THE LOST DANUBE: KARST HYDROGEOLOGY OF THE WESTERN SWABIAN ALB, SOUTH GERMANY

HEINZ HÖTZL
Department of Applied Geology
Karlsruhe University
Kaiserstrasse 12, POB 69 80
D-7500 Karlsruhe
Germany

ABSTRACT

The Swabian Alb formed by an Upper Jurassic carbonate sequence is the most extensive karst area of Germany. The western part of it is crossed by the upper Danube River. The river cuts deeply through the limestone bedrock and receives drainage from portions of the Alb plain and ground-water discharge. The Danube River represents an old, mainly Pliocene, drainage system. It is now restricted by the young Rhine system, which is controlled by the low base level of the upper Rhine graben. This causes a strong headward erosion. Since the Upper Pliocene, the Danube has lost more than 90% of its headwaters. The underground Danube-Aach karst system of the Western Alb represents the last capture of the Rhine, leading periodically to a complete loss of water in the upper Danube.

INTRODUCTION

Germany has several karst areas varying in size and importance, which are mainly in the middle and southern part in the outcrops of Paleozoic, Mesozoic, and Tertiary limestone and gypsum. The largest closed karst area is the Swabian Alb between the Neckar and the upper Danube rivers (Fig. 1). It is a part of the South Germany cuesta landscape and is formed by the outcrop of the Upper Jurassic carbonate sequence. This flat mountain ridge is east of the Black Forest, forms the northern margin of the molasse basin, and, to the southwest it passes into the Swiss Jura Mountains. The Alb extends from SW to NE more than 200 km, and the width is about 40 km. The cuesta faces northwest, and its steep escarpment, the so-called "Albtrauf," rises from 600 m in the foreland to 1,000 m a.s.l., from where the Alb plain dips gently to the southeast.

This carbonate, flat-topped ridge exhibits moderate to intensive karstification. The most pronounced morphological feature is the widely spaced network of dry valleys, which are in sharp contrast to the dense drainage network in the clayey foreland of the Alb. Numerous karst depressions, dolines, cave systems, and sinter formations underline the karst. Active sinkholes and karst springs demonstrate the hydrogeologic efficiency of the underground drainage system discharging towards the Danube River.

The young, dynamic morphology is determined by the relative position of the Danube and Rhine River drainage systems. Favored by the deep base level in the upper Rhine graben (Fig. 1), the Rhine River gradually captures more of the old

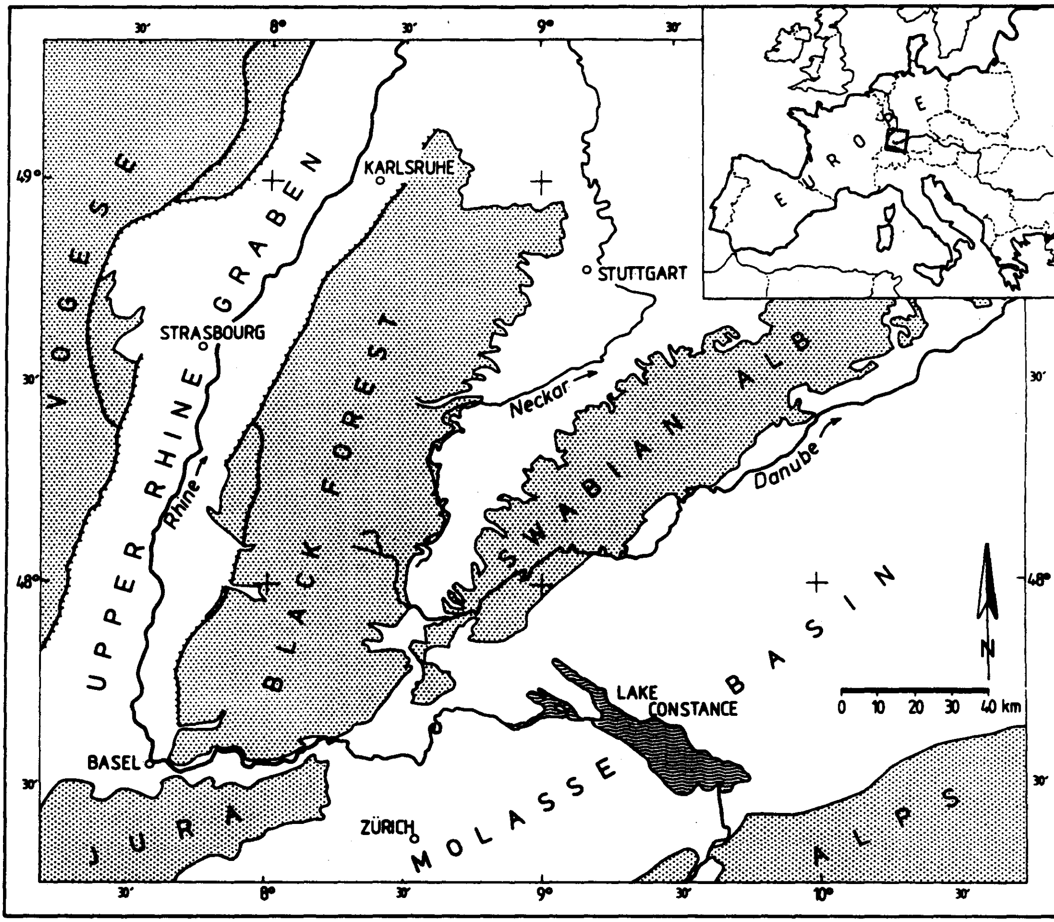

Figure 1. Location of the Swabian Alb in southwest Germany.

Danube drainage area. This affects not only the surface drainage but also the ground water and has led to one of the most spectacular karst phenomena of Germany, the "Donauversickerung", the loss of the Danube water in the western part of the Swabian Alb to the Rhine system. The hydrogeologic details of this phenomena and its development are described in this chapter.

GEOLOGICAL SETTING

The paleogeographical position of the Swabian Alb is in the southern part of the Mesozoic epicontinental basin (Geyer and Gwinner, 1984, 1986). The development of this basin began in the Late Paleozoic. South Germany was part of one of the main Variscian orogenic zones, which was folded, overthrusted, metamorphosed, and intruded by granitic plutons. During the Upper Carboniferous and Permian, the area was uplifted and eroded forming the "basement" which is now cropping out with granite and gneiss in the Black Forest.

The new epicontinental sedimentary cycle started during the Lower Triassic with a sandy continental sequence. Marine carbonates with an evaporitic intercalation succeeded during the Middle Triassic. In the Upper Triassic, continental influence prevailed. During the Jurassic, clayey, marly, and carbonaceous sediment sequences were deposited. The most dissected sequence is the more than 300-m-thick Upper Jurassic carbonates. These consist of light colored, regularly bedded limestone and marl which pass partly into massive algal-sponge limestone with tower-like structures.

At the end of the Jurassic, the area of the Swabian Alb became subaerial and vast parts have remained so. Strong weathering and karstification occurred on the flat carbonate landscape especially during the Cretaceous and Early Tertiary with the warm and humid climate. The topography of the recent landscape originated from the tectonic events of the Middle and Late Tertiary. In the south, the overthrusting of the Alps caused the foredeep basin of the molasse. With the table-like uplifting of the Black Forest in connection with development of the upper Rhine graben, the Swabian Alb tilted southeastward, toward the molasse basin. The molasse basin then controlled futher orientation of the drainage pattern.

DEVELOPMENT OF THE DANUBIAN DRAINAGE PATTERN

With the tilting of the Swabian Alb during the mid-Tertiary, a system of consequent streams was formed. These pre-Danube rivers followed the slightly dipping surface toward the molasse basin (Schreiner, 1974). After the filling of the basin, mainly due to the sediments derived from the Alps in the south, the sea retreated eastward in the Late Miocene. The drainage channels followed the retreating sea, and the early Danube was formed during the Late Miocene and Early Pliocene (Fig. 2). It comprised a huge catchment area, including that of the Aare along with vast parts of the western Swiss Alps (Wagner, 1961; Villinger, 1986). The early Danube River drained more than 20,000 km^2 in one section of the Swabian Alb which now has a drainage area of 900 km^2 (Hötzl, 1973).

During the Pliocene, tectonic displacement and additional uplifting of the northwestern part of the Alb caused a gradual movement of the Danube toward the southeast. Later, it started to dissect the Upper Jurassic limestone, and by the Middle Pliocene it had reached a base level of 200 m below the old surface. The remnants of this old meandering valley can be found as meander-scar terraces about 50 m above the recent valley floor (Wagner, 1961). Only the main tributaries within the Alb followed this deep dissolution. Because of the increasing karstification, larger areas of the new Alb high plains began draining into the subterranean discharge systems. Vertical tectonic displacements during the Pliocene (Illies, 1965) caused additional critical changes in the drainage pattern outside the Alb. In the west, the Saonne River, a tributary of the Rhone River, was captured by eastward erosion of the Aare River. Therefore, the main part of the former upper Danube catchment along with vast parts of the Swiss Alps was draining westward through the Burgundian gate directly to the Mediterranean Sea. In the Late Pliocene, the Danube catchment area was reduced but still included the Alpine Rhine and the early Eschach River with its catchment in the central Black Forest, now part of the Neckar system (Fig. 2). The final phase of development of the drainage area was the activation of the upper Rhine graben in the Late Pliocene (Illies, 1965). The downwarping of the graben basin coupled with additional uplifting of the graben edges, Black Forest and Vogese, produced a new, deeper base level with a short flow path to the North Sea. This was the birth of the new Rhine discharge system. By the Late Pliocene, the Aare River was diverted to the north into the upper Rhine graben. Due to the increased headward erosion of the Aare and the Neckar Rivers, the Rhine system was prograding to the east and capturing more of the Danube headwaters.

Three decisive steps may be mentioned. In connection with the first big glaciation, Danube glacial, about 1.5 million years B.P., the diversion of the Alpine Rhine toward the west to the upper Rhine took place (Villinger, 1986). The next important step was the deep scouring of the Hegau basin during the Mindel-Riss interglacial, approximately 450,000 to 400,000 years B.P. The Mindel gravels are above 600 m a.s.l., and the Riss moraine is at 450 m a.s.l. (Schreiner, 1974). This deep erosion also removed the impermeable molasse cover on the Jurassic limestone in the Hegau basin. Therefore, a new base level was developed for the limestone of the Western Swabian Alb leading finally to the subterreanean Danube-Aach system (Hötzl, 1973). The third step began only 20,000 years ago. At the maximum stage of the Würm glaciation, melting waters of the Feldberg glacier caused the diversion of the Feldberg Danube, now called Wutach, southward to the Rhine (Wagner, 1961). After the loss of these headwaters, the Danube retained only the relatively small headwaters of the Breg and Brigach Rivers encompassing 482 km^2 in the Black Forest. Their junction now forms the "young Danube."

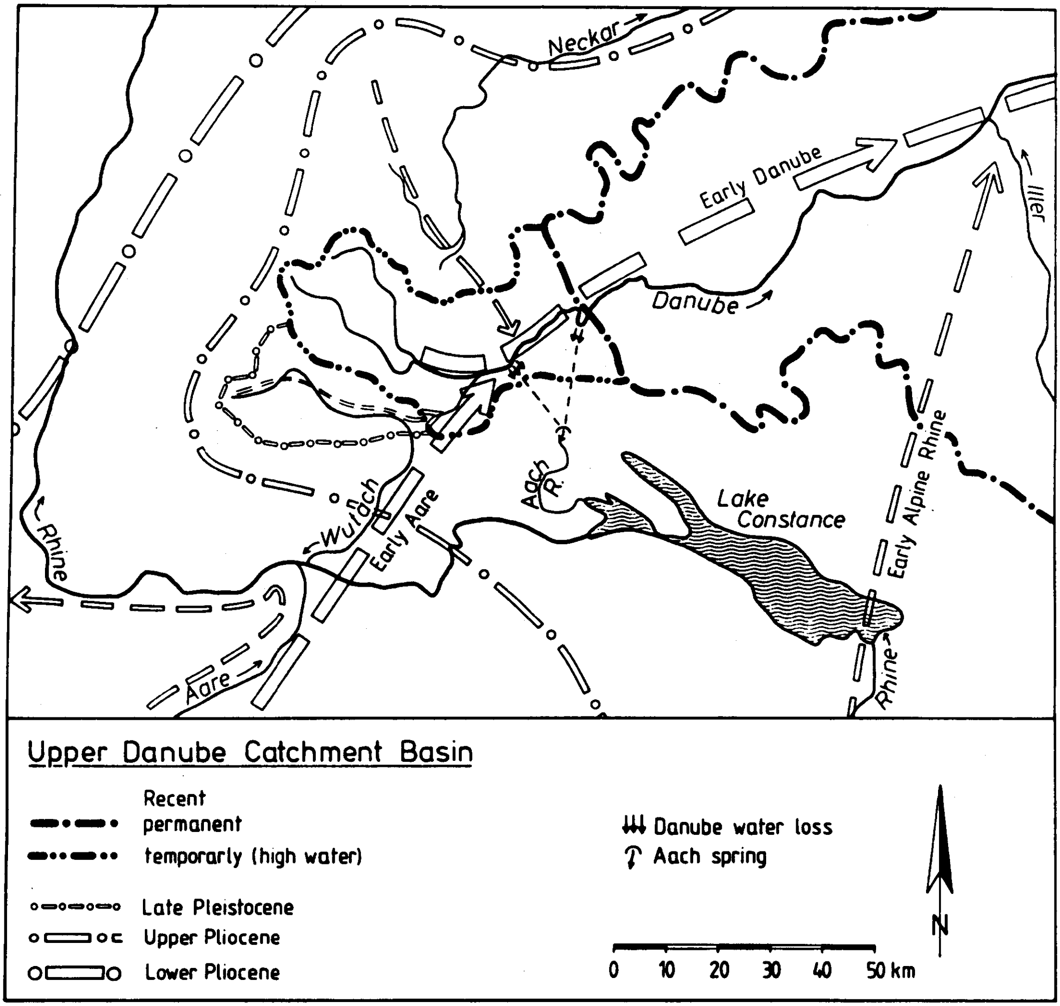

Figure 2. The constriction of the Danube catchment area by the Rhine River system.

