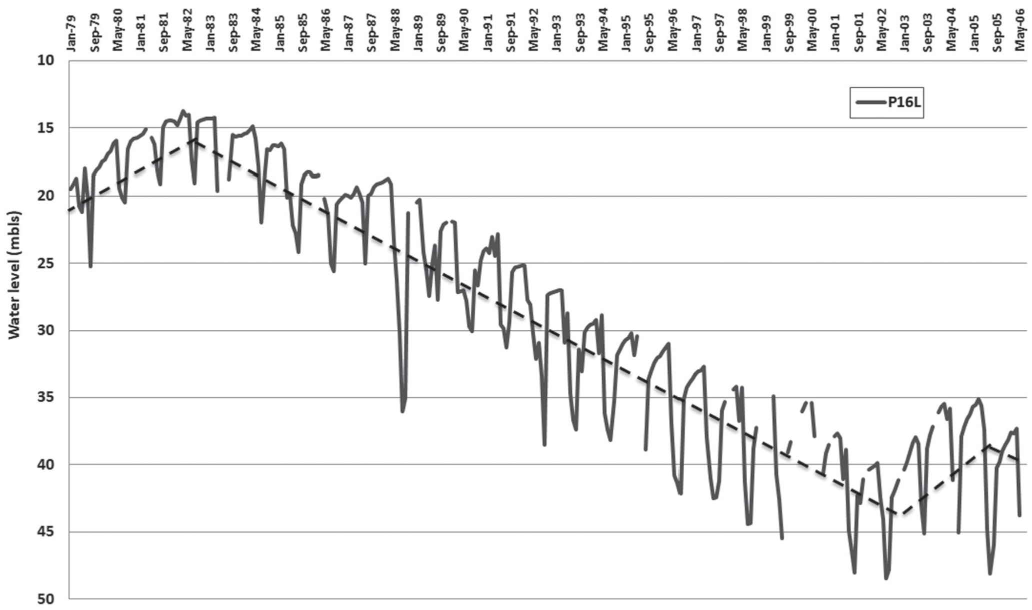

Fig. 3.9. Water-level hydrograph and trend line at borehole P16L in Kalochori sub-basin, based on monthly measurements between 1979 and 2006 (27 years). Trend line is based on simple linear regression between water level and time.

Water-level measurements obtained from the Land Reclamation Services are available for only one monitoring point (borehole P16L), made once in a month, and they cover the period 1979–2006 (27 years). Water-level hydrographs for borehole P16L are presented in Figure 3.9. The water-table elevation is highest in April or May and lowest in July or August. The depth to groundwater in the Kalochori aquifer varies from 12.0 to more than 32.0 mbls.

Perusal of the fluctuation data in borehole P16L of the alluvial aquifer of Kalochori sub-basin shows that water-levels follow a natural cyclic pattern of seasonal fluctuation, typically rising during the winter and spring due to greater precipitation and recharge, then declining during the summer and fall owing to less recharge and greater evapotranspiration. The magnitude of fluctuations in water levels can vary greatly from season to season and from year to year in response to varying climatic conditions.

Superimposed on natural, climate-related fluctuations in groundwater levels are the effects of human activities that alter the natural rates of groundwater recharge or discharge. The withdrawal of groundwater by pumping is the most significant human activity that alters the amount of groundwater in storage and the rate of discharge from an aquifer.

Perusal of the fluctuation data in borehole P16L shows an average seasonal fluctuation of about 8.76 m and a maximum fluctuation of 19.05 m.

3.2.1.4. Groundwater level trend

Groundwater levels indicate that significant hydrologic changes have occurred within the aquifer of the Kalochori sub-basin. Over the years the intense use of groundwater for irrigation in the basin has caused major water-level declines. The depth to the water-table ranges nowadays from less than 14.0 mbls in the Kalochori area near the southern part of the basin, to as much as 30 mbls in the northern part of the Kalochori sub-basin.

At borehole P16L, from 1979 to 2006 the water level declined by 18.67 m (0.69 m/yr.). However, the water level had been increasing at a rate of 5.24 m (1.31 m/yr.) between 1979 and 1983. The decline rate was 23.59 m (1.24 m/yr.) between 1983 and 2002, whilst the water level increased to 2.20 m (0.73 m/yr.) between 2002 and 2005. Finally, the decline rate was 2.52 m (2.52 m/yr.) between 2005 and 2006.

The intensification of exploitation of the groundwater resources was initiated in 1983, reflecting the decreased water reserves stored in the aquifer. As shown in Figure 3.9, the water level measurement dates indicate that most of this regional decline, occurring after 1983, is related to excessive groundwater extraction and most likely the result of less precipitation and groundwater recharge in the post-1983 period. Extractions for mostly almond-irrigation cause seasonal draw-downs of up to 24.0 m (borehole P16L) in the unconfined portion of the Kalochori aquifer. This also indicates that pressure levels have reached a new equilibrium at current extraction levels (apart from the impact of the drought in 1988).

Salient details of trend analyses on piezometric heads show a steady decline in the groundwater level over the mentioned period, indicating that the groundwater extraction exceeds the recharge. The imbalance between recharge and extraction underlines the necessity to limit the use of this aquifer system in order to avoid adverse impacts. In case of over-exploitation, water levels will not rise again because no water is brought back into the aquifer storage.

3.2.1.5. Piezometry

The alluvial aquifer of the Sykourio basin can be characterized as an aquifer stress area where water demand is high and where relatively large numbers of high capacity boreholes extract groundwater in close proximity.

The morphology of the piezometric surface of the aquifers allows us to study the flow of groundwater and the recharge of aquifers, while its seasonal fluctuations determine the variation in ground-water reserves and how they are replenished.

Groundwater level measurements were made in August 1971, on a network of 60 boreholes (public, municipal, and private) distributed throughout the basin in order to determine the piezometric surface of the Sykourio aquifers. The compilation of the piezometric map was based on an up to now unique map created by SOGREAH 1974. The contour lines of the piezometric map relates to the general picture of the cumulative piezometry of all the individual pressurized aquifers.

From the course of the piezometric curves, the main feature is the identification of two hydrogeological sub-basins (Kalochori and Pournari), whose underground watershed, as is often the case, corresponds with surface watershed.

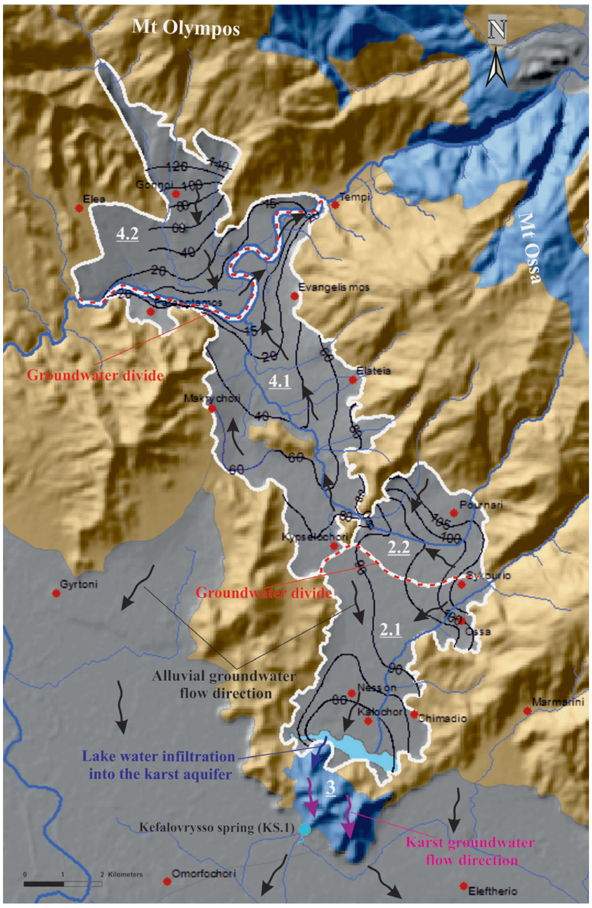

Fig. 3.10. Piezometric maps and groundwater discharge directions of the Kalochori (2.1), Pournari (2.2) alluvial sub-basins and the Elateia (4.1), Gonnoi (4.2) alluvial basins, in August 1971, based on the map compiled by SOGREAH (1974).

Piezometry in the Kalochori sub-basin

The following conclusions are drawn from the study of the piezometric map of the Kalochori sub-basin (Figure 3.10). The main features of the piezo-metric surface morphology are the waveforms of the piezometric contours, with convergent and divergent flow lines. The spacing of piezometric contours and the spatial variability in the aquifer's drainage direction determine the slope of its piezometric sur-face and therefore of its hydraulic gradient.

Groundwater level curves of the piezometric map clearly indicate an aquifer discharge axis from NE to S. This axis follows the course of the Xerias stream, which is the main stream of the hydro-graphic network of the Kalochori sub-basin and is drained towards the south, to the Bara Toibasi Lake. The stream flow seems to indicate that the Xerias consistently loses water to alluvial aquifers along its flow path from where it first flows onto the alluvial fan deposits (Xerias fan). Thus, at the NE part of the basin, near Sykourio, the main recharge is carried out by the surface runoff of the Xerias and shows the largest hydraulic gradient values, ranging between 1.24% and 1.06%.

The hydraulic gradient has an average value of 0.27% in the central part of the Kalochori sub-basin and grows significantly, ranging between 0.53% and 0.60% as it approaches the south border of the Kalochori alluvial aquifer. This is due to the fact that the alluvial aquifer interacts with the surrounding karst aquifer of Chassambali, where it discharges by lateral leakage. The karstic features, due to the well-developed karstification, form favourable hydraulic conditions for handling the large lateral hydraulic loads of the Kalochori alluvial aquifer towards the unconfined Chassambali karst aquifer.

Groundwater movement of alluvial aquifer, as indicated by the potentiometric surface contours (Figure 3.10), is mainly to the south with inflow from the northeast. In general, hydraulic gradients are very low as a result of the large thickness of aquifer and, hence, high transmissivity. It should be noted that no aquifer parameters are available for this sub-basin. In general, the surface topography of the Kalochori alluvial sub-basin is reflected by the depth to the potentiometric surface.

3.2.2. The Pournari alluvial sub-basin

The hydrological watershed of the Pournari sub-basin also has a NE–SW elongated form and occupies an area of 40.6 km^2. Its maximum altitude is 1230 masl (peak Psila Dendra of Mt Ossa, north of the locality Spilia); the altitude at its mouth is ca. 85 masl. The Pournari sub-basin is mainly drained by the Kalamitsa stream and the secondary Livadorema stream, which converges with the Kalamitsa stream at the height of Bounarbasi. The surface water of Kalamitsa stream through the Bounarbasi gorge enters into the Elateia alluvial sub-basin, and the river Pinios is its ultimate recipient. Recognisable in the watershed of Pournari sub-basin are similar conditions regarding the macro-morphological character of the hydrological watershed as in the watershed of Kalochori sub-basin. Steep to very steep slopes of the Ossa mountainous area, which cover 70.0% of the total watershed area, characterize the northeastern part of the watershed.

With an areal extent of about of 9.87 km^2, the Pournari alluvial sub-basin forms the remaining part of the watershed. Flat to gentle slopes, which cover 30.0% of the total area of the watershed, have a slope ranging between 0.2–4.0%.

3.2.2.1. Piezometry in the Pournari sub-basin

The following conclusions are drawn from the study of the piezometric map of the Pournari sub-basin (Figure 3.10).

The main features of the piezometric surface morphology of the Pournari alluvial sub-basin are similar to those of Kalochori. Waveforms of the piezometric contours, with convergent and divergent flow lines can be recognized. The spacing of piezometric contours and the spatial variability in the aquifer's drainage direction determine the slope of its piezometric surface and therefore of its hydraulic gradient.

The groundwater recharge is through surface water infiltration, direct percolation of rainfall, groundwater inflow along the northern, northwestern and partly from southern boundaries and the infiltration of irrigation water.