THE UNDERGROUND DANUBE-AACH SYSTEM

Karst hydrogeologic overview

The Western Swabian Alb (Fig. 3) is obliquely crossed by the deep Danube Valley cutting the Jurassic limestone into a northern and a southern Alb high plain. For both karstified regions, the Danube originally functioned as collecting stream. From this, the Hegaualb, the high plain south of the Danube between Immendingen and Fridingen, is drained by the Rhine system of the Aach River, which is supplied mainly by the Aach Spring, the largest spring in Germany. The underground drainage area of the Aach spring extends headward up to the Danube Valley. Due to the deeper karst water level, sinking of river water occurs in several places, which periodically causes complete loss of the Danube water. The underground connection to the Aach River defines the Danube-Aach system.

The lost Danube is one of the most impressive karst features of this region. The water loss first occurs at the Immendingen weir (Figs. 3 and 4), where the stratified Oxfordian limestone submerges below the valley floor along a flexure zone and the river comes in direct contact with permeable rocks. Several sinkholes have been developed along the river bank. The main water loss occurs 1.5 km downstream along the undercut slope of the Brühl curve between Immendingen and Möhringen, where the water is sinking into the open joints of the Oxfordian limestone or disappearing through the gravel of the river bottom. During dry periods, the water level was measured at 30 m below the river bed. The loss amounts to the entire river discharge under low water conditions so that the river bed downstream from the Brühl curve is dry (Fig. 5). On average, this is the case on 150 days per year, mainly during the summer and autumn. Downstream from the Brühl curve, the Danube is renewed during the dry period by small tributaries and spring discharges coming from the northern part of the Alb plain. Further water losses of the Danube occur in the valley loop of Fridingen 40 km downstream, where the water submerges into the joints and bedding planes of the Kimmeridgean limestone (Fig. 6). The water lost from the Danube emerges in different springs of the Hegau basin (Fig. 3). The distances between the river sinkholes and the springs range from 8 to 19 km. The decisive point is the deep topographic level of the Hegau basin. The Aach Spring, the dominating karst spring with an outflow at 475 m a.s.l, lies 175 m deeper than the Danube at the Brühl curve (distance: 11.7 km).

The hydrogeologic situation can be understood within the context of the local geologic profile (Fig. 7). The base of the karstification is formed by the Oxfordian marl and the underlying marl and slate of the Middle Jurassic. The overlying carbonate sequence acts as one unique, hydraulically connected karst aquifer. Due to disruption by faults and the existence of massive limestone, the intercalated Kimmeridgean and Tithonian marl have minor influence as aquitards. The loss of Danube water occurs for the upstream reach into the Oxfordian limestone and for the downstream segment into Kimmeridgean limestone. Independently from this, the ground water emerges in the Hegau basin from the Tithonian limestone, thus confirming the hydraulic connection of the whole karst body.

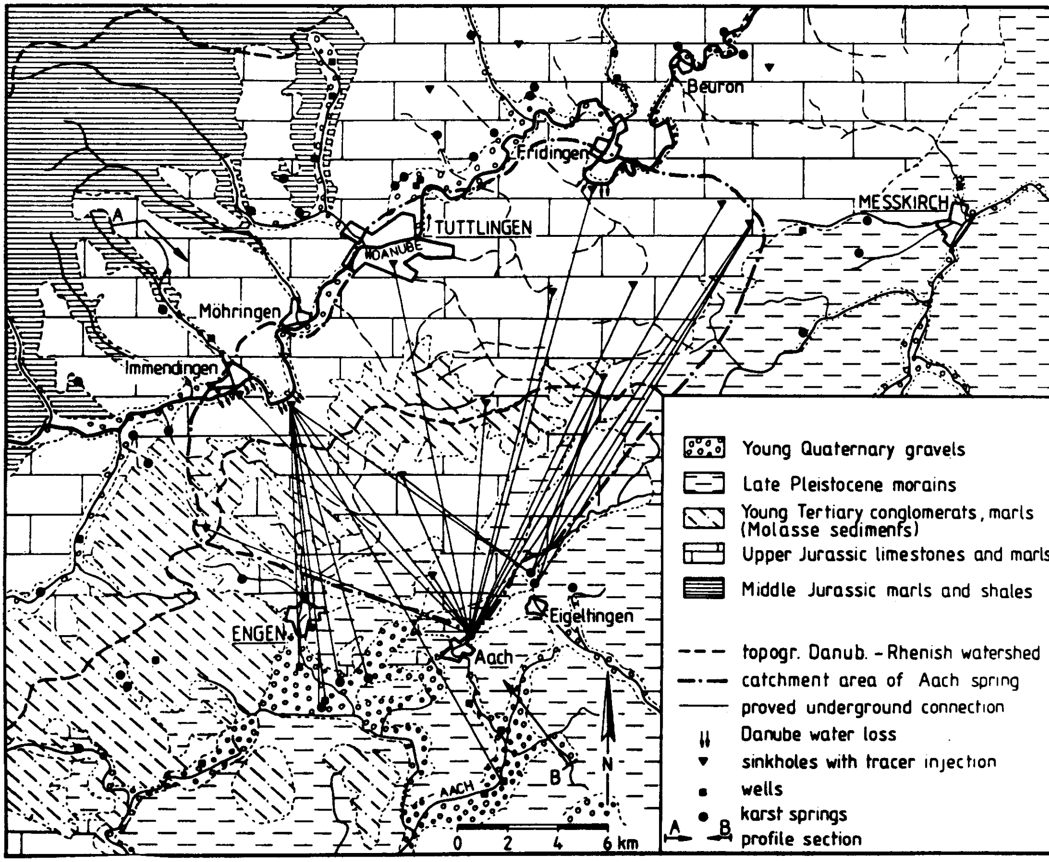

Figure 3. Hydrogeological map of the Western Swabian Alb and of the Hegau basin in the south.

Figure 4. The Danube at the Immendingen weir with sinkholes on the right bank and the well stratified Oxfordian limestone behind.

Figure 5. The Brühl curve of the Danube between Immendingen and Möhringen; on the right bank the undercut slope with the main sinking section showing the downstream Danube completely dry.

Figure 6. The Danube at the Fridingen loop with sinkholes on the left bank within the Kimmeridgean limestone.

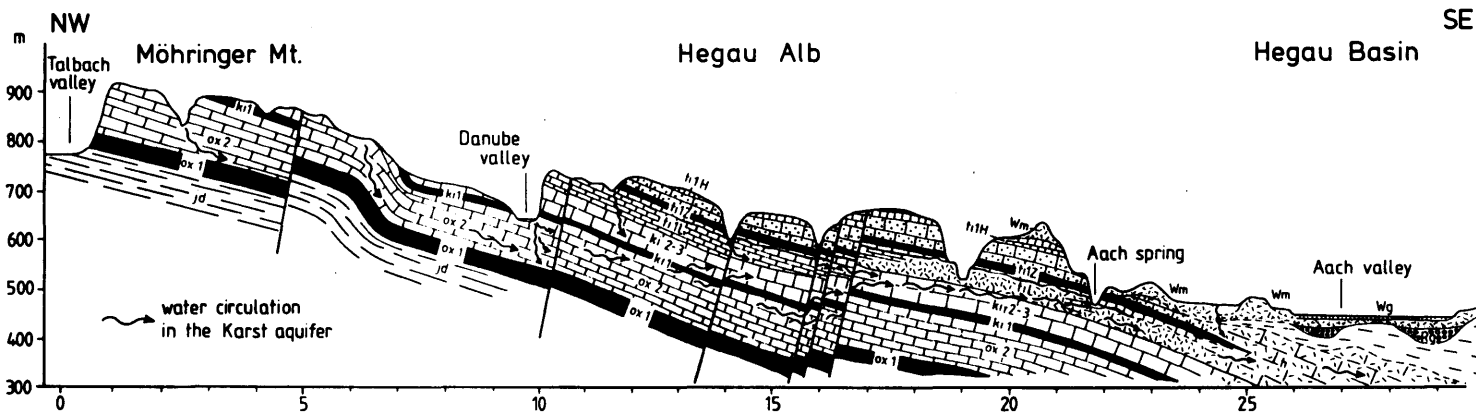

Figure 7. Geological section through the Western Swabian Alb with the Hegau basin in the southeast (according to Schreiner in Batsche et al., 1970, with supplements north of the Danube by Hötzl). Explanation of the symbols: Wg=Würm gravels, Wm=Würm moraine, ti 1H=Lower Tithonian limestones, ti 1Z=Lower Tithonian marl, ti 1L=Lower Tithonian limestone, ki 2-3=Kimmeridgean limestone, ki 1=Kimmeridgean marl, ox 2=Oxfordian limestone, ox 1=Oxfordian marl, jd=Middle Jurassic marl and shale.

Outside the catchment area of the Danube-Aach system, the Danube is still the collecting river of the discharges from the remaining parts of the Western Alb (Hötzl in Batsche et al., 1970). According to the position of the base level below or above the karst base, both types of the classical German karst subdivision into a flat and deep karst are realized there. For example, there are important karst springs east of Tuttlingen just on the northern slope of the Danube valley, where the karst base at the Oxfordian marl crops out a few meters above the valley floor. On the other side east of Fridingen, where the karst base is below the valley floor, karst springs occur on both margins of the valley.

Hydrographic Conditions

Due to the karstification, the river pattern of the Western Alb is poorly developed and restricted on rivers coming from outside of the karst area. This is valid for the main river, the Danube, as well as for tributaries coming from the northern region, while hardly any significant tributary exists from the orographic right side. Therefore, the discharge of the upper Danube River is mainly influenced from outside. Its water regime follows the hydrologic events of the Black Forest with its prevailing winter precipitation leading to high discharge in connection with the snowmelt in April. In the Alb itself the summer rains are stronger, but because of high evaporation most of the small rivers have their maximum discharge in connection with the snowmelt in spring time. The precipitation rate varies from 1,000 mm/a at the Albtrauf in the northwest to approximately 750 mm/a in the southeast.

The Danube at the entrance to the karst area is draining a catchment area of 766 km^2 resulting in a mean yearly discharge of 12.5 m^3/s, while high water may comprise nearly 300 m^3/s. Downstream to Beuron, where the Danube again becomes the main receiving stream, the catchment area increases up to 1,320 km^2 with an average discharge of 10.7 m^3/s. Villinger (1977) determined a discharge rate of 12 to 15 L/s km^2 for the karstified Alb plains. The loss of the Danube water to the Rhine-Aach system comes approximately to 7 m^3/s in the yearly average. Detailed discharge measurements and comparisons of the hydrographs (Türk, 1932, Ganz in LfU, 1980) have shown that the relationship between the river and the underground system is complicated. It depends on the karst water level and the recharge rate in the own karstic catchment. Also the loss of river water is limited by different inflow capacities along the river banks, the discharge capacity of the underground system itself, and the restricted outflow capacity of the Aach spring (Hötzl and Huber, 1972).

At low water levels a flood in the Danube may cause an increasing amount of sinking water, which uses the preferential underground flow channels. By filling them, the water from the karst plain sinking through fine fissures toward the main channels is dammed. In one case a delay of ten days compared with the regular flow time of two days could be proved for the distance from a sinkhole of the Alb plain to the Aach spring with the help of tracers. The rising of the karst water table favors the runoff from the fine fissures into the preferential channels, so that the inflow rate of the Danube is reduced. But even if there is no local recharge, the throughflow capacity is largely limited by the outflow system. Comparison of hydrographs has shown that at a discharge of the Aach spring of more than 20 m^3/s the outflow becomes independent and approximately stagnant in spite of further increases in Danube flood stage (Ganz in LfU, 1980). In the case of a large amount of local recharge, this can completely stop the sinking of river water and can even lead to karst water exfiltration along the same river section.

The average discharge of the Aach spring amounts to about 8.5 m^3/s. The observed highest rate was 24.1 m^3/s, the lowest value 1.3 m^3/s. The water emerges from two parallel adjacent joints from a depth of 17 m. There the water comes through a narrow passage behind which a wide slightly ascending cave occurs, which has been explored by divers more than 600 m to the north (Hasenmayer 1972; Schettler, 1991). Beside the main spring, there are several smaller springs nearby with discharges up to 1 m^3/s, which are already included in the values given above (Hötzl and Huber, 1972). One of these smaller springs, located only 5 m away from the main joints, is of special interest. Water outflow occurs only during total discharge of more than 5 m^3/s, while it becomes a swallow when the total discharge is less than 3 m^3/s (Käss and Hötzl, 1973).

The water balance for the Danube-Aach system was calculated from available hydrologic data and runoff measurements (Hötzl and Huber, 1972). For the Aach spring, about one third of the discharge, 2.7 m^3/s, was estimated as coming from its own catchment area. The other two thirds, about 5.8 m^3/s, are from the karst area north of the Danube, as well as from the more upstream part of the Danube. The remaining 1 m^3/s which is missed at the Beuron discharge station can be assumed to feed the small karst springs near Engen and Eigeltingen as well as the gravel aquifers of the Hegau basin. Similar results were obtained with new data by Villinger (1977) and by Ganz and Villinger (in LfU, 1980) showing more clearly the variation over succeeding years due to the changing hydrologic conditions.

Tracing Experiments

In the first note in the literature on the sinking of Danube water, the belief was expressed that there could be a connection to the Aach spring in the Hegau (Bräuninger, 1719). The first documented tracer experiment was carried out with anilin red by Ten Brink, a spin industrialist from the Aach River in 1869 (Käss, 1969). The experiment failed, probably due to the high adsorptivity and low concentration of the color. The experiment was repeated by Ten Brink together with Knop from Karlsruhe in 1877. Shale oil, fluorescein (that was the first hydrological use of this now very well known groundwater tracer), and sodium salt were injected successively within fourteen days. All three tracers reappeared within two-and-a-half days in the Aach spring. The tracing with sodium salt was the first quantitative experiment. By gravimetric analysis, Knop could prove that 18,530 kg of the injected 20,000 kg reappeared in the Aach spring (Knop, 1878).

Several later tracing experiments were carried out mainly with salt tracers from different sinkholes along the Danube as well as from the high plain. They gave first impressions on the underground flow velocity and the catchment area of the Aach spring. A combined tracer experiment with simultaneous tracer injections (fluorescein dyes, salts, colored spores, and nonpathogenic germs) in eight different places was finally carried out by an international group of scientists (Batsche et al., 1970). For the first time the reappearance of the Danube water, other than at the Aach spring, could be proved in several springs and wells of the Hegau basin. The results with the overlapping fan-like dispersion of the tracers is shown in Figure 3. Therewith, it could be confirmed that not only karst springs but also the gravel aquifers in the Hegau are partly fed (approximately 1 m^3/s) by the Danube water. With these results, it is now possible to outline the direct underground catchment area of the Aach spring. It comprises about 250 km^2 in comparison with 10 km^2 of topographic watershed (Hötzl, 1973). The flow velocity from the sinkholes at the Danube to the Aach spring ranges from 160 up to 435 m/hr for the first reappearance of the tracer and from 120 to 300 m/hr for the maximum of the tracer breakthrough. The flow rates from sinkholes on the Alb plain are slower and vary between 100 to 300 m/hr for the first detection and between 50 to 200 m/hr for the maximum.

Hydrochemistry of the karst water

The characteristic chemical composition of the water is shown by the springs in the Danube valley, which have their catchment area exclusively within the karstified limestone. The relative distribution of the ions is shown in the last line of Figure 9. As expected, there is a clear dominance of calcium and bicarbonate making up about 90% of the cations and anions. The Ca^{2+}/Mg^{2+} ratio of 18 is extremely high. The alkali ions and sulfate make up less than 5 percent of the dissolved solids. The average content of the ions given as the sum of cations (=the sum of anions) is about 5.48 meq/L.

The Danube water from upstream of the Swabian Alb has a different composition due to different geology and environmental factors (Hötzl, 1973). Though the dominant ions are calcium and bicarbonate, there are elevated relative concentrations of magnesium, sodium, chloride, and sulfate. Yet, the total mineralization is less than that of the ground water (Fig. 9). At the second control station of the Danube downstream from the sinking section of the river near Fridingen, the composition shows a distinct adjustment to the karst water. This pattern can be recognized initially from the average concentration values (Fig. 9) and becomes more clear when considering the yearly variation where, for instance, the bicarbonate values rise when the upper part of the Danube becomes dry (Fig. 10). During these dry periods, the Danube water at Fridingen is supplied only from the karst area of the Swabian Alb.

The Aach spring represents a mixture of the upper Danube water and of the ground water. The mixing ratio deduced from the chemical components is approximately 2:1 in favor of the Danube water, which coincides with the value calculated from the water balance. The influence of the Danube becomes especially evident when studying the seasonal variations (Fig. 10). While the non-carbonate components of the Aach spring (Number 8 in Fig. 10) follow the upper Danube water (Number 2 in Fig. 10), the bicarbonate confirms the changing ratios due to the recharge conditions (Fig. 10) in the Aach catchment. From the bicarbonate curves it can be seen that the large amount of melt water coming from the Black Forest in

Figure 8. The Aach spring discharging from Tithonian limestones in the Hegau during low water in the autumn. The main spring is situated in the central notch, while on the orographical right side a smaller outlet along a flat horizontal crack can be seen.

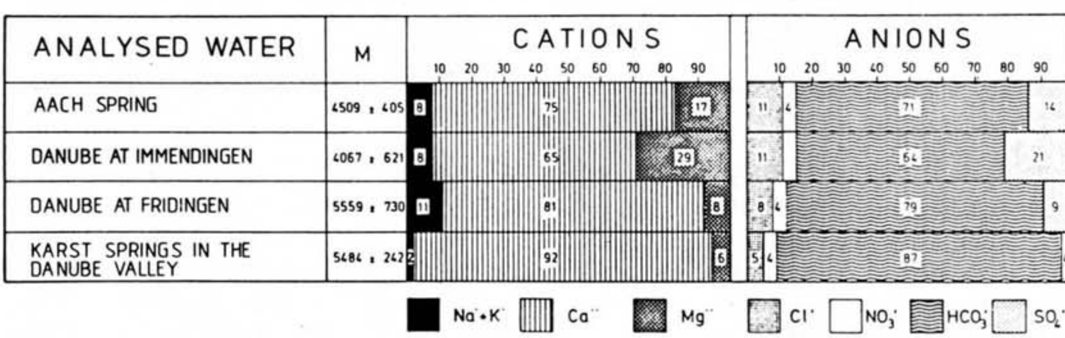

Figure 9. Chemical composition of water from the Danube-Aach area showing the relative ionic distribution. In column M the average values of the mineralization as the sum of cations are given in μeq/L.

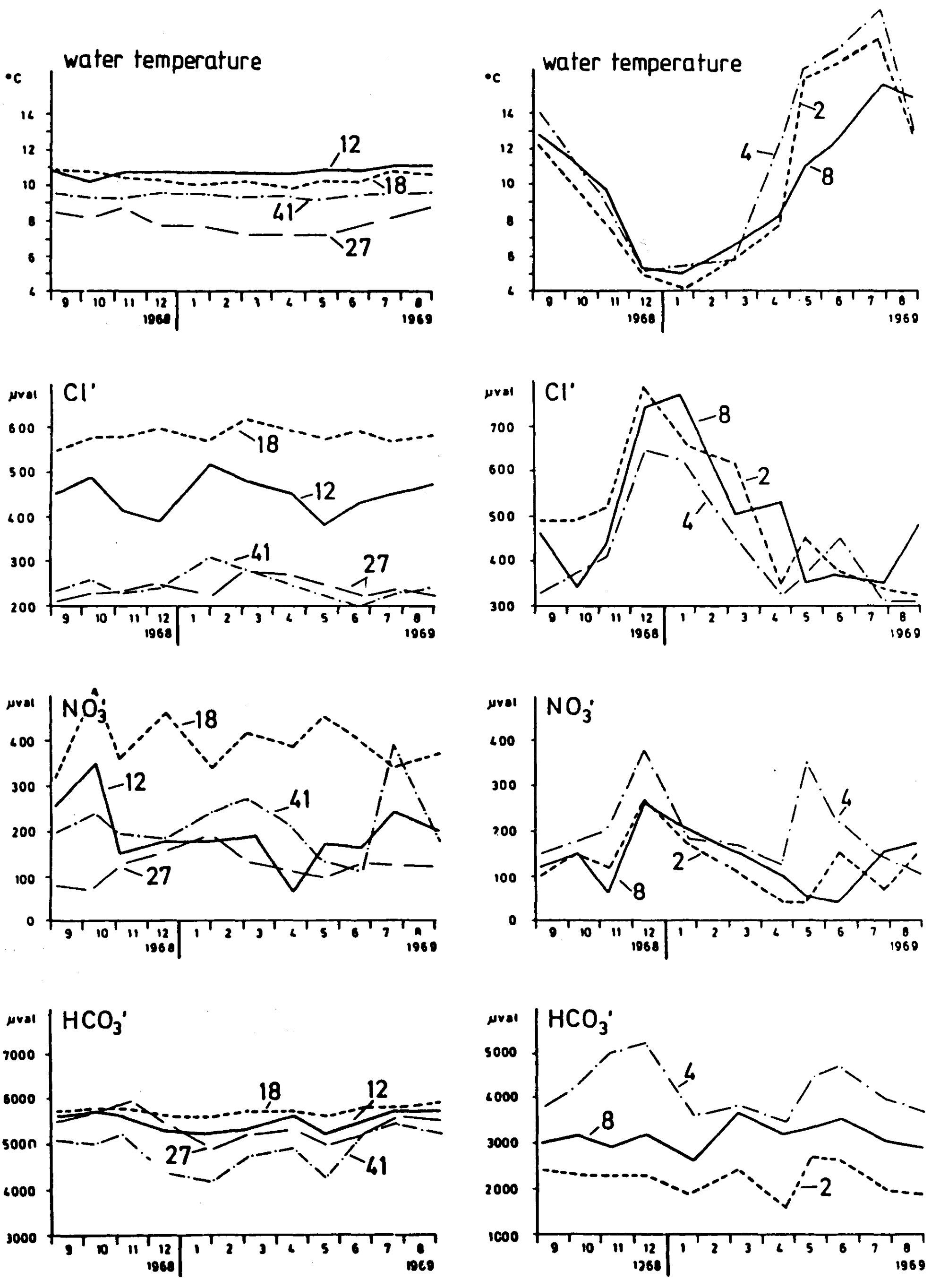

Figure 10. Seasonal variations of some chemical components of the water from the Danube-Aach area. Left column: Spring water from karst (41-Wulfbachquelle, 12-Kressenlochquelle), Tertiary conglomerate (27-Dorsteltal) and Quaternary gravel (18). Right column: Danube at Immendingen (2), Danube at Fridingen (4), and the Aach spring (8).

March through May decreases the bicarbonate content in the lower part of the Danube too (Number 4 in Fig. 10). In spite of high water in the Danube, the share of karst water in the Aach spring is increased due to the high recharge in the own karst catchment, which led to a rise of the bicarbonate content. Similar results were also obtained from tritium measurements (Geyh and Mayrhofer in Batsche et al., 1970). They confirmed greatest similarity between the Danube and Aach during August to November, when the karst water level is low, and greatest dissimilarity during the spring, when the karst water level is high. This is in contrast to the downstream part of the Danube, where greatest dissimiliarity occurs in autumn, due to the complete loss of the river in the upstream section.

From the increased concentrations of the dissolved components added during the underground flow to the Aach spring, the amount of limestone dissolved was calculated at 11.350 tons/a or 6.370 m^3/a (Käss in Schreiner, 1978). Some authors tried to interpret this in terms of the expansion of the storage capacity or the widening of the preferential cave system. In reality, most of the dissolved carbonate is coming from the soil and subsurface zone, while solutional processes along the preferential paths and previously existing cave systems are rather minor due to the fast flow there.

Evolution of the Danube-Aach System

The karstification of limestone has the disadvantage that corrosional processes do not leave direct time markers to date the time of origin. Therefore, several indirect indications and relationships with other geological and climatological processes are required to date the gradual development of a karst system. In the upper Danube, several cycles of alternating phases of sedimentation and erosion have occurred and are well dated. They enable us not only to reconstruct the development of the surface drainage pattern but also to compare it with the development of the underground discharge systems. From that we know that the Danube-Aach system is relatively young, dating back in its initial stages to the mid-Pleistocene. The whole drainage pattern of the Pliocene Danube from the Alps was oriented on the northward-dipping alluvial plain to the baseline of the Early Danube. With the downward cutting of the Danube into the Jurassic limestone plate during the mid-Pliocene, a new karstification cycle developed, but the accompanying processes were only oriented toward the Danube. The decisive change of the drainage pattern for the area south of the Danube occurred with the deep scouring of the Hegau basin during the mid-Pleistocene. The new southward-orientated drainage channels from the Hegau Alb cut through the impermeable sheet of the molasse sediments down to the Jurassic limestone. A new base level was formed nearly 200 m deeper than that of the Danube valley (Schreiner, 1974). Therewith, the limestones of the Hegau Alb could be drained southward and the newly beginning karstification of them followed that direction. This was the initial stage of the Danube-Aach system, of course at that time it was only an early Aach system, which can be dated 400,000 to 450,000 years B.P.

The thick ice sheet of the Riss glacial, reaching up the Hegau Alb to an elevation of more than 650 m a.s.l. temporarily stopped the further development of the karst. But with the melting of this ice sheet even deeper erosion channels were formed. Their bottoms are about 50 m below the recent valley floor. One of the main interglacial valleys could be traced by geophysical surveying up to about 5 km south of the recent Aach spring (Schreiner, 1978). It can be assumed that the dewatering of the karst during the last interglacial with perferential water paths was oriented toward this deep exposure of the limestone, where the Aach spring was originally situated, now being covered by gravels and moraines of the last glacial. An indication that this former karst outlet is partially still in use can be derived from the results of the tracing experiments of 1969. In a well with artesian water, the tracer from the Danube 18 km away could be proved only 8 days after injection. The short flow time can only be explained by well defined karst conduits (Batsche et al., 1970). This underground drainage system was blocked once again by the huge ice sheets of the Würm glaciation, which covered the whole Hegau basin up to an elevation of nearly 600 m a.s.l. The recession of the ice cover started soon after its maximum extension at 20,000 years B.P. The recession stages in the Hegau are very well marked by end moraines as well as by partly refilled melt water channels in front of them. One of these channels, which is dated 18,000 to 16,000 years B.P., was cut into the limestone so deeply that one of the former preferential flow paths leading to the pre-Würm main ground-water outlet 5 km south was opened. Therewith, a new main ground-water outlet, the Aach spring, was formed for the drainage system which has been developed since the mid-Pleistocene.

The time when the headward progressing underground erosion reached the Danube valley located 11 km away could not have been dated exactly until now. The first mention of the complete loss of Danube water in historic documents made only three hundred years ago and the first observation of the complete water loss of the Danube occurring in the middle of the last century are indications of a very recent inclusion of the Danube valley into the underground southward-directed drainage system. This gives certain verification to the interpretation that the underground connection of the Danube with the Rhine-Aach system is of Holocene age (Hötzl, 1973). Other authors deduce from the well developed underground karst system or from the big relief difference that the inclusion of the Danube is older than the Würm glaciation (Villinger, 1977; Schreiner, 1974).

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GEOMORPHOGENY AND KARSTIFICATION OF THE EASTERN PART OF THE NORTHERN LIMESTONE ALPS IN CENTRAL EUROPE

JOSEF G. ZÖTL
Institute of Geothermics and Hydrogeology
Research Centre Joanneum
Elisabethstraße 16
A-8010 Graz
Austria

INTRODUCTION

The various karst phenomena of individual massifs of the Eastern Limestone Alps have been the subject of investigations and publications. What we are still missing is a demonstration of the common source and roots of karstification and its geological history. Figure 1 shows the Eastern Alps and adjacent areas in Central Europe.

Plate and nappe tectonics form the initial basis for later karst development. Orogeny is an early direct influence upon the beginning of common development of geomorphologic features and karstification. A given stage of karstification is the result of interaction of endogenetic and exogenetic forces in the process. Within this complicated interplay, karstification could become a decisive factor in changing the landform, that is, when blind valleys or poljes changed to transverse valleys, or canyons developed when caves collapsed, or waterways were created by headward erosion of underground karst rivers.