Groundwater isodynamic curves of the piezometric map clearly indicate an aquifer discharge axis from east to west. This axis follows the course of the Kalamitsa stream, which is the main stream of the hydrographic network of the Pournari alluvial sub-basin and is drained towards the west to the Bounarbasi gorge.

In the eastern part of the basin, north of Sykourio, the main recharge is carried out by the surface water runoff of Kalamitsa stream and shows the largest hydraulic gradient values, ranging 0.95%.

In the central region of the Pournari alluvial sub-basin the hydraulic gradient has an average value ranging between 0.40% and 0.46%.

In the southwestern part of the sub-basin, near Bounarbasi gorge, the alluvial aquifers are bounded by the impermeable crystalline schists, which form the bedrock of the basin and also the western limit of the aquifers. The hydraulic gradient grows significantly, ranging between 1.32%–1.92%, with an average value of 1.41%. The higher hydraulic load is due to the wedging of the aquifer against rising bedrock in the west-northwest and to the re-

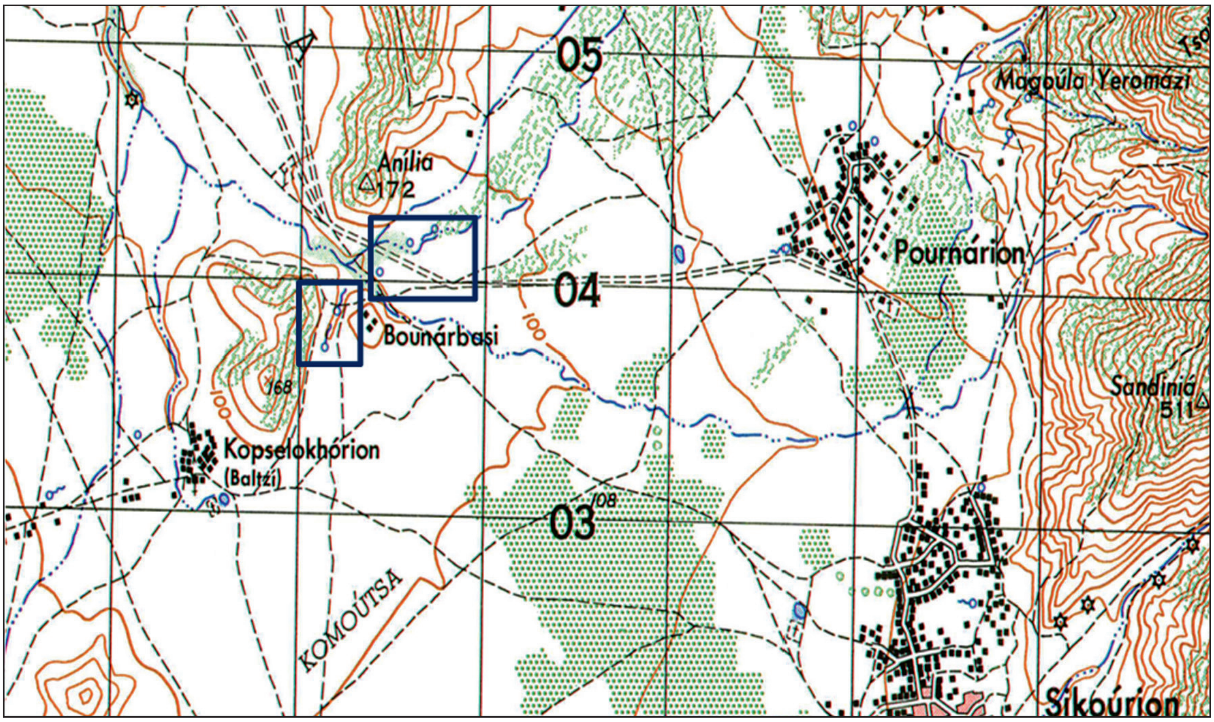

Fig. 3.11. Spring outlets in the Bounarbasi gorge (left square) and in the stream beds near the confluence of the Kalamitsa and Livadorema (right square). Source: Topographical Map of the USA, Army Map Service 1945.

duction in thickness of the water bearing formations in the Pournari alluvial sub-basin.

The seasonal runoff values differ largely for winter and autumn. According to them, the runoff values are generally high in the rainy period for the wider catchment, indicating a high susceptibility to erosion and sedimentation. The runoff coefficient is higher for April, although February and March receive the most precipitation. The end of the rainy season explains this higher value. The soil moisture storage is saturated, water tables have risen, many intermittent springs and streams appear, and no further infiltration can occur. These result in more runoff than during April, compared with the receiving precipitation.

These hydraulic conditions are due to spring outlets, which occur at different altitudes near the Bounarbasi gorge and which have dried up. Nonetheless, their presence is attested by older topographic maps and oral testimonies of local residents as well as by the appearance of springs in the Kalamitsa and Livadorema stream beds (Figure 3.11). The general piezometric image in both sub-basins remains the same due to the similar average hydraulic gradient: that of Kalochori is 0.82% and that of the Pournari is 0.97%.

A large hydraulic gradient is observed at the margins of the two sub-basins, and in particular from the piezometric contours 105.0 masl to 95.0 masl. For the Kalochori sub-basin the hydraulic gradient was calculated at 1.19%, and for the Pournari sub-basin at 0.81%. Hydraulic gradients decrease from 95 masl to 90 masl, calculated at 0.27% and 0.46% respectively.

The groundwater contours in the aquifer of the Pournari sub-basin range between 105.0 masl and 80.0 masl. Because groundwater flows perpendicular to the piezometric potential lines, the direction of the groundwater flow is from east, northeast and southeast towards the west along the Kalamitsa stream. Interaction between groundwater and Kalamitsa and Livadorema streams is observable in the vicinity of the rivers, where groundwater infiltrates the river streams by spring outputs.

3.3. The Chassambali karst aquifer

Karst is a unique hydrogeological environment that is characterized by features not found in any other aquifer and is controlled by processes not occurring in any other hydrogeological environment. The

term "karst aquifer" applies to terrains composed of limestone, dolomites, gypsum, halite and in general soluble rocks.[4] The geological processes that destroy soluble rocks, thus forming unique morphological features and a specific type of porosity, or a specific hydrogeological environment, are denoted by the term karstification.[5]
Karst is characterized by mixed flow conditions (turbulent and/or laminar), through joints, fractures or fracture zones, bedding planes and generally through discontinuities of the rock mass, the dimensions of which are enhanced by hydrochemical, mainly dissolution processes[6].
Because of the nature of flow in a karst domain, significant heterogeneity and anisotropy characterize the formation's hydraulic properties, both in horizontal and in vertical directions. The width and frequency of fractures caused by tectonic deformation decrease with depth and so does karstification, as LeGrand and Stringfield concluded.[7] The main characteristics of a karst terrain are high infiltration rates and rapid water movement via the discontinuities of the rock mass.

3.3.1. Characteristics of the Chassambali karst aquifer

The Chassambali karst aquifer developed in the limestones of the Pre-Cretaceous tectonic nappe of the Pelagonian zone. The karst system crops out in the southern margin of the Kalochori alluvial aquifer (Figure 3.12). The total areal extent is only 5.65 km^2. The altitude of the karst outcrop varies from 63 masl to about 445 masl. It displays a dip of 25° in northwestern direction, namely, to the Kalochori alluvial sub-basin, and its thickness is estimated to be 310 m. The basement of the limestones in the area consists mainly of serpentinites, but also of metamorphic gneissic rocks of the Pre-Cretaceous tectonic cover of the Pelagonian zone (compare Chapter 5).
Only a small water capacity aquifer developed in the Chassambali limestone outcrop, due to the limited areal extent. Practically no soil cover or vegetation exists, which suggests that infiltration rate is quite high. Impressive karst features such as caves or big caverns do not exist, implying that the karstification has developed in a homogeneous pattern rather than in a few large karst channels.
Oral information regarding the karst features of the two boreholes P1 and P4 (Figure 3.7) drilled within the karst suggest that it is well developed to a depth of 80.0 m, while below that karstification decreases and effectively stops at a depth of 110.0 m.
Therefore, the main karst outcrop of Chassambali was accepted as an individual hydrogeological entity, which only interacts with the surrounding alluvial aquifers of Kalochori and eastern Thessaly and with the surficial water of Bara Toibasi Lake. The recharge of the aquifer is dominated by rainfall infiltration on the limestone surface, by the lateral groundwater infiltration of the alluvial aquifer Kalochori, and by leakage of Bara Toibasi Lake.
As mentioned above the southern edge of the Bara Toibasi Lake is determined by the Chassambali Hills. Hydrogeological observations carried out in September 2017 confirm that at an altitude of 86.0–87.0 masl and about 800 meters in length karst features (fractures, conduits, pipes and small sinkholes) were identified, a result of karstification, that work as drainage paths. Through those karst features surface water of Bara Toibasi Lake infiltrates and feeds the Chassambali karstic aquifer, which developed within the carbonate formation. Thus, it can be concluded that the karstic aquifer works as a permeable water buffer, which creates preferential water flow paths and connects hydraulically surficial lake water and groundwater of Sykourio alluvial aquifer with:

- The karst springs that emerged at 63.0–65.0 masl altitude at the contact zone between the limestones of Chassambali karst body and the alluvial aquifer of the Eastern Thessalian Plain.
- The upper and deeper alluvial aquifer of the Eastern Thessalian Plain.

At this point it should be clarified that the strong hydraulic connection between aquifers of the Kalochori sub-basin and the Eastern Thessalian Plain, through the Chassambali karst aquifer as hydrogeological conductor, presents a completely new insight. It has emerged in the present paper

[4] Ford and Williams 1989.
[5] Milanović 1981.
[6] Ford and Williams 1989.
[7] LeGrand and Stringfield 1973.

Fig. 3.12. Chassambali limestone outcrop at the southern border of the Kalochori sub-basin with tectonic and karst features (photo by the author).

due to the study of groundwater levels in boreholes of the two basins.

It has been suggested that the three aquifers, Kalochori alluvial aquifer, Chassambali karst system and eastern Thessalian alluvial aquifer, are hydraulically interconnected, forming a wider hydrogeological system in the area (Figure 3.8). Therefore, groundwater exploitation on the aquifers of the Eastern Thessalian Plain will negatively affect the other two aquifers.

The similar water level of September 1998 in boreholes P1 and P4 of the Chassambali karst aquifer and SR72 in the Eastern Thessalian Plain alluvial aquifer (Figure 3.7) indicates the direct hydraulic connection of the two water bodies.

This is also confirmed by the extremely high positive mean correlation coefficient (r=94.03%) between the water level (1979–2006) in the boreholes P16L of the Kalochori aquifer and SR72 of the Eastern Thessalian Plain (Figure 3.13) and by the parallel long-term line on monthly water level fluctuation between these boreholes (Figure 3.14). For reasons of better comparison, in Figure 3.14 the monthly water level measurements are plotted on the same scale but with vertically offset of 20.0 m for the monthly water level of borehole SR72. Actually, despite the fact that the two aquifers are separated by the thick carbonate body of the Chassambali Hills, they are in strong direct hydraulic communication.

3.3.2. Springs in Chassambali karst aquifer

Tectonics is very important for the discharge behaviour of the Chassambali karst aquifer, which is mainly discharged by three karst springs. Kephalovrysso spring (KS.1) emerges in the southern part of the limestone outcrop, at the contact of the karst foothills with the sediments of the Eastern Thessalian Plain (Figure 3.15).

While KS.1 emerges along a NNW–SSE fault line at the lowest topographic altitudes (63.0 masl), the two others spring outlets (KS.2 and KS.3), are present along an NNE–SSW fault at an elevation of about 64.0 and 65.0 masl respectively. The presence of these springs is documented by topographic maps of the 20th century.

The existence of an underground watershed forming a separate hydrogeological sub-basin for the two

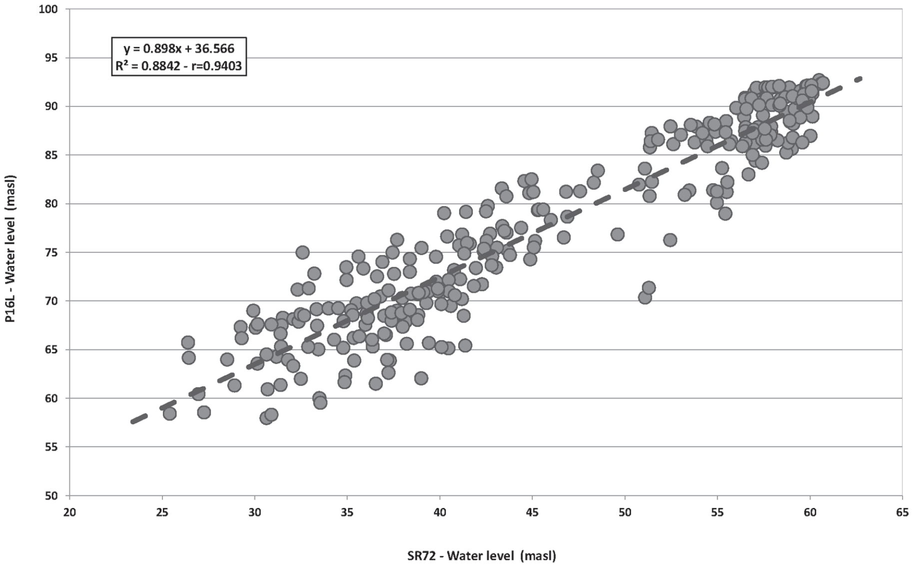

Fig. 3.13. Relation between piezometric levels in boreholes P16L and SR72 (1979–2006). The trend line (r=0.9403) is based on simple linear regression.

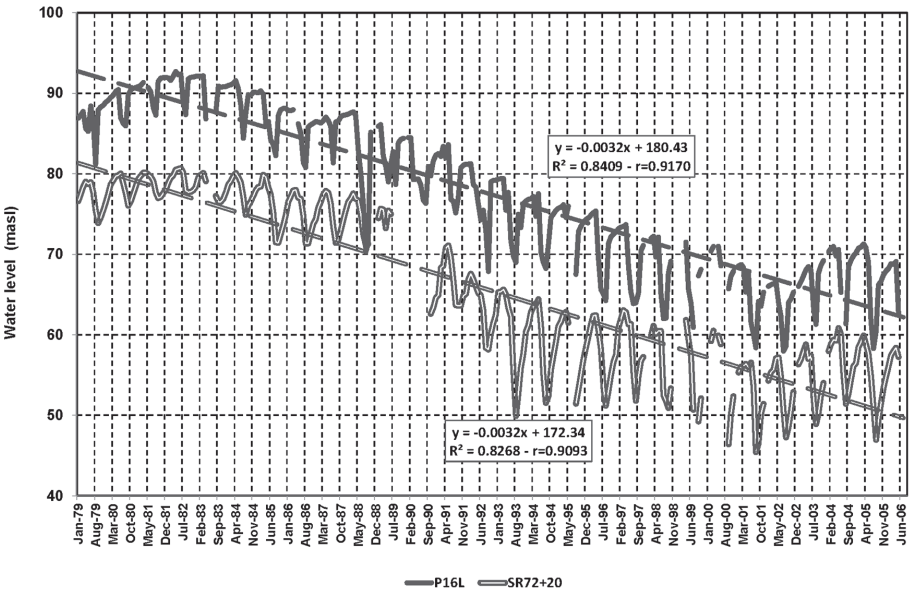

Fig. 3.14. Correlation of water-level fluctuation and long-term trend line at borehole P16L of Kalochori aquifer and at borehole SR72 of Eastern Thessalian aquifer, based on monthly measurements between 1979 and 2006 (27 years) plotted to same scale, but with a vertical offset of 20m for the boreholes SR72. The trend line is based on simple linear regression.

springs cannot be proven by the existent geological data. Nevertheless, these springs can operate even without the presence of an underground watershed. In this case, the aquifer is discharged first by the Kephalovrysso spring, while the other springs start to operate only when the karst water table rises above the altitude of the Kephalovrysso spring.

All the karst springs can be characterized as overflow type. The positioning, the common hydrogeological characteristics and the altitudes of the springs show a uniform, common catchment area for all the springs. Due to the overflowing conditions of the springs, the groundwater accumulates in the contact zone of the karst aquifer and the post-

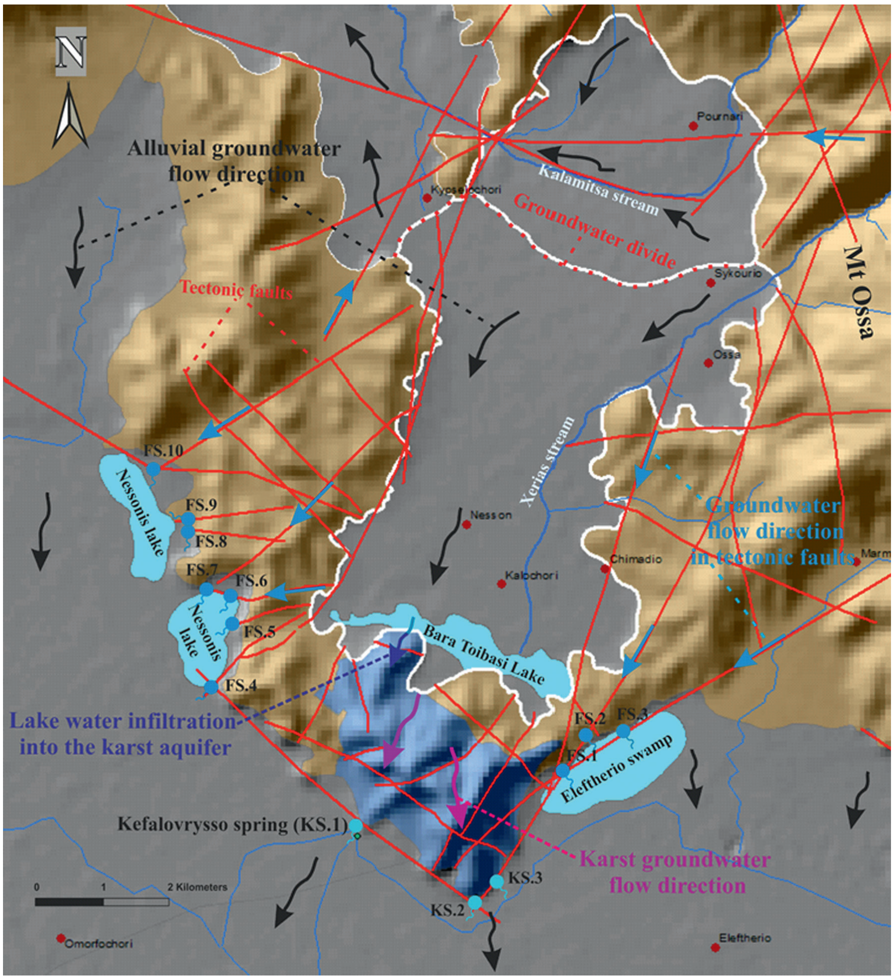

Fig. 3.15. Tectonic faults, spring outlets and swamps on the border of Chassambali karst aquifer (KS=karst spring; FS=fault spring).

alpine formations of the Eastern Thessalian Plain, which possesses lower permeability in comparison to the karst body. This structure greatly reduces the quantity of the groundwater discharging the Chassambali karst aquifer into the alluvial aquifers of the Eastern Thessalian Plain. Consequently, a minor amount of karst water drains to the alluvial aquifer of the Eastern Thessalian Plain and the remaining amounts are discharged by the springs.

From personal testimonies and photographic material of the year 1977, the particularly large discharge of the Kephalovrysso spring is documented. This occurs usually in wet years, when the annual recharge may exceed annual discharge resulting in an increase in storage and rise in the water table. These saturated conditions are achieved by the following hydrological factors:

- Precipitation increase has a relative large impact on and a strong relationship with the karst aquifer water level and consequently the increase of Kephalovrysso spring discharge rate.
- The lateral groundwater inflow derived from the Kalochori alluvial aquifer apparently causes the slightly higher groundwater rise in the Chassambali karst aquifer.
- The continuous lake water infiltration into the Chassambali karst aquifer, which we also characterize as the “leakage process”, affects decisively the water regime of the lake. It should be noted that the amount of the water infiltration depends on the lake’s water level. Thus, with an increase of the lake’s level also an increase of the lake water infiltration rate is expected, due to both the increase in lake area and the hydraulic load at sites where the karst is covered by lake water. Regarding the amount of lake water that can penetrate into the karst aquifer, it is noteworthy that, globally, many dam

reservoirs constructed on karstic limestone (for Greece, e.g., the Perdikkas Dam) were never, or only partly filled, despite extensive investigations and remedial works. Even at low water levels, water losses of 1.2 m^3/sec occurred. For example, of the 46.6 hm^3 of water flowing into the reservoir of the Perdikkas Dam between September 1962 and March 1963, it lost 32.2 hm^3; the leakage was accordingly approx. 69% of the inflow.[8] It means that also the water of the rising Bara Toibasi level that penetrates into the Chassambali karst body causes an increase in the amount of the groundwater level in the karst acquifer. This, in turn, raises the operation of the karst springs (KS.1–3).

3.3.3. Fault springs around the Chassambali karst aquifer

Chassambali karst aquifer is bounded by serpentinites in the east and schists in the west. Usually, aquifer systems of low transmissivity are developed in faulting zones, where impermeable rocks, such as serpentinites, show secondary porosity. The result was the emergence of three fault springs (FS.1–FS.3) of minor discharge in serpentinites located at the northeastern border of the Chassambali karst aquifer, along the fault line in NNW–SSE direction (Figure 3.15).
Thus, these springs are associated with faults that are aligned with the inferred direction of groundwater flow. According to observations on topographic maps of the 20th century it is obvious that the springs enriched the surface water of an old swamp NW of Eleftherio.
The same hydrogeological properties are presented in the area NW of the Chassambali karst aquifer. A large number of minor discharge fault springs (FS.4–FS.10) emerged in schists and recharged in the former Lake Nessonis in the northwestern part of the Eastern Thessalian Plain, close to the Pinios River.

[8] Kleinsorge 1972.

3.3.4. Cessation of the Kephalovrysso spring

Discharge measurements of Kephalovrysso spring (KS.1) have not been recorded systematically in the past, due to the low water capacity of the aquifer. Therefore, an exact explanation why the Kephalovrysso spring flow completely ceased cannot be offered.
As mentioned above, the eastern Thessalian aquifer receives amounts of karst water through lateral inflow. Thus, the hydraulic relation of the two water systems is considered certain, and a hydraulic equilibrium is reached between the inflow of water from the eastern Thessalian alluvial aquifer and the discharge of the Chassambali karst aquifer. For this reason, reliable information for the spring's cessation is obtained from borehole SR72 at a distance of only 4.6 km. It penetrates into the eastern Thessalian aquifer, and water level fluctuations for the period 1975–1989 can be studied and compared to the Kephalovrysso spring outlet at 63.0 masl.
In Figure 3.16 the hydrograph of the borehole SR72 during 1975–1989 shows that the initial water table in the years 1975 and 1976 lay at very shallow depths. The difference of the borehole water level and the altitude of the Kephalovrysso spring vary between 1.45 m and 2.54 m. In this period Kephalovrysso spring was in operation throughout the whole year.
In the following years (1977–1979) the water level showed a decline that has been obviously affected, as presented in the above figure, by less precipitation. During this time the Kephalovrysso spring emerged only in the wet months of the year. This can be documented from oral testimonies and photos provided by locals, showing women washing their carpets in Kephalovrysso spring in the springtime, until at least 1977.[9]
The water level fluctuation remained unchanged between 1980 and 1985. In the years 1986–1988 increased water exploitation had reduced the piezometric surface in borehole SR72 by about 2.0 m, and in 1989 it showed another drawdown by 2.5 m. Most of this regional decline occurring

[9] We would like to thank Theodoros Paliougkas for providing us with these photos with clear indications regarding the year and the season in which they were taken.