Cave horizons are a further expression of geomorphogeny and karstification (paleokarst) in the Northern Limestone Alps. The eastern part of the Enns Valley is used here as an example of this sort of development, and to illustrate the development of the drainage pattern and karstification of the region in general. The changing climatic periods were also influential exogenetic forces, but acted to different degrees in karstic and nonkarstic regions. Pleistocene glaciations and glacial erosion caused hanging valleys that are still to some extent a local base level of erosion and of karst hydrogeology.

Plate and Nappe Tectonics

Because of the lack of basic information in many descriptions of karst regions it is difficult to understand karst phenomena in regions with complicated tectonics. To understand the geologic setting of the Northeastern Limestone Alps requires the reconstruction of its development and knowledge of the topography. This is especially true for regions where lithological differences play a decisive hydrogeological role in developing the geomorphology and drainage pattern.

Tollmann (1986), a highly regarded expert on Alpine tectonics, claims that the Alps are the most complicated orogenic system in the world, created by the collision of the African and Eurasian plates. With regard to nappe tectonics, we generally

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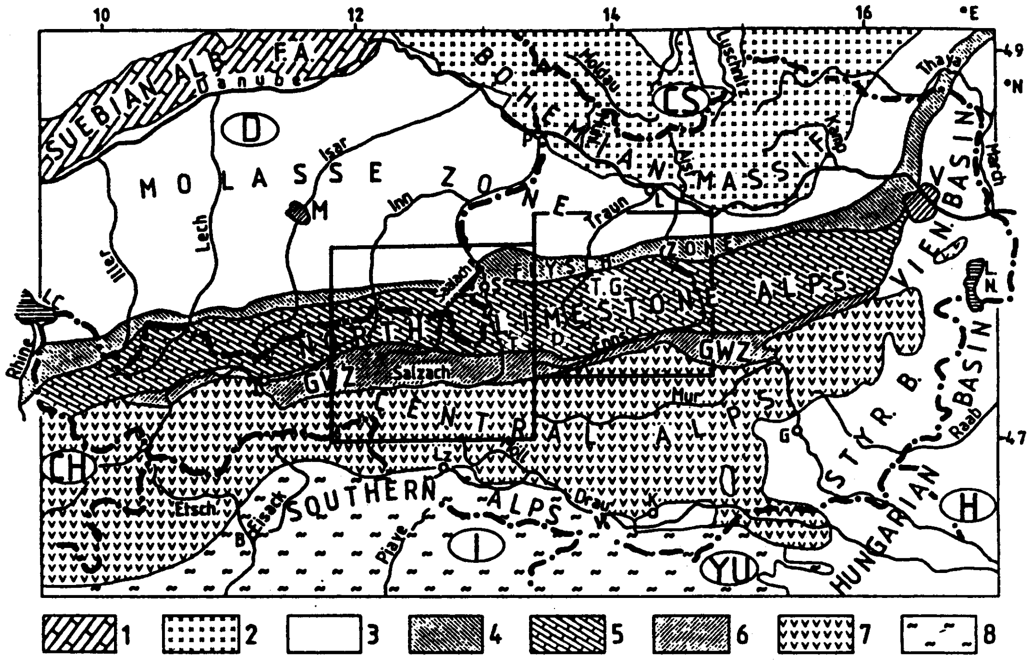

Figure 1. Generalized geologic map of Central Europe (Regions from north to south). 1 - Jurassic limestone (Malm), Germany. 2 - gneiss and granites of the Bohemian massif, Germany, CSSR, Austria. 3 - Tertiary and Quaternary sediments of the Molasse basin, Germany, Austria, CSSR; Tertiary basins of Styria and Vienna; inneralpine basins. 4 - Flysch zone; Germany, Austria, CSSR. 5 - Triassic and Jurassic carbonate rocks of the Northern Limestone Alps; Germany, Austria. 6 - graywacke zone (GWZ); Austria. 7 - Central Alps; crystalline schists, mainly gneiss with intercalary beds of marble; Switzerland, Italy, Austria, Yugoslavia. 8 - carbonate rocks and schists of the Southern Alps; Italy, Austria, Yusoglavia. B - Bozen, D - Dachstein massif, G - Graz, I - Innsbruck, K - Klagenfurt, L - Linz, Lz - Lienz, L.C. - Lake Constance, L. N. - Lake Neusiedl, P - Passau, S - Salzburg, T - Tennengebirge, T. G. - Totes Gebirge, V - Vienna, Squares - areas of investigation.

presume that the Northern Limestone Alps and the graywacke zone lie allochthonously upon a penninic underground as a nappe system overthrusted from an area south of the present Tauern mountains (Central Alps, Figs. 2 and 3). Table 4 in Tollmann (1986, p. 90-91) shows the geosynclinal and orogenic alpidic phases of the eastern Alps. In the area of the present Northern Limestone Alps, the most important phases are the creation and transport of the large east-alpine nappe systems during the Hauterivian stage (Lower Cretaceous, 125 Ma ago). This involved transposition of the East Alpine nappe systems mainly toward the north with folding and faulting of limestone nappes, and the main overthrusts in the Limestone Alps and Pre-Alps in the Turonian stage (Upper Cretaceous, 90 Ma ago).

The early orogenetic activities of the east alpine nappe tectonics, beginning for the Limestone Alps as early as 135 Ma, are calculated from radiometric age measurements. These results provide important information on metamorphism in the Central Alps and the beginning of the process of subduction of parts of the crust. However, the transport toward the north must have covered the whole undisturbed alpine limestone bedding sequence before it divided into individual nappe series. The interior nappe tectonics of the Upper East Alpine Limestone systems occurred much later, mainly during the Upper Eocene. The results of these sketchily described tectonic events are illustrated in Figures 3 and 4.

GEOMORPHOLOGY AND KARSTIFICATION

The "Augenstein" Landscape

The nappe tectonics occurred mainly as submarine events. Later an isostatic inequilibrium caused the beginning uplift of the region, and, in the Early Oligocene, the area of the present Central Alps, for the first time, became a coherent dry region. The uplift continued in phases of rise and repose during the Oligocene and resulted in a central mountain region of crystalline rocks with the limestone nappes and the flysch zone at its northern flank. Erosive actions during the repose periods created piedmont benchlands (Spreitzer, 1962). The different altitudes of old denudation planes are therefore not primarily the result of local tectonics as commonly believed.

During the Oligocene and the early Miocene, the longitudinal valleys of the present Inn, Salzach, and Enns rivers did not exist, and therefore a S-N drainage from the Central Alps crossed hilly limestone mountains and flowed into the "Molasse" sea (Fig. 3). Because the central zone rose faster, erosional activities increased dramatically and a thick cover of crystalline gravel and sand accumulated on the limestone surface.

Origin of the term "Augenstein" landscape (English "eye-stone" landscape) is as follows. After denudation of this gravel cover and separation of the various limestone massifs, quartz gravel and remnants of quartz-sand layers were found on the plateaus. For a long time, no proposed explanation was acceptable for this phenomenon. Hunters and shepherds noticed the difference between the limestone and the round, polished, quartz pebbles and called them "eye stones." It was Simony (1895) who discovered the connection with the paleodrainage patterns.

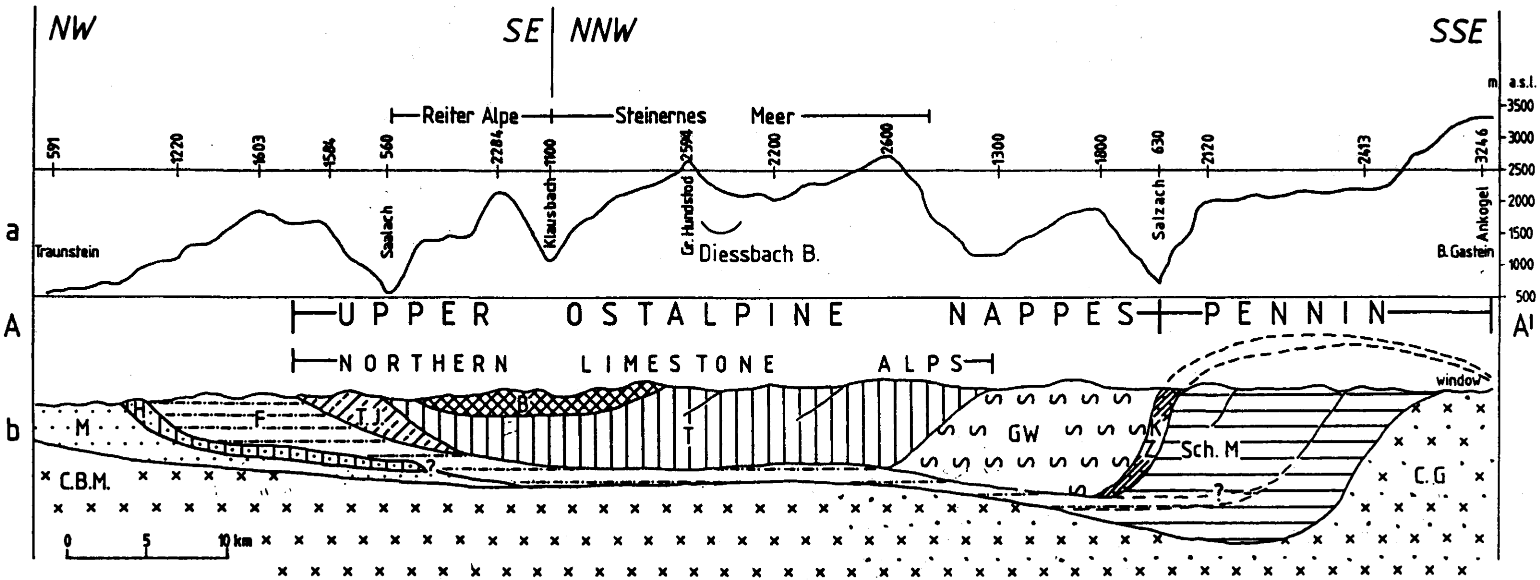

Figure 2. Tectonic and orographic profile A - A', see map Fig. 4. First tectonic and petrographic preconditions for the development of the longitudinal valley of Salzach River in easier erodible schists and the development of the karst massifs on the Northern Limestone Alps in more resistant carbonate rocks during the orogeny (b). Profile "a" shows the present orographic features. CBM - crystalline basement, i. e., underground continuation of the Bohemian massif. CG - alpine central gneiss. Sch. M. - Penninic crystalline schist mantle of Central Alps. - Penninic "Klamm"-limestone. Upper Ostalpine nappes: GW - graywacke, T - Tyrolian nappes, B - Berchtesgarden nappe, TJ - Low Bajuvarian nappe. M - Molasse sediments, H - subducted Helvetian sediments, F - Flysch. Tectonic profile (mainly after Gwinner, 1971).

Tollmann (1986) dated the beginning (Early Oligocene) and the end (Late Oligocene/Early Miocene) of this period by finding the change from crystalline sediments to gravel from nearby areas (Helvet and Flysch zone) in the sediments in the Molasse trough.

With regard to the later karstification, two scales may be used to estimate the amount and time of denudation and early separation of the Central and Limestone Alps. These are the thickness of the sediments in the Molasse basin between the Alps and the Bohemian massif and the radiologic measurements of cooling of minerals. Tollmann (1986) calculates an uplift of 0.2 to 0.5 mm/a and erosion of the slow-moving Venediger Group in this region amounting to 11,000 m since the Upper Oligocene (27 Ma), 6,000 m since the Badenian (Miocene), and 3,000 m since the Pannonian stage. Comparable measurements in the Zillertaler Alps, 40 km west of the Hohe Tauern, show a denudation of 5,000 m since the lower Pannonian stage in the crystalline Central Alps, but only 3,000 m in the graywacke zone because that region uplifted more slowly than the crystalline central mountains.

Karstification and Development of the Present Drainage Pattern

Only estimates are available of the thickness of the Augenstein accumulation upon the limestone nappes of the later Northern Limestone Alps. The highest altitude at which remnants have been found in fissures near the top of the Gjaidstein, one of the highest peaks of the Dachstein massif, is about 2,800 m. The largest concentrations of Augenstein remnants now lie in karst hollows, caves, and cave ruins in the plateaus between 1,700 and 2,000 m (Fig. 3) and are secondary or even tertiary deposits. After denudation of the Augenstein sheet from the carbonate rocks, the erosion in the Central Alps decreased with the loss of water on the karstifying plateau surfaces. This, in turn, promoted the development of longitudinal valleys along the limestone border in the less resistant schists of the lower Trias and in the graywacke.

The limestone massifs became more strongly uplifted but were also denuded by surface erosion. Under present climatic conditions, an average of 15 mm/1,000 a surface erosion is measured on bare alpine plateau karst (Zötl, 1979; Bögli, 1978). Since the Middle Miocene (20 Ma), we estimate a minimum of 300 m and perhaps as much as 500 m of surface erosion without considering the extreme variations in climate. Such climatic variations lose significance with increasing altitude, as in the present situation in central Africa where, for example, Kilimanjaro has a permanent cape of snow and ice. Weathering processes in lower regions are more sensitive to changes, especially changes in the timberline. Bögli (1978) estimates the amount of surface corrosion in wooded alpine areas at more than 80 mm/1,000 a, that is, 800 m in 10 Ma, and I think that even Terzaghi (1913) saw these changes as the reason for the lower average altitude of the Limestone Pre-Alps.

Karstification began before the denudation of the entire Augenstein cover; the present remnants of these sediments in recent uvalas, dolines, cave ruins, and the caves of the massif bear witness to this. Cave horizons at different altitudes show phases of the development of the individual massifs of the Northern Limestone Alps and their underground drainage. The first stage was the

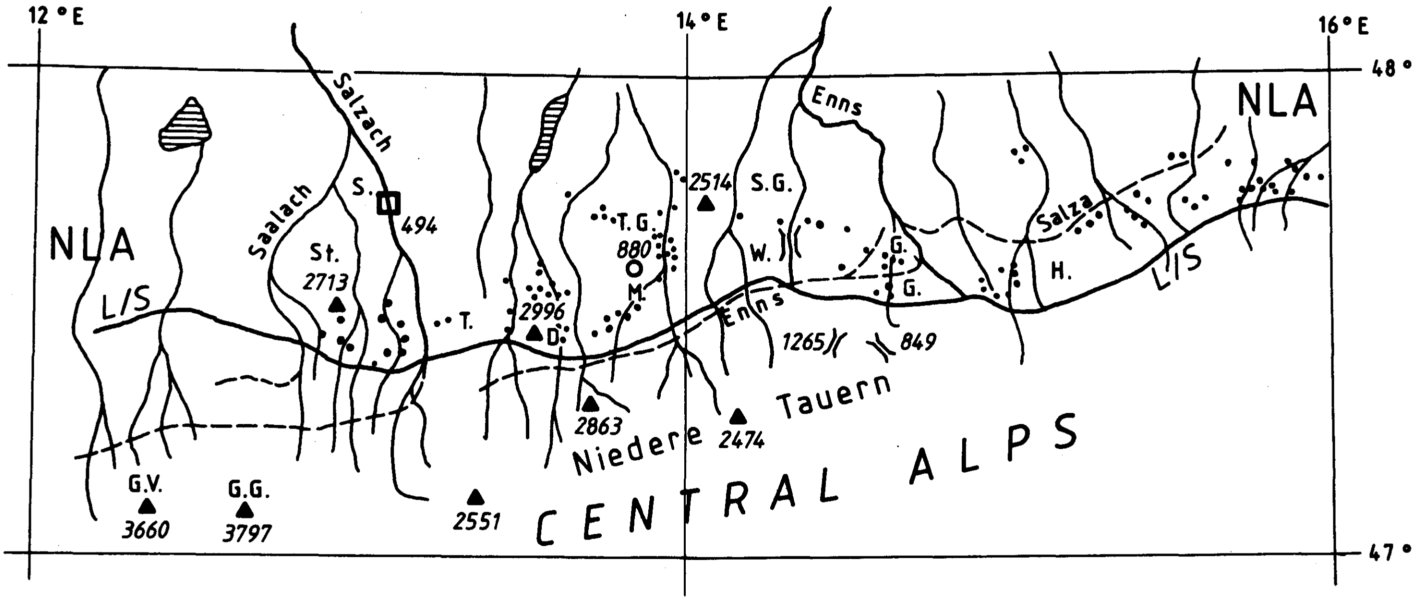

Figure 3. Consequent drainage patterns from south to north during the Augenstein landscape (Oligocene). Broken lines later developed longitudinal valleys of the rivers Saalach, Salzach, Enns and Salza. NLA - Northern Limestone Alps. L/S - boundary line between the carbonate rocks of the Northern Limestone Alps on one side and schists (Triassic "Werfener" schist and pal. graywacke) on the other. Dots - remnants of "Augenstein" sediments upon the plateaus of limestone massifs. D - Dachstein massif, G - "Gesäuse" mountains, G.G. - Groß-glockner, highest peak of the eastern Central Alps, G.V. - Groß-Venediger, H - Hochschwab massif, M - Basin of Mitterndorf, S - Salzburg, S. G. - Sengsen-Gebirge, St.-Steineres Meer massif, T - Tennengebirge massif, T. G. - Totes Gebirge massif, W - Warscheneck massif (partly from Winkler-Hermaden, 1957).

Augenstein landscape in the Late Oligocene. The surface runoff carried material from the Central Alps over the carbonate rocks into the Molasse basin in the north. The next stage shows the situation in the Miocene. During the early Miocene, the Augenstein cover had already been severely eroded; leakage of surface water into joints of the underground carbonate rocks forced the process of karstification. Large cave systems, such as the Mammut Höhle at present elevation of 1,600 m with active cave rivers, were formed by both chemical and physical erosion (Figs. 5 and 6).

The next step was a probable stage in the Pliocene with development of large cave systems at a present altitude of 1,200 to 1,400 m and the level of typical hanging valleys in limestone (Zötl, 1974, 1984). The present karst massifs of the Northern Limestone Alps show a so-called "shallow karst," large springs on the northern slope of the karst massif and small springs on the boundary of the limestone and schist in the south. All of them have seasonal variations in discharge. It should be noted that in the area from the Dachstein toward the east, the limestone massifs are higher than the Central Alps, Niedere Tauern, from which the large amount of Augenstein sediments had been eroded and accumulated upon the limestone region during the Oligocene.

Karstification and Climate

Figure 4 shows the massifs of the High Limestone Alps, the Mitterndorf basin, and the present drainage pattern. During the long period from the Miocene to the Quaternary period there were remarkable climatic fluctuations, but for long stretches more uniform phases dominated the Tertiary. Some authors have assumed that the highest peaks of the Dachstein massif are remnants of a tropical tower karst (Niederer Dachstein 2,934 m; Hoher Dachstein 2,995 m; Huner Kogel 2,694 m; Kleiner Gjaidstein 2,735 m; Hoher Gjaidstein 2,794 m; Niederer Gjaidstein 2,482 m; Mitterstein 2,417 m; Koppenkarstein 2,866 m). This may be possible, but I do not see any conclusive proof; I think it looks like a typical nunatak landscape. Gareis (1981) and Büdel and Gareis (1986) have investigated lower inner-alpine basins much more critically. They studied the fluctuations of local glacial ice tongues and the role of dead ice during temporary periods of climatic depression in the Late (Würm) Glacial in the basins of Bad Mitterndorf, Austria, Feltre, and Belluno, Italy, and the valleys from Sargens to the Walensee, Switzerland. One of the important results of these studies is that erosion was surprisingly slight in these low, wide alpine basins, which were already formed in preglacial times as intramontane plains.

The Basin of Bad Mitterndorf

As I mentioned (1964) it was Spengler (1918) who found that the present arcuate Traun valley running from the basin of Bad Mitterndorf toward the Hallstätter Lake can only be explained in development from a primal underground drainage, "Höhlenfluß," that is, cave river, caused by the influence of layers of the "Gamsfeld" nappe. The valley now dividing the Sarstein from the Zinken, northeastern Dachstein massif, was formed by collapses during the Pliocene. For information on the Pleistocene ice streams in the Bad Mitterndorf basin see Penck

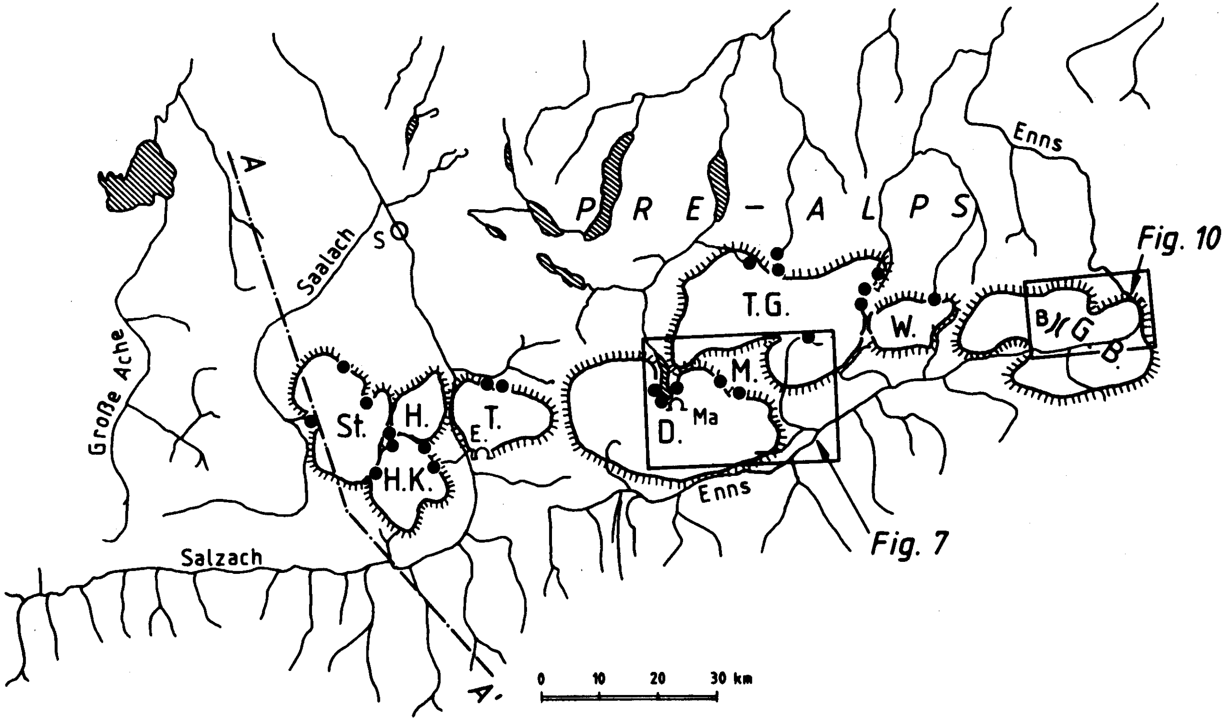

Figure 4. The present drainage pattern of the Northeastern Limestone Alps between the Groβe Ache and the Enns River and the northern slope of the Central Alps (Fig. 3). Very impressive the asymmetric river patterns of Salzach and Enns. Full black circles springs of a discharge of 100 - >1,000 L/s. A - A' profile Fig. 2. The largest limestone massif are St - Steinernes Meer, HK - Hochkönig, H - Hagengebirge, T - Tennengebirge, D - Dachsteingebirge, T.G.-Totes Gebirge, W - Warscheneck, G.B.-Gesäuseberge. B (in Fig. 10) - Buchauer Sattel, M (Fig. 7) - Mitterndorf basin, Ma - Mammut Cave (Dachstein massif, Fig. 5), E - Riesen-Eishöhle, Tennengebirge (Figs. 4 and 6).

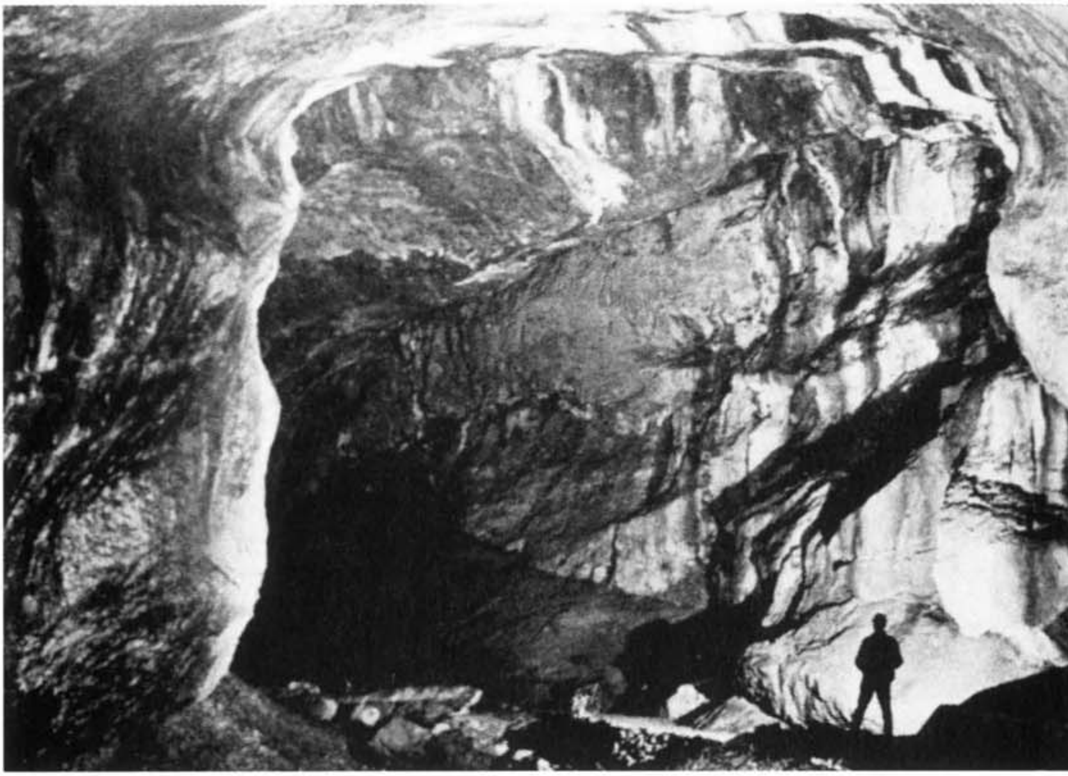

Figure 5. The most important gallery of the Dachstein Mammut Cave (ca. 1,500 m). The cave was formed by corrosion as well as erosion. Geodetic measured length of the cave is 40 km (Photograph by K. Mais, 1970).

Figure 6. View from the entrance into the "Eisriesenwelt" cave (1,650 m) towards the Salzach Valley (valley floor ca. 600 m). Geodetic measured length of the cave is 43 km (Photograph by G. Abel, 1968).