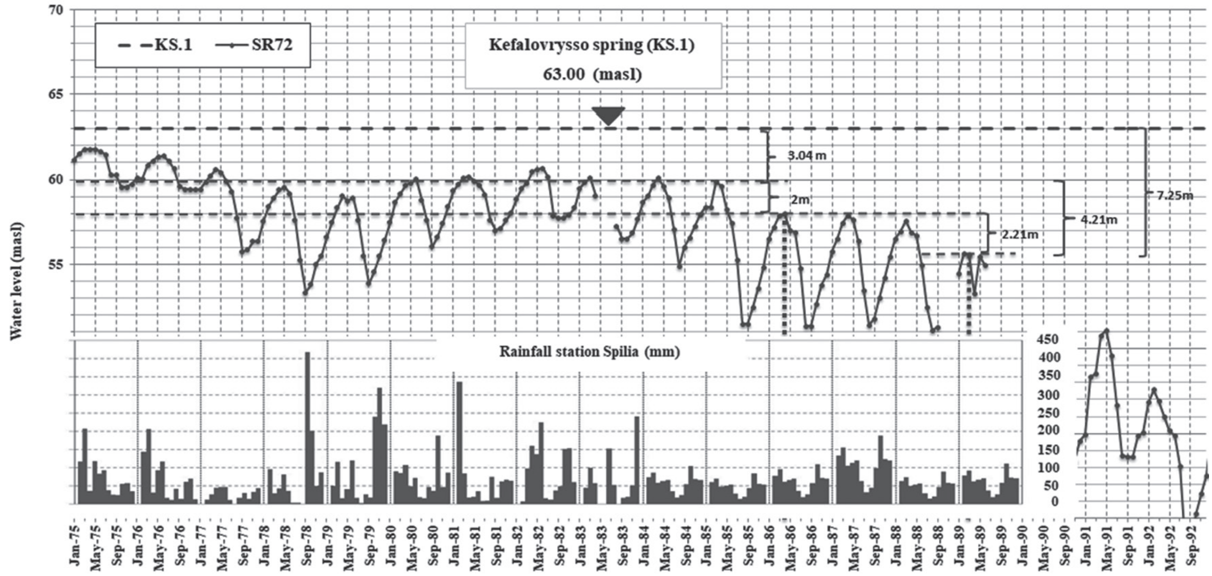

Fig. 3.16. Groundwater level fluctuations (masl) in borehole SR72 plotted against rainfall (mm) at the Spilia station for the period 1975–1989.

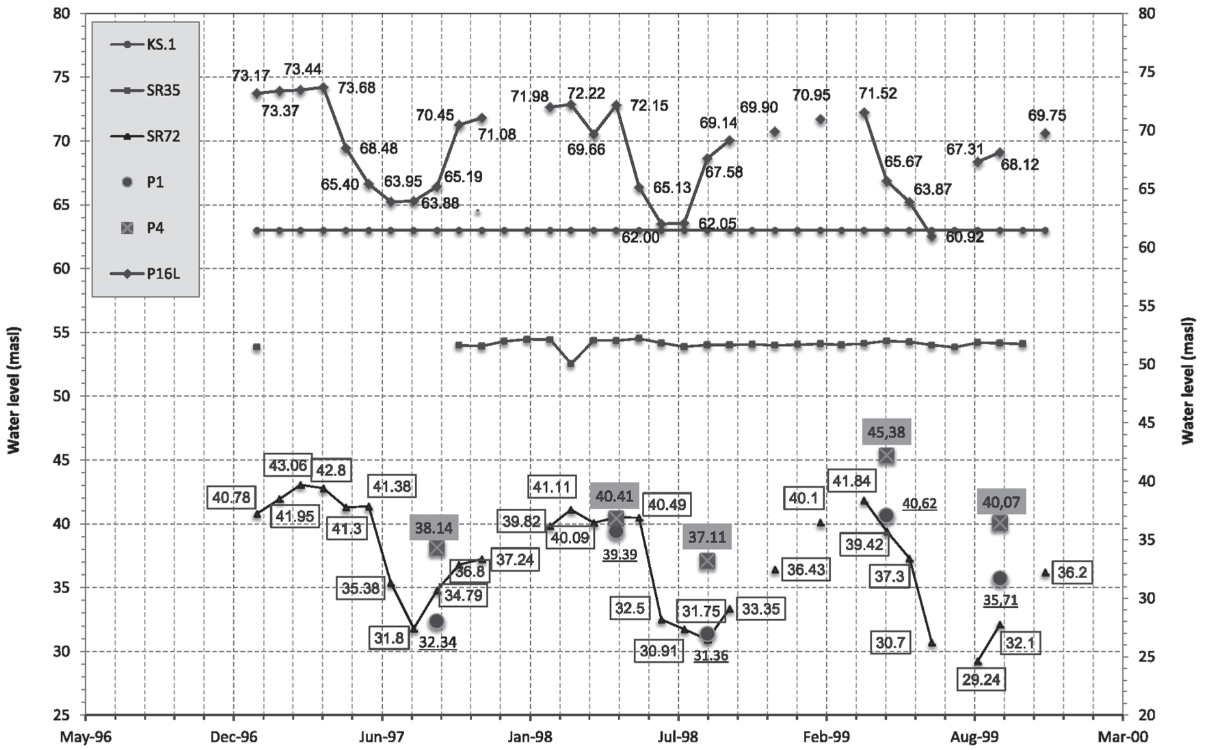

Fig. 3.17. Monthly groundwater level fluctuations (Jan.1997–Dec.1999) in boreholes of Kalochori (P16L), Eastern Thessalian alluvial aquifers (SR35, SR72) and Chassambali karst aquifer (P1, P4). In the graph is also depicted the outlet elevation of Kephalovrysso spring (KS.1).

after 1986 is related to the excessive groundwater inception, leading to the complete flow cessation of Kephalovrysso spring. This occurs despite monthly rainfalls, yet did not present significant variabilities from the average values between 9.0% and 1.3% in the years 1986–1992, as measured at the rainfall stations of Larissa and Spilia. This stands in complete accordance with the conclusion by Panagopoulos[10], who defined the intensification of the groundwater resources exploitation in the alluvial basin of Tirnavos during the year 1986.

Based on the water level fluctuation characteristics of borehole SR72, the Kephalovrysso spring flow can be divided into two stages:

- first period (1975–1985), when the spring flow was affected only by climate variation;
- second period (1986–1989), when the flow was affected by both climate variation and anthropogenic activities.

Therefore, the data processing and analysis of the water level fluctuation in borehole SR72 led to the understanding of the water flow within the karst aquifer and the behaviour of the Kephalovrysso spring.

This is confirmed by Figure 3.17, where the water levels in boreholes P1 and P4 of the Chassambali karstic aquifer during the dry and wet periods of the years 1997–1999 show values similar to the SR72 levels of the eastern Thessalian aquifer.

3.4. The catchment of the Pinios River

To the catchment of the Pinios River, west of Mt Ossa and south of Mt Olympos, belong two alluvial basins: Elateia and Gonnoi (Figure 3.18).

3.4.1. The Elateia basin

The alluvial basin of Elateia forms the western part of Mt Ossa and contains important quantities of groundwater.

The watershed of the Elateia hydrological basin has an elongated NW–SE direction and occupies an area of 67.4 km^2, with a minimum and maximum elevation of about 14.0 masl, and 959.0 masl respectively.

Steep to very steep slopes of the Ossa mountains cover 55.6% of the total area of the watershed, especially in its northeastern part. Flat to gentle slopes cover 44.4% of the total area of the watershed and have a slope inclination of 1.4–2.0%. The Elateia aquifer extends over 30.0 km^2 and has an average thickness of 74.0 m.

Ephemeral streams flow out of Mt Ossa usually only during wet months of the year, and they discharge into the plain of Elateia. Kalamitsa is the major stream, it has an intermittent character and flows at gentle slopes, before discharging into the Pinios River. The Kalamitsa enters the Elateia basin from the southeast through the gorge of Bounarbasi. Large volumes of water from the surface runoff of Kalamitsa feed the alluvial aquifers of the Elateia basin, and therefore only a minor amount of its water reaches and recharges the Pinios River.

The margins of the Elateia alluvial basin are formed in the east, south and west by Paleozoic gneisses and schists. The sediments of the Gonnoi basin form the northern basin margin, where the course of the Pinios river represents the underground watershed for the Elateia and Gonnoi aquifers. Additionally, they contain a thick sequence of Quaternary alluvial deposits.

The alluvial aquifer of Elateia basin can be characterized as an aquifer stress area, where water demand is high, and large numbers of high-capacity wells extract groundwater in close proximity.

3.4.1.1. Structure

The Elateia basin contains alluvial deposits of mixed fluvial, lacustrine and terrestrial origin. The thickness of the Quaternary alluvial deposits exceeds 160.0 m in its central part, and some trends exist with regard to the distribution of their lithology.

An extensive aquifer system is present in the alluvial deposits of Elateia basin. Lithological sections of two boreholes, drilled in the years 1975 (borehole 17L) and 1976 (borehole 18L) and obtained from the Land Reclamation Services are illustrated in Figure 3.19, where borehole 18L is located in the central part of the alluvial basin and 17L is near its northwestern margin. Borehole 18L reached the bedrock of the basin at a depth of 114.0 mbls and borehole 17L at a depth of 86.0 mbls.

The drill holes that were used in this study produce a suitable spatial distribution of alluvial basin-fill stratigraphy and the presence of high-quality lithologic descriptions that could be interpreted in

[10] Panagopoulos 1995.

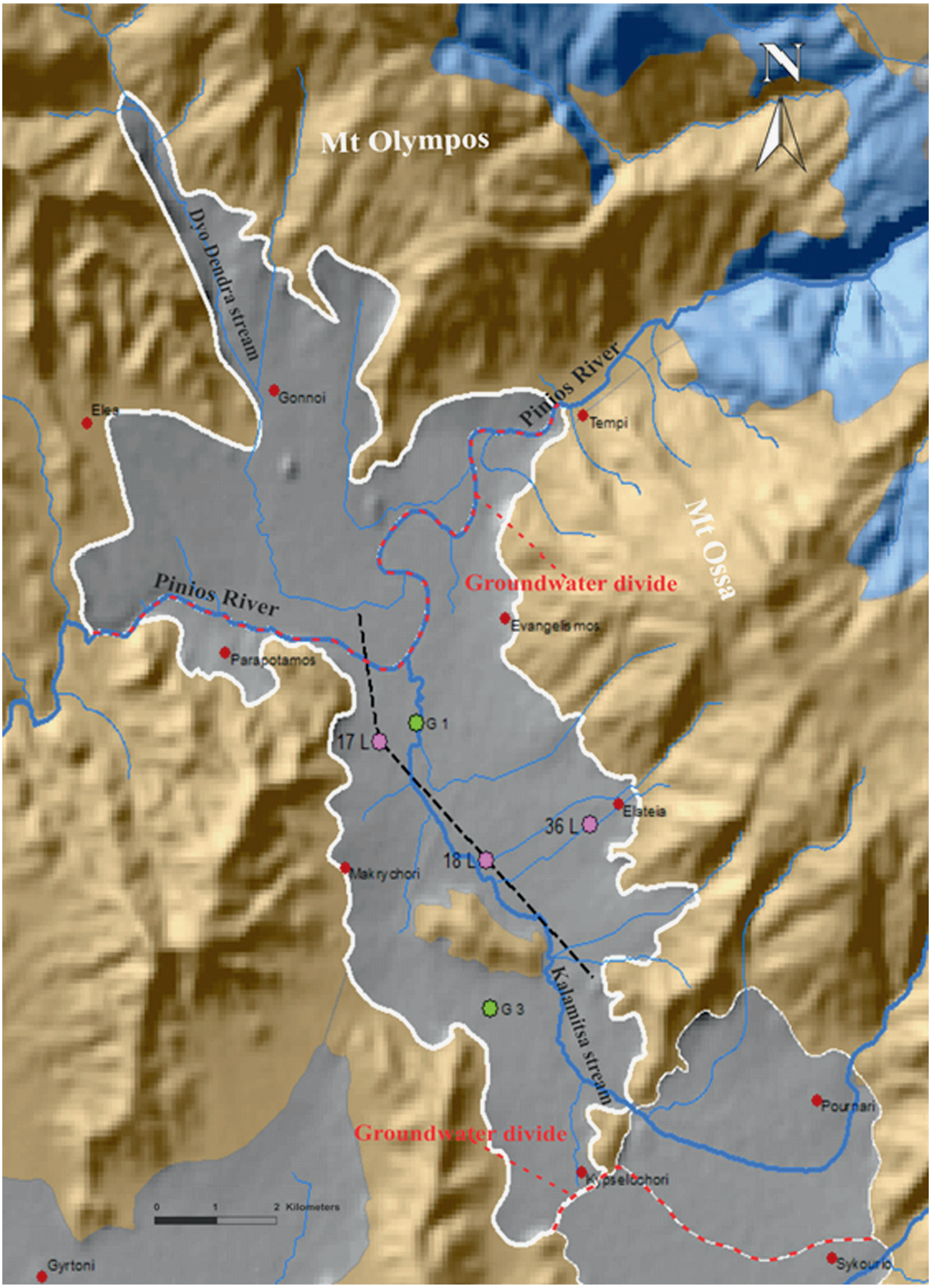

Fig. 3.18. Alluvial basins of Elateia and Gonnoi. Locations of boreholes and directions of lithological and hydrogeological sections are illustrated.

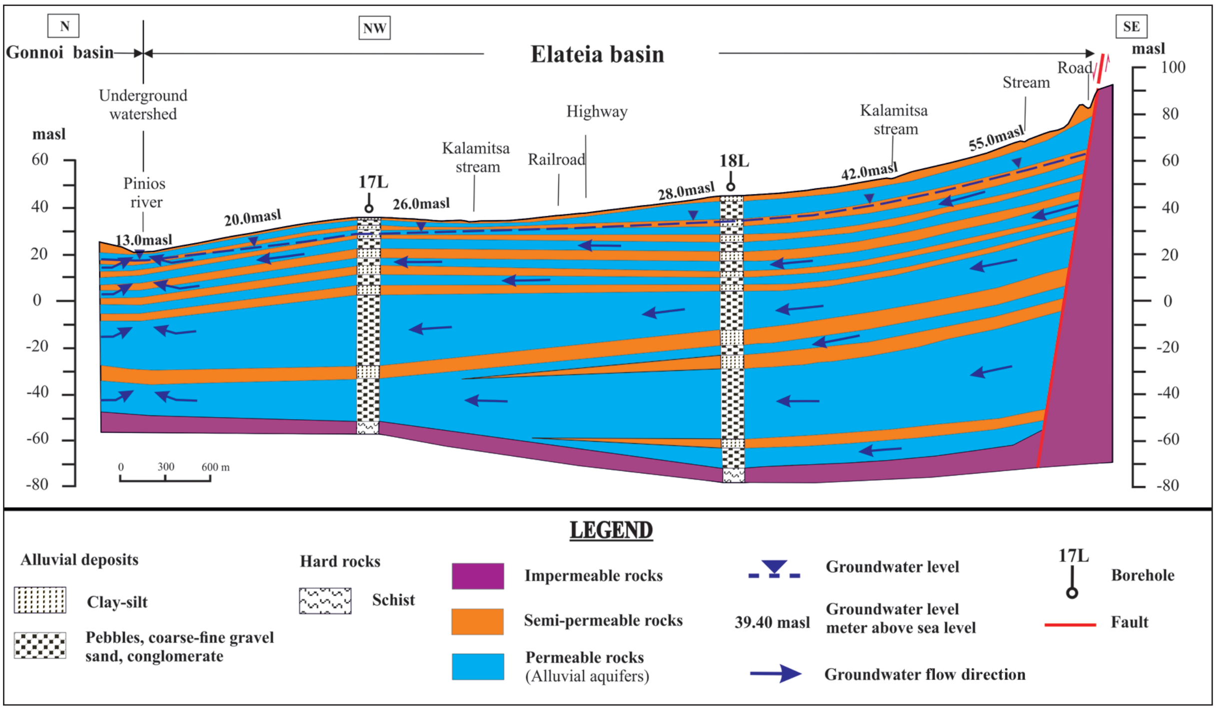

Fig. 3.19. Hydrogeological section across Elateia alluvial basin. The exact direction of the section and the locations of the two boreholes are indicated in Figure 4.1.

terms of depositional environment and facies information.

Examination of the lithological sections in the upper parts of the basin showed an alternation from thin (2.0–6.0 m) clay and silt deposits serving as aquitards-semi aquitards layers and thin (4.0–5.0 m) deposits of coarse materials (pebbles, coarse-fine gravel sand and conglomerates) as water bearing layers. This sequence was identified in the two typical lithological sections of the studied boreholes until the depth of about 40.0 mbls.

In deeper parts of the basin these bands of water bearing gravel and sands increase in thickness (50.0–57.0 m) while the thickness of the aquitards, semi-aquitards layers with clay and silt deposits remains the same as in the upper part. Crystalline schist also forms the bedrock of the basin. The Elateia aquifer thus contains important quantities of groundwater. Based on the compiled geological map and the examination of the alluvium lithology, a geological section was constructed across the Elateia basin (Figure 3.19). The exact direction of the section along with the location of the considered boreholes are illustrated in Figure 3.18. The different deposits act as one aquifer with direct vertical hydraulic contact, i.e. there is no hydraulic separation into separate aquifers.

3.4.1.2. Groundwater fluctuation

The study of the water level fluctuation was based on the water level inventory kept by the Land Reclamation Services in of Larissa. Data records are available for two monitoring points: borehole G1, near Parapotamos, and borehole G3 in the Makrychori area. They cover the period 1994–2006 (13 years). A map showing the position of the monitoring points on the aquifer is illustrated in Figure 3.18. Individual hydrographs for each borehole are presented in Figure 3.20. The water table in borehole G3 was fairly shallow, lying at a depth between 3.0 to 21.0 mbls (1994–2006), as compared with the depth of the water table in borehole G1, which was between 8.0 to 15.0 mbls (1994–2006).

As may be expected, the annual variation of groundwater level in borehole G3 is large compared with the hydrographs of the borehole G1 (Figure 3.20). This likely indicates a particularly low specific yield and greater amount of groundwater abstraction in borehole G3. Borehole G3 has

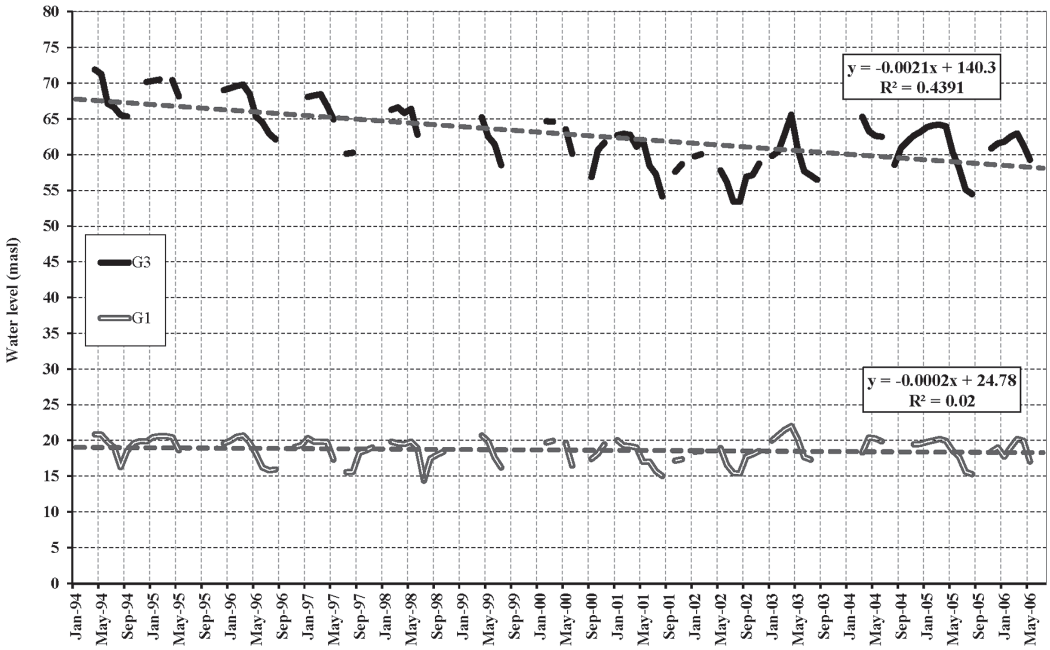

Fig. 3.20. Long-term hydrographs for the boreholes in Elateia alluvial basin. The trend lines are based on simple linear regression between water level and time.

a higher annual variation (average of 7.5 m), compared with the borehole G1 (average of 3.8 m).

3.4.1.3. Piezometry in the Elateia basin

The alluvial aquifer systems of Elateia can be characterized also as aquifer stress areas, where water demand is high and where relatively large numbers of high-capacity boreholes are extracting groundwater in close proximity.

Groundwater level measurements were made in August 1971 within a network of 30 boreholes (public, municipal, and private), distributed throughout the basin in order to determine the piezometric surface of the Elateia aquifer. The compilation of the piezometric map was based on the up to now unique map created by SOGREAH. The contour lines of the piezometric map relate to the general picture of the cumulative piezometry of all the individual pressurized aquifers.

From the course of the piezometric curves, two separate hydrogeological basins, Elateia and Gonnoi, can be identified. The Pinios River represents the underground drainage line of the aquifers for both basins (Figure 3.10).

The following conclusions are drawn from the study of the piezometric map of the Elateia aquifer. Figure 3.10 shows the piezometric map of the Elateia aquifer with a 20-m interval of subsequent piezometric lines.

- The main features of the piezometric surface morphology of Elateia alluvial basin are similar to those of Kalochori and Pournari. Recognisable are waveforms of the piezometric contours with convergent and divergent flow lines.
- The groundwater recharge is through surface water infiltration by the Kalamitsa stream, direct percolation of rainfall, groundwater inflow along the southeast, southern, and partly from eastern and northeastern boundaries, and infiltration of irrigation water.
- The groundwater level ranges from 80.0 to 15.0 masl.
- The piezometric contour of 80.0 masl in the southeastern border of the Pournari subbasin is continued into the upper part of the Elateia basin. This implies an existing strong hydraulic connection between the two alluvial basins.
- Groundwater level curves of the piezometric map clearly indicate an aquifer discharge axis from SE to NW. This axis roughly follows the Kalamitsa stream course, which is the

main stream of the hydrographic network of Elateia alluvial basin, and drains towards the river Pinios floodplain, where it discharges by lateral leakage into the riverbed sediments.

- In the southeastern part of the basin, near Magoula Bounarbasi, the main recharge is carried out by the surface water runoff of the Kalamitsa stream and shows hydraulic gradient values, ranging between 1.28% and 1.68%.
- In the central part of the Elateia basin the hydraulic gradient has an average value ranging between 2.26% and 2.97%.
- In the northwestern part of the basin, at the northwestern hydraulic limit near the river Pinios, the hydraulic gradient of the alluvial aquifers has an average value ranging between 0.53% and 0.58%.
- The fact that the piezometric lines near the Pinios River show a vertical offset towards a northeastern direction in both aquifers of Elateia and Gonnoi and follow the current riverbed (which means a flow parallel to the river course), indicates the presence of common underground drainage lines that follow the current riverbed. A small amount of groundwater that discharges into the Pinios River forms the base flow component of the river's surficial runoff.
- The potentiometric contours indicate a high gradient to the central part of the basin, which probably is a result of the large thickness and hence high transmissivity of the aquifer.
- The aquifer is fairly uniform across the geological formations of the Elateia basin.
- The southwestern borders of the basin are formed by impermeable Paleozoic gneisses. Thus, between the aquifers of the Elateia basin and the Eastern Thessalian Plain no hydraulic connection exists.