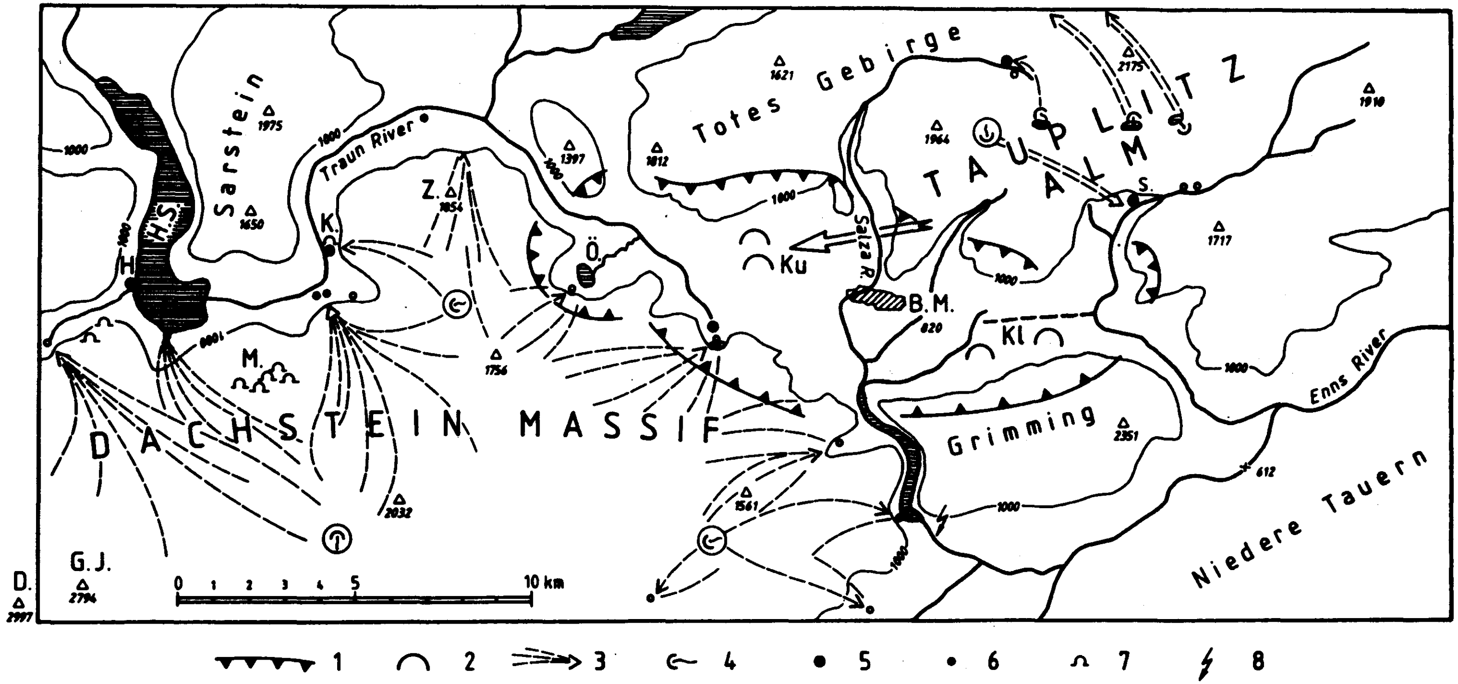

Figure 7. The basin of Mitterndorf and the surrounding limestone massifs (Fig. 4). 1 - borderline of the basin, 2 - karst kegel (cockpit), 3 - flow direction of karst water in the eastern part of the Dachstein massif and the southern part of Totes Gebirge (Tauplitz Alm), 4 - sinkhole (ponor): sinkhole with circle point of injection of tracers, small arrows output of tracers, 5 - springs with perennial flow, 6 - intermittent spring, 7 - cave, 8 - electric power plant; B.M. - Bad Mitterndorf, D - Hoher Dachstein, G.J. - Gjaidstein, H - Hallstatt, H.S. - Hallstätter See (lake), K - Koppenbrüller Höhle (cave), Ku - Kumitzberg, Kl - Kulm, M - Dachstein Mammut-Höhle (cave), Ö - Öden-See (lake), S - Sagtümpel, Z - Zinken (Fig. 8), arrow - view direction Fig. 8.

and Brückner (1901-1909), Lichtenberger (1956), van Husen (1968), Gareis, (1981), and Büdel and Gareis (1986).

Gareis (1981) and Büdel and Gareis (1986) show that the Pleistocene ice streams entered a wide intramontane plain, and "inselbergs" like the Kamp, Kumitzberg, or Kulm in the Mitterndorf basin are neither nunataks nor glacial "Rundhöcker" (roche moutonée, sheep backs), but remnants of tropical karst morphology (Fig. 8). Figure 9 shows an example from northern Malaysia comparable in form and size to the "inselbergs" in the Mitterndorf basin. The morphogeny of the gorge of the Gesäuse, the eastern part of the Enns Valley, is an example of interaction of Pleistocene glaciation and karstification.

The Enns Valley

Karstification of the Dachstein massif (Miocene) was concomitant with the development of the longitudinal valley of the Enns River. Almost all the upper branches of what had been the consequent drainage pattern from the Central Alps became tributaries of this longitudinal valley, as is demonstrated impressively by the presently asymmetrical river pattern. Typically, the longitudinal valley has a rather flat bottom with an average width of 2 km; this changes about 100 km downstream. There, the river enters a gorge, the gradient increases dramatically from 0.33% to 16%, and the river crosses the limestone massifs of the Gesäuseberge (Fig. 10) into the lower Gams basin at the eastern side of the gorge.

A cursory look at the geology provides no immediate explanation for this development. However, there is clear evidence of older S-N drainage on both sides of the gorge (Fig. 3). Detailed geological mapping (Ampferer, 1935) shows a gravel layer of the Enns River about 150 m thick and a ground moraine at the top of the Buchauer Sattel pass at 372 m elevation (Fig. 10B), that is, a present altitude 270 m higher than the entrance of the river into the gorge. Here, the river bed is the Werfener schist, which is more susceptible to erosion than the hard limestone of the gorge. After new drilling in the Enns Valley near the entrance into the gorge, I conclude that the gorge is the result of headward erosion by underground drainage of karst water from the Gesäuseberge into the tectonic basin of Gams east of the gorge of the Enns River at present elevation of 500 m. This was the nearest local base level of erosion. The gorge is the result of collapses, similar to phenomena in the classic karst, that is, the lowest part of the Reka Valley where the Reka River enters the Skozianska jama (Zötl, 1963).

The details are rather complicated. Sinkholes and seepage of river water may have started in the Upper Pleistocene. On the other hand, the Enns Valley further up the gorge has a sediment fill of 195 m thick with only layers of Würm moraines and lake sediments. The age of the erosion is not quite clear. Zailer (1910) found five horizons of postglacial lakes with marsh and peat layers. The exact bottom of the gorge, where the river flows turbulently between large blocks, is not known. What is clear, however, is the role of karstification in this development. As to the influence of climatic variations, it was the work of the glaciers during the cold stages of the Pleistocene that created the dominant features of glacial cirques over deepened valleys and tributaries as hanging valleys.

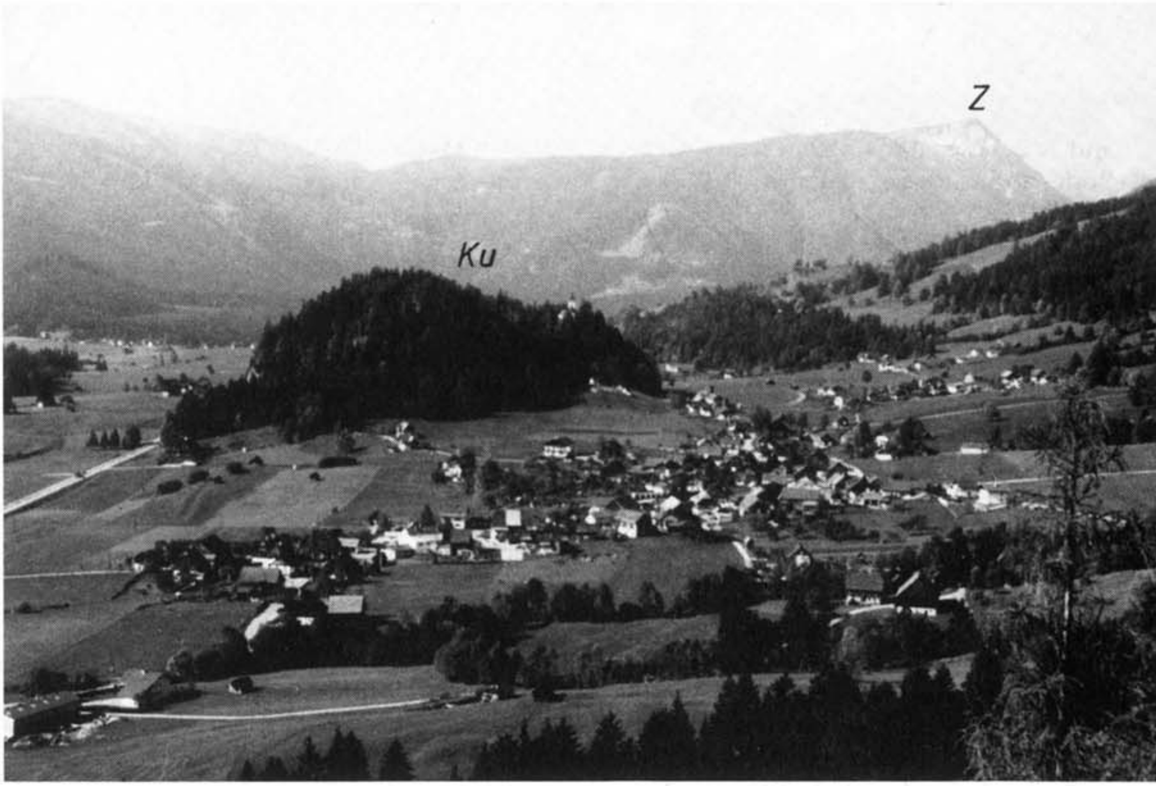

Figure 8. View into the basin of Mitterndorf. Ku - Kumitzberg (cockpit or kegelberg of the paleokarst, Fig. 7), Z - Zinken, northern peak of the eastern part of the Dachstein massif (Fig. 7) (Photograph by H. Kain, 1987).

 Figure 9. Cockpit (karst kegel) near Ipoh/Malaysia (Photograph by J. Zötl, 1981).

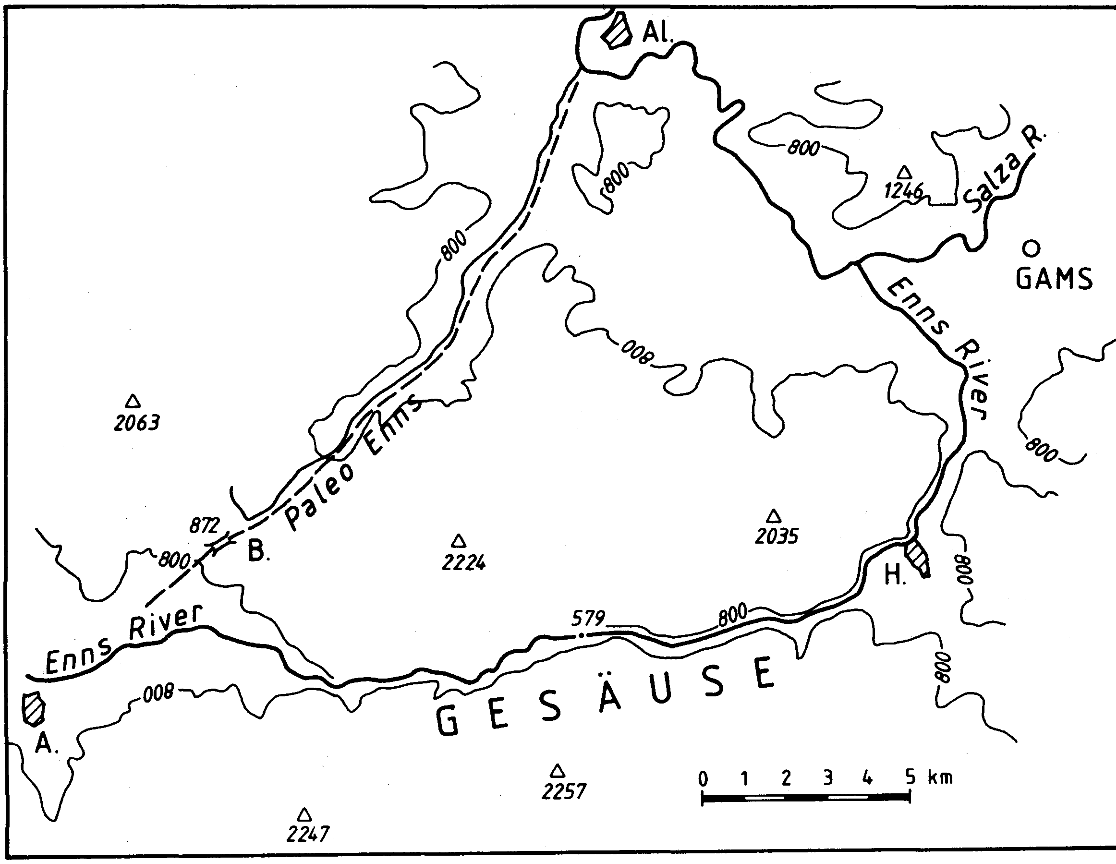

Figure 10. Paleo Enns River and the present river gorge "Gesäuse". A - Admont, Al - Altenmarkt, B. - Buchauer Sattel (pass), H - Hieflau (Fig. 4).

UNDERGROUND DRAINAGE AND WATER BALANCE

Horizons of hanging valleys, caves, and springs show a rather clear connection with old or present local base levels of erosion. Cave systems at high altitudes are to be found in most of the massifs of the Northeastern High Limestone Alps. In part, they are evidence of the karstification that has occurred since the end of the Augenstein landscape. Unlike some hanging valleys, they lost their function as a local base level of underground drainage and are dry. The karst water discharges at the present base level around the massifs from large karst springs. The discharge of these large springs fluctuates both episodically and seasonally. Intermittent springs with overflow discharge may vary in discharge from 0 to 10 m^3/s (Zötl, 1961).

Radiohydrological studies of springs from the Totes Gebirge massif provide a general hydrogeological description of the karst massifs of the High Northern Limestone Alps (Dincer et al., 1972). Replenishment of the ground water varies seasonally with rainwater and melted snow mingling in different proportions in different regions and seasons. The low water in the springs in autumn is entirely rainwater from the previous summer. In the late winter and spring a certain amount of seepage from melting snow occurs (Fig. 7). In early summer the discharge of the large karst springs may be due entirely to melted snow from the plateau region. Then, during the summer, the proportion of rainwater becomes larger.

The most important conclusion from the isotopic measurements is that no appreciable storage of water exists in the high karst massifs of the Northeastern Limestone Alps. The flow curves indicate active storage, that is, storage above the point of emergence. From the tritium and oxygen-18 content of the spring and lake water in the Totes Gebirge, it is evident that all the feed water in this area is active, that is, it lies above the altitude of the low-lying springs and lakes. This is typical shallow karst drainage. This demonstrates a special type of subterranean karst drainage where two components, a high altitude reservoir and a deep karst water reservoir, control ground-water flow which is fundamentally different from the Mediterranean type.