3.4.2. The Gonnoi basin

The Gonnoi alluvial basin forms the northwestern part of the Pinios catchment area and contains important quantities of groundwater. The watershed of the Gonnoi hydrological basin has an elongated NW–SE and a NE–SW direction in the eastern part of the basin near the Pinios River. Its natural basin occupies an area of 77.4 km^2 with a minimum and maximum elevation of about 14.0 masl and 1402.0 masl. Steep to very steep slopes of Mt Olympos characterize the northwestern part and cover 69.1% of the total watershed area. The remaining parts of the basin are characterized by flat to gentle slopes, which cover 30.9% of the total watershed area with a slope inclination ranging between 2.7–4.5%. The Gonnoi aquifer extends over 23.9 km^2, but its thickness is unknown due to the lack of borehole logs.

Ephemeral streams flow out of Mt Olympos usually only during wet seasons of the year and discharge into the plain of Gonnoi. The major stream "Dyo Dendra" has an intermittent character and flows on gentle slopes, before discharging into the Pinios River.

The margins of the Gonnoi alluvial basin, which contains an unknown thick sequence of Quaternary alluvial deposits, are formed in the northeast, north and west direction by Paleozoic gneisses and schists and by the sediments of the Elateia basin in the southeast, where - as previously mentioned - the course of Pinios River presents the underground watershed both for the Gonnoi and the Elateia alluvial basins.

Public, municipal and private boreholes intersect the Gonnoi alluvial aquifer. The data used to evaluate the hydrogeology of Gonnoi basin were primarily generated during the 1970s. Most of the current hydrogeological understanding was obtained only from the piezometric map of August 1971, established by SOGREAH. Thus, it is difficult to use these results to assess the present day hydrogeological conditions of Gonnoi basin.

3.4.2.1. Piezometry in the Gonnoi basin

The following conclusions are drawn from the study of the piezometric map (Figure 3.10) of the Gonnoi aquifer.

- The groundwater level ranges from 150.0 to 15.0 masl.
- The recharge is through surface water infiltration by the "Dyo Dendra" stream, direct percolation of rainfall and infiltration of irrigation water.
- Groundwater level curves of the piezometric map clearly indicate an aquifer discharge axis from NNW to SSE. This axis follows the course of "Dyo Dendra" stream, which is the main stream of the hydrographic network of Gonnoi alluvial basin and drains towards the floodplain of the river Pinios.

- North of Gonnoi municipality, the main recharge is carried out by the surface water run-off of unnamed streams and shows the largest hydraulic gradient values, ranging between 4.67% and 6.21%.
- The central part the hydraulic gradient has an average value ranging between 2.79% and 4.18%, while in the southern part of the basin, near the Pinios River, the hydraulic gradient average value ranges between 0.85% and 2.57%.
- From the above can be concluded that the values of the mean hydraulic gradient in the two alluvial basins drained by the river Pinios indicate a completely different hydraulic pattern. For the Elateia basin it amounts to 1.46%, while in the Gonnoi sub-basin it is at 3.61%.

3.5. Some final remarks regarding the hydraulic connection between the aquifers of the study area

In the previous chapters the hydrogeological conditions prevailing in aquifers of the study area were examined. From the detailed study mainly of the piezometric maps but also from comparisons of the groundwater level fluctuation in boreholes, important conclusions emerged regarding the hydrogeological characteristics of the alluvial aquifer systems and their hydraulic relations to the surrounding formations.

Studies of the piezometric maps reveal the main flow paths in the hydrogeological units and possible hydraulic interaction between the aquifers under study and those from the surroundings, whilst parallel the main recharge-discharge areas are depicted.

Figure 3.10 shows the groundwater flow direction of the aquifers, which allow us to identify and evaluate the hydraulic communication between them.

The piezometric map of the Sykourio alluvial basin compiled by SOGREAH in August 1971 defines two hydrogeological sub-basins of Kalochori and Pournari.

Groundwater level curves in the piezometric map of Kalochori sub-basin clearly indicate an aquifer discharge axis from NE to S. The southern border of the alluvial aquifer consists of the limestone of the Chassambali karst aquifer.

The karstic features, due to the well-developed karstification, form favourable hydraulic conditions for handling the large lateral hydraulic loads of the Kalochori alluvial aquifer towards the Chassambali karst aquifer.

The karst aquifer is mainly discharged by three springs. Kephalovrysso is the major karst spring and emerges in the southern part of the limestone outcrop, on the contact zone of the karst foothills with the sediments of the Eastern Thessalian Plain. This situation creates hydraulic interactions between the two hydrogeological units, discharging amounts of karst water through lateral inflow into the eastern Thessalian aquifer, due to a strong hydraulic communication existing between the alluvial aquifer system and the Chassambali karst system to the northeast.

Groundwater level curves of the Pournari piezometric map clearly indicate an aquifer discharge axis from NE and SE to NW. This axis follows the Kalamitsa stream course discharging towards the Pinios River, whose course represents the NW and SE hydraulic limit for both the Elateia and the Gonnoi alluvial aquifers. Recognisable across this hydraulic limit is a groundwater flow diversion of almost 90 degrees in NE direction, feeding most likely the Lower Olympos–Ossa karst aquifer.

The piezometric contour of 80.0 masl in the southeastern border of Pournari sub-basin is continued into the upper part of the Elateia basin pointing to a strong hydraulic connection between the two aquifers.

Finally, the southwestern and western borders of the Elateia basin are formed by Paleozoic gneisses. Thus, no hydraulic connection exists between the aquifers of the Elateia basin and the Eastern Thessalian Plain.

3.6. Bara Toibasi Lake: the prehistoric sites and the historical lake water balance

3.6.1. Bara Toibasi Lake and the prehistoric sites

As mentioned above, the Kalochori alluvial sub-basin forms the central and southwestern part of the Sykourio basin and contains important quantities of groundwater. The watershed of the

Kalochori hydrological sub-basin has an extent of about 81.7 km^2, and the alluvial aquifer covers an area of 22.4 km^2. Xerias (or Megalo Rema) is the major stream with a torrential intermittent character. Surface runoff from the watershed discharging via the Xerias stream supplied the lake, but only a minor amount of water reached and recharged Bara Toibasi, probably rather in wet winters. The lake is a closed-catchment lake with no surface water outflow in the lowest southern part of the Kalochori sub-basin.

The prehistoric sites around the Bara Toibasi Lake do not indicate a random distribution over the prehistoric period, but form clusters suggesting an alternation of lower and higher, climatically driven lake-level phases. They provide evidence of a rather unstable climate (Figure 3.21).[11]

Based on the spatial distribution and the chronology of the investigated archaeological sites (for a more detailed study on this topic compare Reingruber, Chapter 6) the extent of the Lake should not be considered as stable and permanent throughout the Neolithic. Certainly, without knowledge derived from pilot cores (e.g. geochemical, sedimentological and physical investigations), one cannot safely establish either the lake's contours nor its changing depths and storage volume.

Concerning the extent of Bara Toibasi Lake at a recent historical time frame the following should be mentioned. Maps from different decades of the 20th century indicate that Bara Toibasi Lake has been subdued to considerable fluctuations. It seems that from the beginning of the 20th century until 1980 it was always a shallow lake with water altitude ranges between 87–90 masl. In this time period Bara Toibasi Lake served as a natural reservoir holding a large water volume. Therefore, its size and depth varied through the ages and has not always covered the same surface within the basin.

Nowadays Bara Toibasi Lake has become almost or completely dry, with only a few stagnant pools of water remaining. Water diversion for irrigation and agricultural use as well as drainage have affected the area mainly since modern times (after 1980).

3.6.2. Bara Toibasi Lake and the water balance for the period 1971–1980

3.6.2.1. Conceptual Approach and Methodology

An interesting question concerning the extent of Bara Toibasi Lake at a recent historical time frame arises as to what could be the water level fluctuation (min, max and average) of the lake during the period 1971–1980. In order to approximately reconstruct the lake's hydrological conditions, it was considered appropriate to establish the lake's physical characteristics for this period.

To calculate the surface area (A) and the storage volume (V) of Bara Toibasi Lake, in relation to the different water altitudes (H) of the lake for the period 1971–1980, a digital terrain model was created using digital maps of the Land Registry in a scale of 1:5000.

Paper topographic maps produced by HMGS (Hellenic Military Geographical Service) and geological maps produced by IGME (Institute of Geology and Mineral Exploration) were georeferenced and important features (contours, streams, lake shore, and extent of alluvial deposits) were digitized. Hence, by using the digitized contours, streams and lake boundaries, the digital terrain model – Triangular Irregular Network (TIN) – was constructed. This model, which includes the altitude information of the area, was generated with the ArcGIS 10.3 spatial analyst package.

Subsequently, several contour lines that may have corresponded to the lake's levels were used to determine the lake surface area, and a lake bathymetry map was derived by spatial interpolation from the lake's levels. Finally, from the bathymetric map, the area and volume corresponding to different lake depths were calculated using the "surface volume" tool in ArcGIS 3D analyst. Thus, the main characteristics of the lake are presented in Figure 3.22 in terms of the area it covers (A) and the stored volume of water (V) in relation to the altitude of the lake water level (H). In Figure 3.23 the relationship between storage volume (V) and lake surface area (A) is attributed. The specific Level-Area-Volume relationships for Bara Toibasi Lake have also been calculated.

According to Figures 3.22 and 3.23 for Bara Toibasi Lake the power and quadratic polynomial functions are most suitable to describe their H–A, H–V and V–A relationships, respectively.

[11] Reingruber and Toufexis 2021.

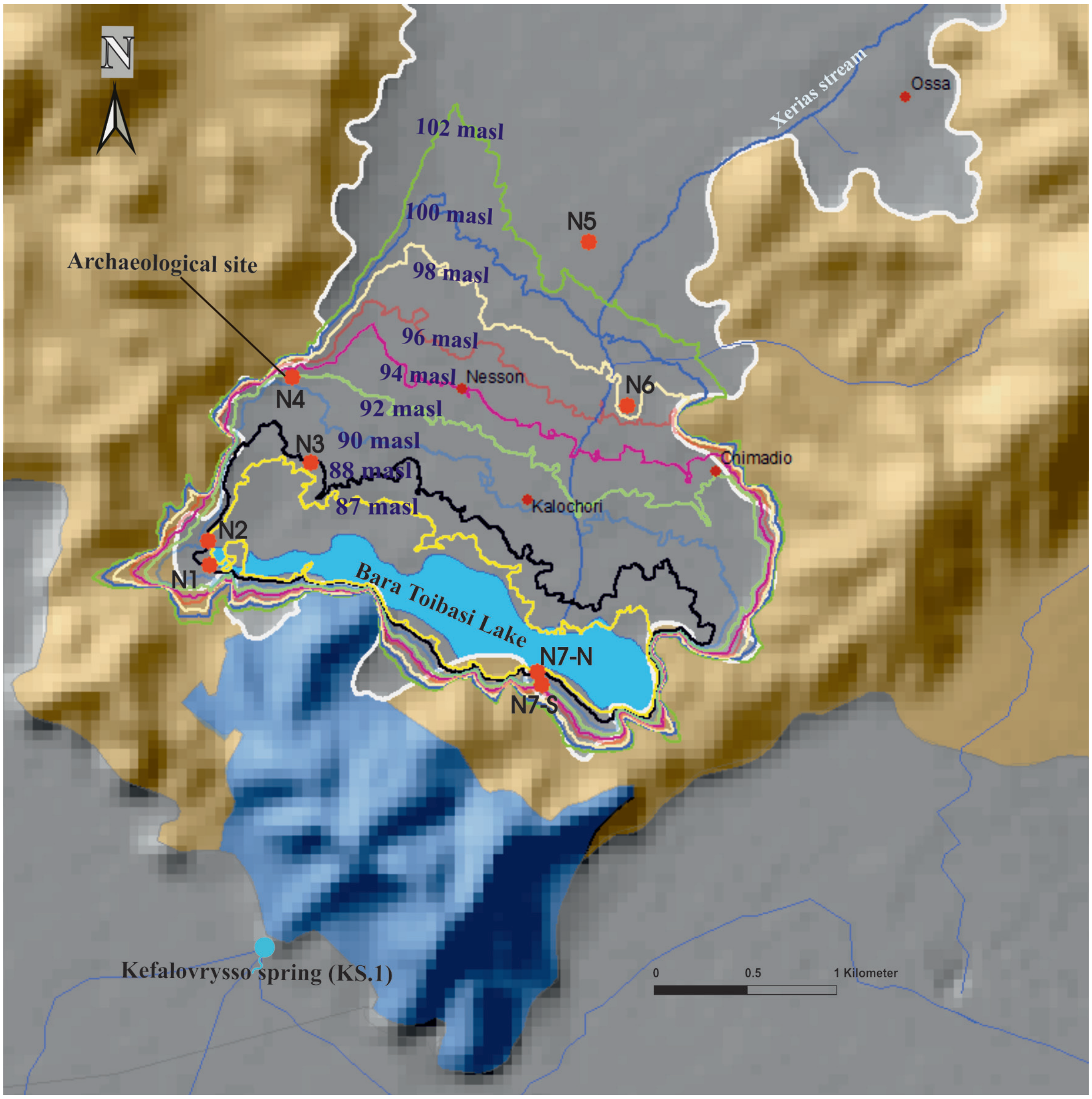

Fig. 3.21. Location of archaeological sites and Bara Toibasi Lake extent (year 1945) in relation to the water level altitude of the lake indicated and labelled in masl.

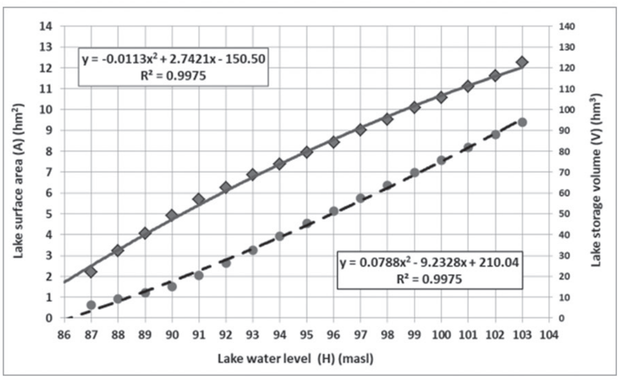

Fig. 3.22. Lake surface area (A) and storage volume (V) in relation to the water level (H).

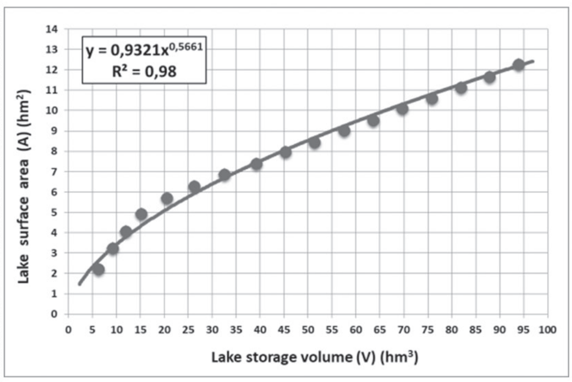

Fig. 3.23. Relationship between the lake storage volume (V) and surface area (A).

3.6.2.2. Lake water balance for the period 1971–1980

Water balance studies have been extensively implemented to make quantitative estimates of water resources. Water balance helps to quantitatively evaluate the contribution of individual sources of water in the system in time and space, and it studies the degree of variation in water regime due to changes in components of the system.
Studying the water balance is to determine whether the input balances the output or creates a positive or negative stock in order to gain a deeper understanding of the lake's hydrology. Unfortunately, the number of valuable hydrological proxies from the Kalochori sub-basin that could provide accurate water balance estimation is insufficient. Concerning the question of what could be the maximum level of the lake, we tried to approach this indirectly by establishing the water balance of the lake for the 1970s (1971–1980). The lake water surface area (A) for the year 1970 was determined first, deriving from the SOGREAH piezometric map of 1971, considering that this map depicts the spatial hydrological condition of the lake for the year 1970.
The method used to estimate the extent of the lake water surface area (A) was the careful digitization, which admittedly is the most accurate, although it is time consuming and tedious. The underlying lake topography is relatively uniform and varies slightly. This reduces the error rates in the initial assessment of the lake area, altitude and storage volume.
The initial hydrological conditions, i.e. the lake surface area (A), altitude (H) and storage volume (V) for the year 1970, were established using regression analysis obtained by the derived equation of the relationship between the aforementioned water balance components (Figures 3.22–3.25).
In order to clearly explain how to perform such analytical integration, let us imagine that the volume (V) – area (A) relation can be described as a second-polynomial function:

$$V=f(V)=aA^2+bA+c$$

where V is the volume in hm^3, A is the surface area in km^2 and a, b, c are coefficients determined by regression analysis.
As Bara Toibasi Lake is a closed catchment with no surface water outflow, the annual hydrological water balance equation can be expressed as follows (Table 3.1):

$$\Delta h=P_L-E_L+R-I_K$$

Where Δh: is the yearly water level variation (mm),
P_L: is the yearly precipitation on the lake surface (mm),
E_L: is the yearly evaporation from the lake surface (mm),
R: is the yearly surface runoff into the lake (mm),
I_k: is the yearly infiltration into the karst aquifer (mm).

Compiling the hydrological balance of the lake the elements that we took into account are the following:

- Ideally, *in situ* measurements of each water balance component should provide a complete understanding of inflow, storage changes and outflow of Bara Toibasi Lake.
- The water area of Bara Toibasi Lake was converted to water level and storage volume using the area (A)–level (H) and area (A)–volume (V) relationships as developed above.
- Any water that may potentially accumulate in the lake will be a result of precipitation.
- Monthly total rainfall data were obtained from the Spilia rain gauge station, which is installed in the catchment area of the lake.
- The total water inflow into Bara Toibasi Lake was calculated using the catchment runoff coefficient resulting from the hydrological balance of studies carried out in similar alluvial basins in the region of Thessaly.[12]
- The amount of water evaporating from the lake's surface was calculated with a coefficient from a study carried out in reservoirs and shallow lakes of Larissa prefecture.[13]
- Water abstraction from Bara Toibasi Lake for the period 1971–1980 was assumed negligible.
- In the annual hydrological water balance equation groundwater inflow and outflow from or to the alluvial aquifer of Kalochori sub-basin were set to zero due to the thick firm (~40 m) clay at the bottom of the lake.

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[12] Manakos 2001; 2010.
[13] Kotsopoulos et. al 2006.

Annual height (mm)	Driest year (1977)	Average year (1976)	Wettest year (1979)	Hypothetical case 1	Hypothetical case 2	Hypothetical case 3
P_L	329.9	821.4	1185.9	2000.0	2500.0	3000.0
E_L	1226.5	2140.0	2533.5	3355.4	4024.6	4508.4
R	1259.0	1796.4	2190.8	2789.6	2910.1	3114.4
I_K	251.8	359.3	438.2	557.9	582.0	622.9
Δh	110.6	118.5	405.1	876.3	807.5	983.2
Lake altitude (masl)	86.49	87.80	88.56	90.45	92.25	93.68
Lake area (km^2)	2.78	4.86	5.75	7.62	9.13	10.23

Tab. 3.1. Components of Bara Toibasi Lake water balance according to the hydrological data for the period 1971–1980.

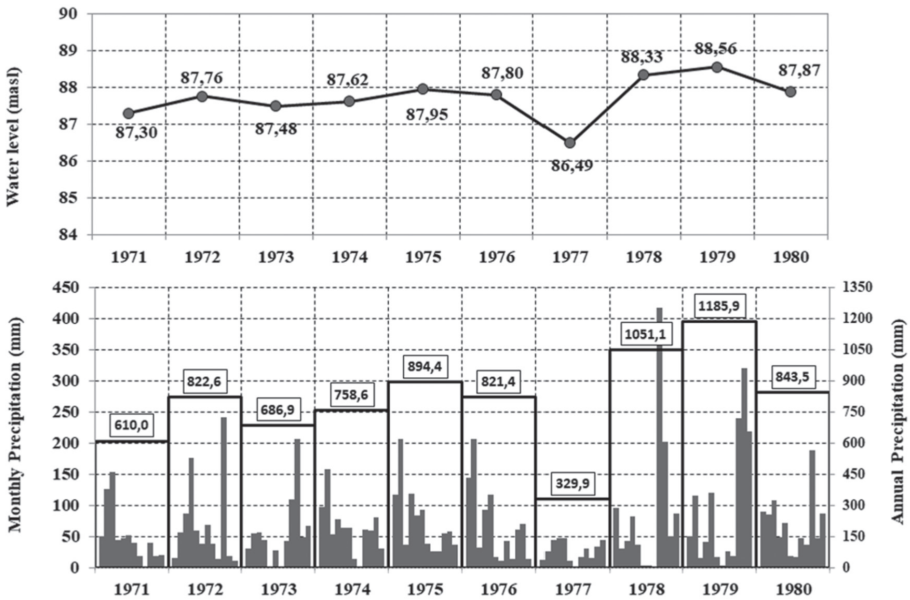

Fig. 3.24. Range of predicted annual water level fluctuation (masl) of Bara Toibasi Lake compared with monthly and annual total rainfall record (mm) at Spilia station (1971–1980). Data shown are for periods of no significant human intervention in the lake's hydrology.

- There was no discharge to any existing surface water bodies or courses as a result of extraction activities.
- The inflow into Bara Toibasi Lake was estimated as the surface runoff from all catchments draining into the lake.
- Water from the lake that enters the Chassambali karst aquifer is probably the most difficult water balance component to estimate due to the requirements of laborious and costly measurements.
- The amount of lake water infiltrated into the Chassambali karst aquifer (leakage process), which determines the altitude of the lake to a large extent, was calculated as the 20% rate of the total inflow into the Lake. At this point it should be noted that this rate can be considered acceptable for low lake water levels

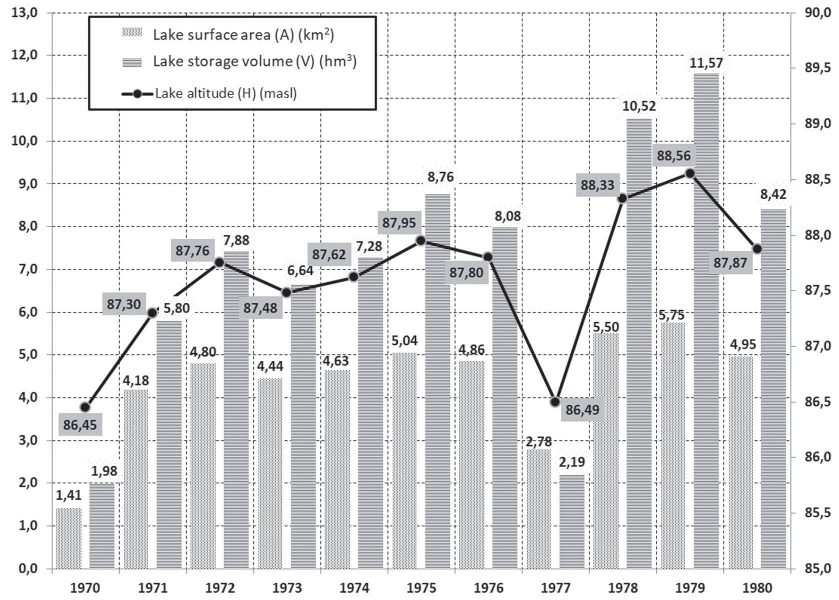

Fig. 3.25. Interannual variation trend of water storage volume (V) surface area (A) and water level altitude (H) of Bara Toibasi Lake for the period 1970–1980.