SUMMARY

Nappe tectonics created the basic prerequisites for the present general geological features of the Northern Limestone Alps. The local limestone massifs are the product of nappe tectonics, uplift, and selective exogenetic powers at work since the Oligocene. Since this period, a consequent S-N surface drainage pattern, Augenstein landscape, has changed to tributaries of a subsequent longitudinal stream with the denudation of the Augenstein cover and karstification of the present Northern Limestone Alps. For the surface karst, a minimum of 300 m of erosion is estimated on the plateaus of the massifs. Large caves presently located at high altitudes, now paleokarst, also show that underground drainage and karstification were active. Cave horizons illustrate stages of geomorphology and karstification in the limestone massifs.

There are relationships between karstification and climatic conditions both in stable climates and in Pleistocene glaciation. The wide Mitterndorf basin between the Dachstein massif and the Totes Gebirge still has many features of the old tropic intramontane plain with its inselbergs. On the other hand, Pleistocene ice streams with thicknesses of many hundreds of meters and strong erosional power overdeepened the main valleys, reduced the erosion of the smaller ice streams from the tributaries, and created the typical hanging valleys. In some cases, these hanging valleys are still the local base level of erosion and underground karst drainage, such as Diessbach basin (Zötl, 1984). Depending on the strata series given by nappe tectonics, the High Northern Limestone Alps are more or less individual limestone massifs with the hydrogeological character of shallow karst, that is, all the feed water of the karst massif is active and lies above the altitude of the base level of erosion.

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HYDROGEOLOGICAL CHARACTERISTICS OF GEOSYNCLINE KARST AQUIFERS WITH AN EXAMPLE OF THE TREBISNJICA CATCHMENT

PETAR MILANOVIC
Energoprojekt
POB 20
11070 New Belgrade
Yugoslavia

CHARACTERISTICS OF DEEP KARST

Basic Factors of Genesis of Karst

The term karst describes a complex geological environment with specific hydrogeological, hydrological, and geomorphological features. Karstification is a specific exodynamic process in which the structure of rocks and the continuity of large soluble rock masses (limestone, dolomite, gypsum, and halite) are changed by water. In contrast with other natural exodynamic processes, which are mainly limited to the shallow surface zones, karstification also encompasses deeper masses. Its vertical and horizontal extensions are genetically closely connected with discontinuities of endodynamic origin.

The principal conditions for karstification are soluble rock mass, permeability (joints, fissures, and faults), and large quantities of water. The karstification process starts with the chemical action of water, by corrosion. Where faster water circulation occurs, erosion plays a considerable role. The intensity of karstification depends to a considerable extent on temperature; with a decrease of temperature this process becomes more intensive. The intensity of karstification is not continuous on a geological time scale. Several phases of karstification occur. It starts with the initial phase, passes through intensive karstification, and then reaches its maximum in the phase of complete karstification. Thereafter, intensity of karstification decreases gradually to the final stage of karst forms. Neotectonic processes influence the direction and intensity of karstification. In some regions the evolutionary cycles repeat themselves, while in other regions they may be interrupted at any developmental stage.

Basis for Determining Types of Karst

Among the numerous karst classifications, those based on the regional geological, hydrogeological, and geomorphological characteristics usually prevail. Regional geological factors determine the dynamics and complexity of karst processes in broad geotectonic units, so that the difference between the karst geosyncline and epicontinental areas is clearly visible (Herak, 1977). Local structural characteristics and neotectonic activity are of decisive importance for the origin of hydrogeological characteristics of smaller units. The evolution of karst aquifers in geosyncline areas is incomparably more dynamic then in other areas. It takes place under conditions of complex structure which makes the classification of karst environment considerably more difficult. Zones of tension and compression alternate under such geotectonic conditions. The compression zones, especially

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where connected to the lithological members of hydrogeologically impervious strata, represent the rocks that are subject to limited karstification; in the part of structures subjected to tension, karstification develops at full intensity.

Knowledge of the type of karst should serve as a basis for solving practical engineering problems. Such problems include determining ground- and surface-water budgets, construction of surface reservoirs, and the use of karst aquifers. For both the geosyncline and the epicontinental karst, the division can include shallow and deep karst, open and closed, and continental and littoral. Their identification is based on hydrogeological and geomorphological differences and specificities. However, the regional geological and tectogenetic aspects are, in essence, the basis of all classifications.

All further considerations in this chapter refer to the karst of the Dinarides, identified as a prototype for mature karst in which the depth of carbonate rocks is several kilometers, or the base of permeable rocks is far below the level of all regional erosion levels and the water circulates through deep karst channels. This discussion also covers the karst of the Helenides and the Taurides which represents parts of the same orogenic zone as the Dinarides.

Porosity and Depth of Karstification

The primary porosity is negligible in the carbonate rocks of the Dinarides, whereas the secondary porosity of tectonic origin makes up a large part of the total porosity. The brittle carbonate rocks under geosynclinal conditions are nearly continually subjected to tectonic forces; therefore, a dense network of cracks, joints, and faults is created. Vertical joints and bedding plains are particularly favorable for rapid development of karstification. The zones of dissolution are formed at cross joints and at contacts of limestone with less permeable rocks. At the surface these places are sinkholes, dolines, or poljes; within the rock mass, these zones are conduits or channels. The karstification intensity and direction of development depend on the position of erosion base level. The erosion base contributes to the direction in which the karst process develops.

The karstified rock mass is characterized by an uneven distribution of porosity. Large-scale porosity (channels and caverns) also has an uneven distribution. They are often concentrated in certain zones and horizons. The karstification of rock mass decreases with the depth. Often it follows the exponential law of $\varepsilon = ae^{-bH}$ ((1) in Fig. 1). According to this distribution (Milanovic, 1980), the depth of karstification of Dinaric karst ranges from 200 to 250 m. The depth of karstification is to the transition zone above the rock not being karstified. Although this so-called base of karstification is represented as an approximate zone, there is no sharp boundary between the karstified and non-karstified rocks. The intensity of karstification of the zone close to the surface is usually high and is not subject to the exponential distribution.

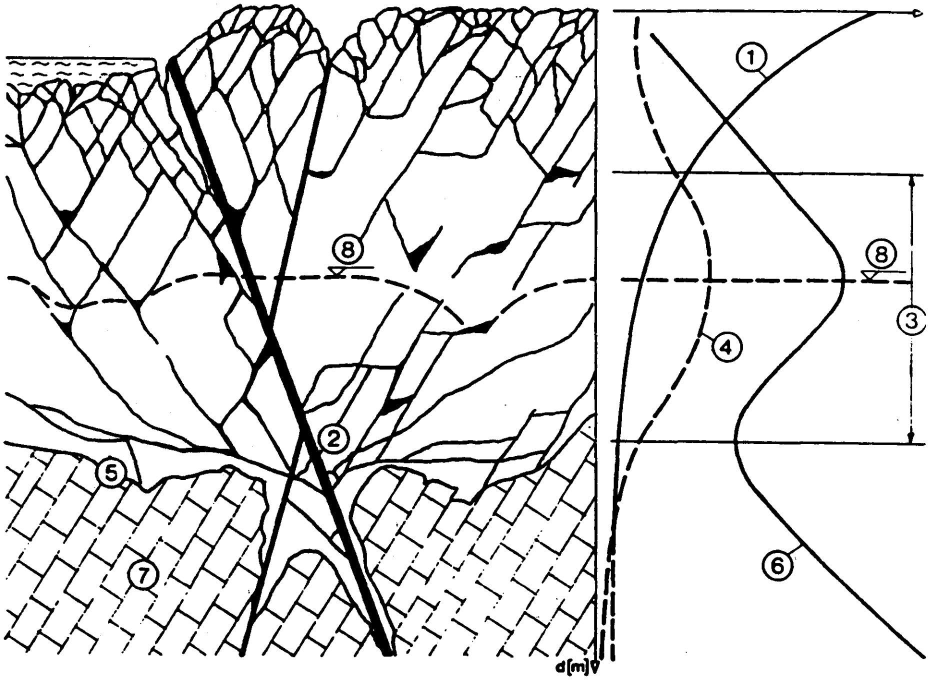

Figure 1. Schematic representation and depth. 1. Relationship between karstification and depth. 2. Zone of base channels. 3. Zone of water-table fluctuation. 4. Graph of vertical distribution of active karst porosity. 5. Base of karstification. 6. Diagram of electrical sounding. 7. Non-karstified limestone. 8 . Level of water table.

Characteristics of Underground Flow and Storage

The deeper parts of the underground drainage network are dominated by individual conduits that convey water toward the base channels ((2) in Fig. 1). These channels represent erosion base levels within the aquifer. They are formed in zones in which the karstification base is lowered with regard to its level in the surrounding rock mass. The concentrated conduits in this zone are then formed with the whole aquifer drained through them. Following the general characteristics of such a drainage system, it may be compared to a surface drainage system, provided that the volume of rock mass instead of the surface of the catchment is drained.

A subsurface system contrary to the surface-drainage system can be formed in the karst aquifer. For instance, the subsurface flow which originated from concentrated recharge may be sub-divided into an increased number of individual conduits, each leading to a different erosion base level. In that case, the structural and lithological characteristics and the neotectonic activity all play a decisive role in influencing the evolution of karst aquifers.

The highest intensity of karstification is in the karst systems with the largest storage capacity in the zone of water-table fluctuations ((3) in Fig. 1). This is also shown on the graph of vertical distribution of active karst porosity ((4) in Fig. 1). The graph of electrical resistivity ((6) in Fig. 1), represents the depth of karstification (Arandjelovic, 1969). According to this graph, the zone with the most storage is the region of water-table fluctuation.

Porosity decreases below the lowest ground-water levels. The base of karstification ((5) in Fig. 1) and the minimum water-table level coincide. Generally the slope of the base leads to the zone of aquifer discharge. The potentiometric line is generally inclined to this zone as well. However, locally this line ((8) in Fig. 1) is inclined toward the nearest karst channel with prevailing transport capacity in that part of the aquifer. The most active karst channels are directly above the base of karstification. The existence of karst drains below this level is not excluded, however, they are rare and are of limited transport capacities when they do exist. When deep tunnels are excavated, large karst forms have been encountered at depths exceeding 500 m. Deep boreholes have encountered karstified zones at depths exceeding 2000 m.

Specific Features of Karst Catchments

One of the basic consequences of the karstification process is the lack of organized surface drainage systems and fluvial erosion is replaced by the underground karst erosion. As a result of the evolution process of karst, ground-water flow often follows directions not in accordance with flow directions of the ancient system of surface drainage. The karst process destroys the individual surface drainage systems under some geological conditions; it connects them and transforms them in one catchment area. Under some other tectogenetic conditions this process destroys vast surface catchment areas abundant in water, and creates from these a great number of smaller karst catchment areas. The transformation of the surface catchment area (A) into several catchment and subcatchment areas (B) is represented in Figure 2.

Neotectonic movements and the karstification process can destroy and disorganize a well-developed surface drainage network. Ground-water flow replaces permanent surface-water flow, and only temporary flow in dry valleys remains at the surface. Zones of concentrated infiltration emerge, and the original catchment area becomes divided into several independent catchment areas. The limits of individual catchments are usually not permanent because they change with the flow. For instance, water from the subcatchment area (9) in Figure 2, flowing on the surface, becomes water of the catchment area (11), while that flowing underground in (9) belongs to catchment area (10). Each karst polje is a local erosion base level for a certain subcatchment area. Simultaneously, it belongs to the catchment area of a lower erosion base. It often happens that the karst polje, although being a morphologically uniform unit, does not belong to only one catchment area.

The generalized scheme of transformation processes from several separate aquifers into a single aquifer is presented in Figures 3 and 4. As a consequence of differential movements of separated tectonic blocks, the zones of concentrated infiltration are formed in cascades. Each of these zones of the poljes represents an erosion base of a separate aquifer in the initial phase of formation of karst aquifers. The karstification reaches the deeper parts of the rock mass below the erosion base adapting itself to the lowest discharge zone. It connects them into a complex hydrogeological entity provided the karstification intensity increases from higher levels down to the lower erosion bases. The erosion base of the highest individual catchment area ((I) in Fig. 3) receives water from the permanent drainage system. The following lower step ((II) in Fig. 3) represents the erosion base level for part of ground water as well as for part of the surface water. Ground water is partially discharged below this erosion base level towards the lower base, ((III) and (IV) in Fig. 3). The erosion base (III) receives water from the catchment area exclusively via the underground. It is also a bifurcation area from which water is partially drained in the direction of another catchment area. Water flows mostly in directions which enable the emptying of the entire aquifer ((IV) in Fig. 3). The bases (II) and (III) in Figure 3 are not active in the dry period, and water flows in the underground direction of the base flow. The aquifer adapts itself to the lowest erosion base in the next evolutionary phase ((IV) in Fig. 4). The karst process reaches the rock masses below the local bases and connects the separate karst aquifers. The depth of development of karst aquifers is adjusted to the lowest discharge point. The karstification base is deep below the bases ((II) and (III) in Fig. 4) in the zone of base flow.

Identification of Underground Watersheds

A complex investigation procedure is required to identify the underground or surface karst catchment and watershed. Considering the size of a catchment is one of the key factors in water-budget analysis, so an effort should be made to determine this size as exactly as possible. The orographic study of karst rarely provides satisfactory results for catchment area determination.