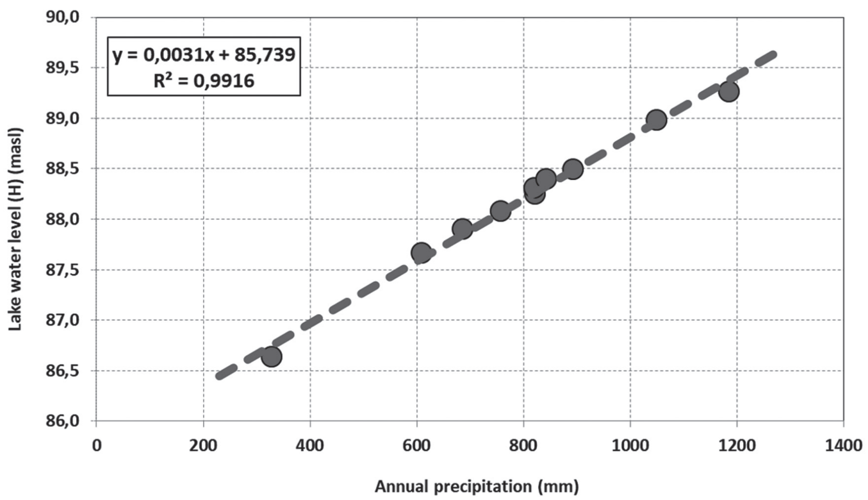

Fig. 3.26. Relationship between annual rainfall record (mm) at Spilia gauge station and predicted annual water level fluctuation (masl) of Bara Toibasi Lake.

(~86–88 masl). Corresponding with the increase of the lake level (up to 88 masl), an increase of lake water infiltration rate is expected, which, as mentioned above, can reach up to ~60%, due to the increase of the lake area e.g. higher volume and hydraulic load (stress) at sites where the karst is covered by lake water.

- Thus, the Bara Toibasi Lake water balance is dominated by runoff into the lake from the catchment area, recharge from precipitation over the lake's surface, evaporation from the lake surface and infiltration (leakage) of lake water into the Chassambali karstic aquifer.
- Finally, it is noted that the water balance established for the 1970s helps to explore the hydrologic state of the lake under natural conditions prior to the massive human intervention.

Start elevation	Start area	Start volume	End elevation	End area	End volume	Lake level changes	Lake area changes	Water balance
(masl)	(km^2)	(hm^3)	(masl)	(km^2)	(hm^3)	(m)	(km^2)	(hm^3)
86.45	1.41	1.98	87.87	4.95	8.42	1.42	3.54	6.44
Date	Date	Date	Date	Date	Date	Rate	Rate	Volume change rate
(year)	(year)	(year)	(year)	(year)	(year)	(m/a)	(km^2/a)	(hm^3/a)
1970	1970	1970	1980	1980	1980	0.13	0.32	0.59

Tab. 3.2. Values of lake level, area and volume changes matching with start and end dates obtained from the water balance between 1971 and 1980.

The water balance of Bara Toibasi Lake is summarized in Tables 3.1 and 3.2.

3.6.2.3. Summary and Conclusions on Bara Toibasi Lake water balance for the period 1971–1980

This study has provided an estimation method and a comprehensive 10-year analysis of inter-annual variations in Bara Toibasi Lake's surface area, water level and storage volume using digital terrain model data, bathymetric maps, historical meteorological records, and GIS techniques.

All of the above data were combined in order to figure out the hydrological conditions of the Bara Toibasi Lake, during the time period in which the lake served as a natural reservoir holding a large water volume with no significant human intervention in the lake's hydrology. Therefore, the established water balance of Bara Toibasi Lake for the period 1971–1980 may contribute to exploring its hydrological status (regime) under natural conditions.

The following interesting conclusions emerge from the evaluation of the compiled results of the approximate hydrological balance of the lake for the period 1971–1980 and taking into account the conditions, limitations and reservations raised regarding the reliability of the data used which in general we consider to satisfactorily approach reality:

- The lake's morphology determines the relationship between lake water elevation (H), lake surface area (A) and storage volume (V). For Bara Toibasi Lake the results show that the power and quadratic polynomial functions are most suitable to describe their H–A, H–V and A–V relationships, respectively (Figures 3.22 and 3.23).
- The annual precipitation ranged from 329.9 mm to 1185.9 mm with an average 800.4 mm, with a fluctuating but generally increasing trend (Figures 3.24 and 3.25).
- Precipitation changes have a relatively large impact and strong relationship on the lake's water level (H) as shown in Figure 3.26.
- The average yearly discharge of the Xerias stream is about 0.27 m^3/s representing 68% of the total annual inflow into the lake.
- The significant drops in the Xerias runoff in the year 1977 generally correspond to low precipitation and high evaporation (Table 3.1).
- There was no discharge to any existing surface water bodies or courses as a result of extraction activities.
- During 1970–1980 the area (A) of the lake increased by 3.54 km^2 and the variation rate was 0.32 km^2/a. H and V changes showed increased values of 1.42 m and 6.44 hm^3 with variation rates of 0.13 m/a and 0.59 hm^3/a respectively (Table 3.2).
- From 1971 to 1980, the lake water volume obviously increased from 1.98 hm^3 to 8.42 hm^3 and described an annual cycle with alternating filling and emptying phases (Table 3.2).
- In the 1970s lake water levels fluctuated about 2.07 m between dry and wet years (Table 3.1).
- Finally, the data from Table 3.1 show that for hydrological conditions similar to those of the 1970s the upper level of the lake does not exceed 90.00 masl. Even with the hypothetical rainfall values of 2000 mm, 2500 mm and 3000 mm the estimated upper level of the lake does not exceed 90.45 masl, 92.25 masl and 93.68 masl, respectively.

- Given that, as mentioned above (Chapter 3.2), with the increase of the lake level we expect an increase of the quantities of water that penetrate into the Chassambali karstic aquifer, lake levels for rainfall values 2000 mm, 2500 mm and 3000 mm must be considered overestimated.
- Therefore, we consider that lake water levels cannot rise further than 96.00 masl due to the leakage process even for the hypothetical cases by which the average annual rainfall would be between 2000 mm, 2500 mm and 3000 mm, i.e. almost four times as much as in the 1970s.
- The surface runoff, the excessive evaporation from the lake surface and especially the leakage into the Chassambali karst aquifer limit the extent, storage volume and altitude of the lake.
- Finally, Bara Toibasi Lake presents a dramatic example of a water body that has drastically declined in size due to high surface water abstraction and diversion mainly for agricultural use.

3.7. Final conclusions

Situated at the southwestern slopes of Mt Olympos–Ossa range in northeastern Thessaly are the study areas of the Sykourio and Elateia alluvial basins. The available geological, hydrological and hydrogeological data are limited and non-standardised. However, this study makes an effort to identify and understand the main hydrological and hydrogeological characteristics of the aquifers in the area.

Borehole geological and lithological data are used to study the lithological variations and the geometry of the aquifers on the vertical scale. Water-level measurements from observations on boreholes are the principal source of information about hydrological stresses acting on aquifers and how these stresses affect groundwater recharge, storage and discharge.

The intense use of groundwater for irrigation in the basins has caused major water-level declines. Trend analysis on piezometric heads show a steady decline in groundwater level for a period up to 27 years (1979–2006), indicating that the groundwater extraction exceeds the recharge.

The available piezometric data were processed and analysed. Studies of the piezometric maps reveal the main flow paths in the hydrogeological units and possible hydraulic interaction between the studied and surrounding aquifers, whilst in simultaneously the main recharge-discharge areas are depicted. From the course of the piezometric curves in the Sykourio basin the main observation is the identification of two hydrogeological sub-basins (Kalochori and Pournari).

The Chassambali karst aquifer formed in the southern margin of the Kalochori alluvial aquifer. The aquifer is mainly discharged by the Kephalovrysso spring emerging in the contact zone of the karst foothills with the sediments of the eastern Thessalian alluvial basin. The basic characteristics of the karst aquifer and its flow mechanisms were identified by analysing the hydrographs of the borehole SR72. Based on the same analysis, the complete cessation of the Kephalovrysso spring in 1986 was confirmed.

The strong hydraulic connection between the aquifers of the Kalochori sub-basin and the Eastern Thessalian Plain, through the Chassambali karst aquifer presents a completely new insight that emerged first in the present paper from the study of groundwater levels in boreholes of the two basins.

Groundwater level curves of piezometric maps in the Pournari and Elateia alluvial aquifers clearly indicate an aquifer discharge axis from SE to NW, due to the strong hydraulic communication between the two aquifers. This axis follows the Kalamitsa stream course discharging towards the Pinios River.

The course of the Pinios River presents the NW and SE hydraulic limit for the Elateia and Gonnoi aquifers, respectively. Across this hydraulic limit is recognized a groundwater flow diversion (almost 90 degrees) of the two aquifers towards NE, feeding most likely the Lower Olympos–Ossa karst aquifer.

Ephemeral streams flow out of Mt Ossa usually only in wet months of the year and discharge onto the plain of Sykourio. Large water amounts of the major stream Xerias recharge the permeable sediments of the Kalochori aquifer and dissipate by evaporation. Thus, only a minor water amount of the Xerias reaches and recharges Bara Toibasi Lake, which is a closed-catchment lake with no surface water outflow and occupies the lowest southern part of the Kalochori sub-basin.

Today the Bara Toibasi Lake has completely dried out and is covered only in its deepest parts in springtime by melt water. Considering the historical extent of Bara Toibasi Lake, it became clear that it did not exceed a level of 87.0–89.0 masl; karstification of limestones (tectonic fractures, cracks, small fissures, pipes and sinkholes) as present at the southern edges of the lake, are not favourable for fully impermeable conditions. Through those karst patterns lake water infiltrates and feeds the karstic aquifer. This leakage process presents another main factor affecting lake water fluctuations and making them unpredictable. The water level of the lake is also perceptively sensitive to climatic fluctuations.

Early human activity (since the Neolithic Age) developed around the Bara Toibasi Lake, especially due to the region's high levels of wealth in water resources and fertile ground, making it a pivotal crossroad for both human interaction and settlement. As early as the late 1960s archeological investigations revealed traces of Neolithic settlements in the basin of Sykourio.[14] These investigations have been enlarged in the last four years with new information added (compare Ch. 2, Reingruber and Toufexis). Taking into consideration the spactial distribution and the chronology of the investigated archaeological sites, the extent of the lake should not be considered as stable and permanent throughout prehistory.

Topographical maps from different decades of the 20th century indicate that Bara Toibasi Lake has undergone considerable fluctuations. It seems that from the beginning of the 20th century until 1980 it was always a shallow lake with water altitude ranges between 87–90 masl. In this time Bara Toibasi Lake served as a natural reservoir holding a large water volume. Therefore, its size and depth varied throughout the ages and has not always covered the same surface within the basin.

The approximate hydrological balance of the lake established for the period 1971–1980 reveals that under the influence of similar hydrological conditions to those of the 1970s the upper level of the lake does not exceed 90.00 masl. Even with the hypothetical rainfall values of 2000 mm, 2500 mm and 3000 mm the estimated upper level of the lake does not exceed 90.45 masl, 92.25 masl and 93.68 masl, respectively. Therefore, we consider that due to the leakage process the water level of the lake cannot rise above 96.00 masl under the climatic conditions of the late Holocene.

[14] Theocharis 1962; Gallis 1992.

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