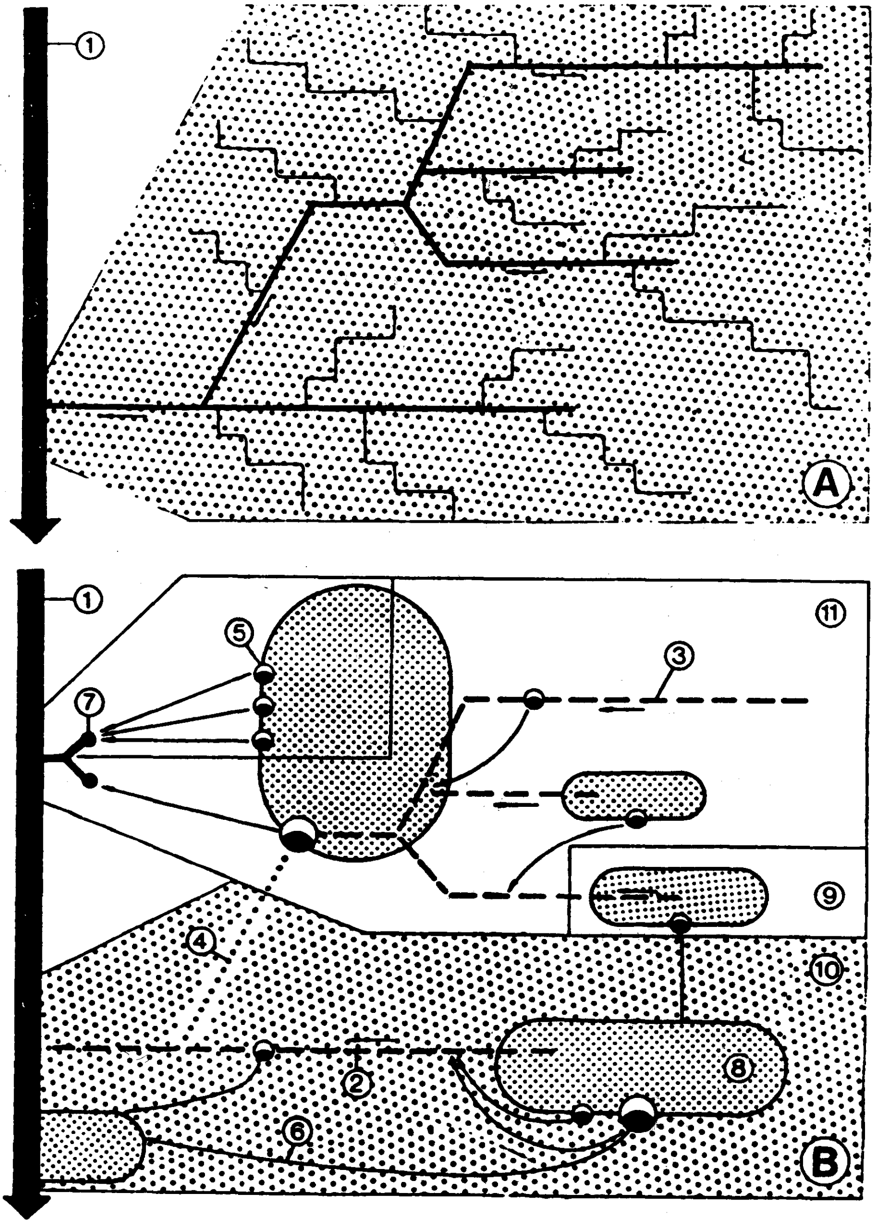

Figure 2. Scheme of karst aquifer evolution. A. Surface catchment area, before karstification. B. Subcatchments in some area after long time of karstification. 1. Main surface flow (regional erosion base level). 2. Permanent surface flow. 3. Temporary surface flow. 4. Dry valley. 5. Ponor. 6. Subsurface flow. 7. Spring. 8. Karst polje. 9. Highest subcatchment. 10. Subcatchment with increasing ground-water deficit. 11. The limit of the subcatchment developed by evolution of karst aquifer.

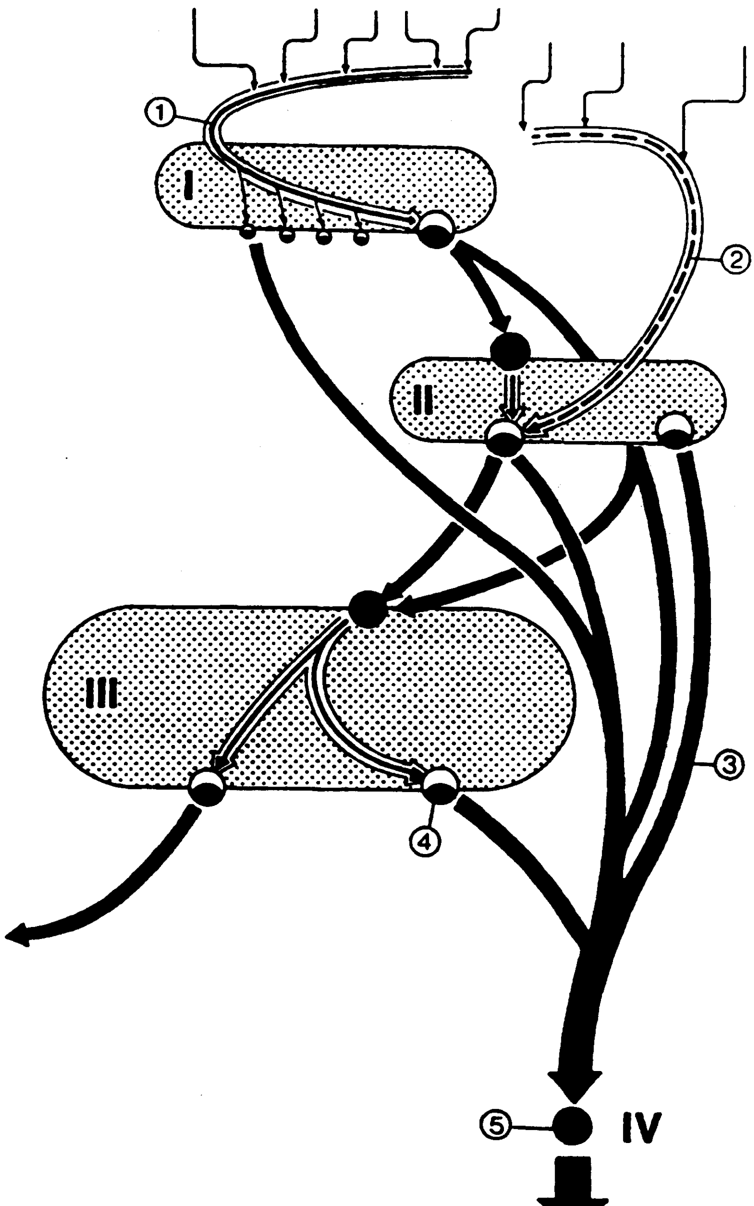

Figure 3. Model of transformation of separated karst aquifers into a single aquifer. 1. Permanent surface-water flow. 2. Seasonal surface-water flow. 3. Underground water flow direction. 4. Ponor (shallow hole). 5. Spring.

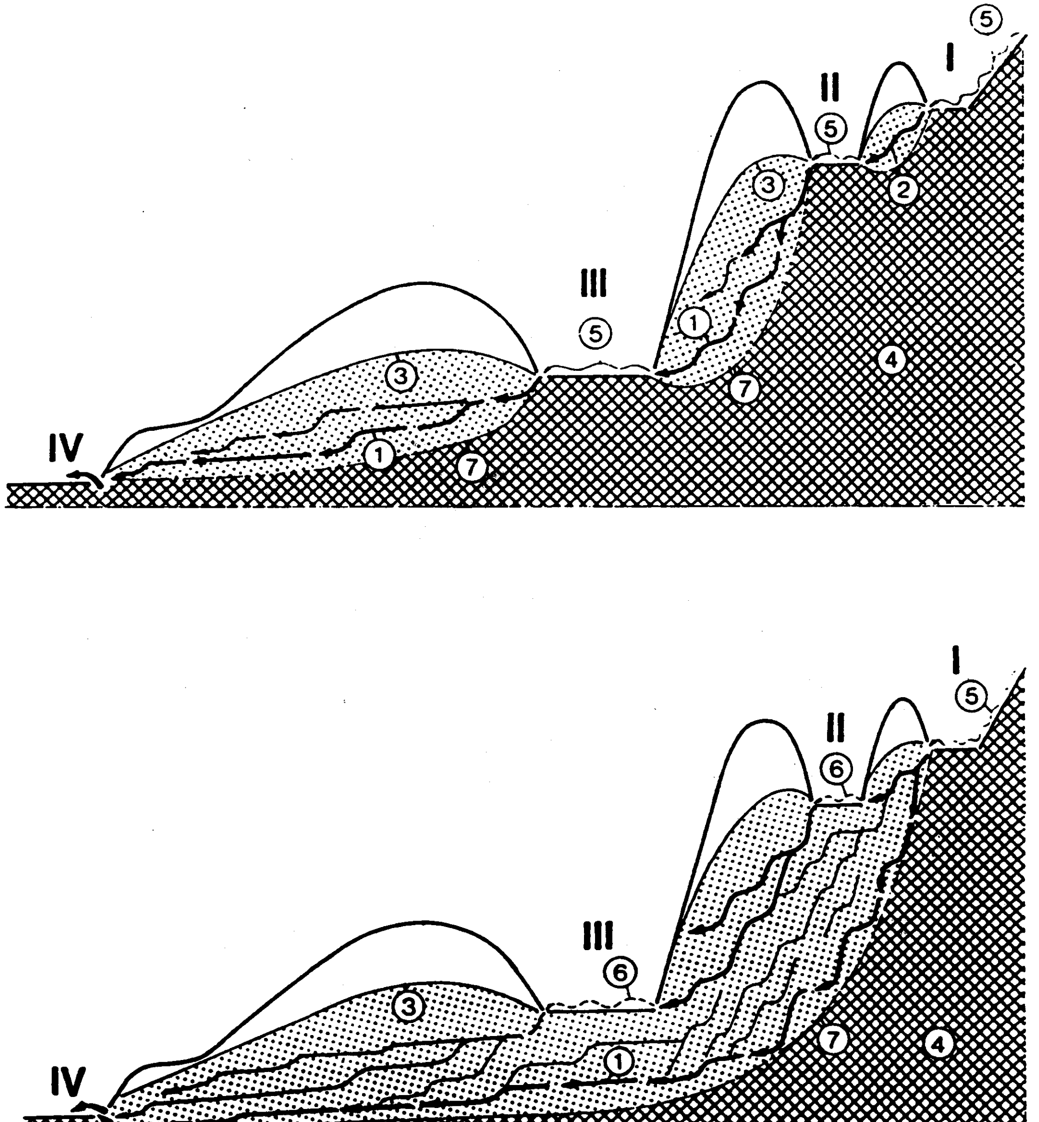

Figure 4. Model of transformation of separated karst aquifers into a single aquifer (schematic cross section). 1. Ground-water flows in separated aquifers. 2. Ground-water flows in single aquifer. 3. Highest level of water table. 4. Non-karstified rock mass. 5. Surface-water flow (permanent). 6. Seasonal surface-water flow. 7. Base of karstification.

No single investigation method can provide sufficient information for identification of the watershed of a karst catchment, with corresponding subcatchments, nor can it provide an understanding of hydrogeological characteristics of the karst aquifer of that catchment. Therefore, it is always necessary to undertake versatile investigation approaches. To become informed about hydrogeological and hydrological factors it is necessary to use a large number of investigative methods, such as geological mapping, surveying maps of hydrogeological features such as ponors, springs, estavelles, sinkholes, and caverns, speleological investigations, remote sensing, hydrological flow or water-level measurements and monitoring, tracing of underground flows, various geophysical methods, and hydrogeological monitoring. Lithostratigraphic and structural characteristics of karst represent the base of all analyses. The geological structure of rocks subject to karstification is determined by the methods of geological mapping.

Experience acquired so far indicates that investigations in the zone of concentrated discharge, the hinterland of the spring, make possible the collection of data and the quantification of most hydrogeological characteristics of karst aquifers and, through them, the characteristics of the karst catchment. The zone of concentrated infiltration, that is the zone of ponors or swallow holes, is the next important zone of investigation. Investigations performed in these zones usually provide information on inlets and outlets of a karst aquifer. The karstified rock mass, within which water storage and transport occur, is found between these two zones. Because of low and non-uniform porosity of karstified rocks, the characteristics of karstified rocks between inlets and outlets usually remain unknown. Collection of hydrogeological information on this part of aquifer is feasible only by investigation via piezometric boreholes. These boreholes often provide a contact with the channels of preferential groundwater flow, that is with the subterranean localized streams.

In contrast to investigations carried out in infiltration zones and the outflow areas, investigations in the central parts of the karst catchment and the corresponding aquifers represent complex and often uncertain exploratory undertakings. Identification of rupture tectonics by remote sensing methods, field checks of their findings, and geophysical prospecting are usually guided by the information provided by detailed geological maps. They enable a distinction between the potential exploratory sites which may establish the contacts with the subterranean concentrated flows within a karst aquifer, which are usually far away from the zones of water infiltration and outflow. Faults and fault zones as indicators and controls of karst processes, in addition to analysis of compression dislocation, represent the most significant geological factors in the selection of exploration sites. They make feasible the observation of water flow in karst aquifers. They may manifest themselves on the surface in the form of clear morphological features, and they affect the homogeneity of larger morphological forms, such as the lineaments along which the vegetation is considerably more intense.

Scarce vegetation cover and photogenic properties of the karst usually are conducive for remote sensing methods in analyzing structural, geomorphological, and hydrogeological characteristics of karst and in planning investigations. Global structures and regional faults can be clearly identified on satellite images. The photos taken from the heights of 1 to 4.5 km, in scales of 1:35,000, 1:25,000, and 1:5,000, offer the high quality data necessary for the analysis of zones of infiltration and outlets (springs) and especially for the identification of fault and fracture systems. These data are especially valuable for determination of location of borehole sites, since experience has shown that the zones with ruptures produce water concentration routes (Fig. 5) and more so the intersections of significant faults determine the approximate positions of parts of the aquifer with concentrated ground-water conduit flows. The character of fault zones should be taken into account. If they are sufficiently wide, with mylonitic clayey material, they may play an important role as a hydrogeological inhibiter or barrier.

The reconstruction of neotectonic activity within the evolution of karst processes represents an important step in the analysis of karst catchments. The movement of tectonic blocks is often opposite to the general neotectonic trends. Morphostructural entities are formed and are confined by tectonic discontinuities. Localization of these discontinuities is based on the geological conditions. This is important in the formation of structural models which are a starting point for spatial and temporal reconstructions of the evolution of karst processes and identification of karst aquifer watershed lines.

One specific feature of karst is an exceptionally dynamic regime of both surface- and subsurface-water flow. Transition to full saturation of the aquifer and consequent flooding of the surface of the catchment to the conditions of full water deficit is often abrupt. These changes occur in time intervals often shorter than 24 hours. These short periods are very important for investigations. The significance of direct field observations is often neglected, because the times of these variations are difficult to predict sufficiently in advance.

Base flow functions as a hydraulic system under pressure during periods of heavy precipitation. A high coefficient of correlation exists between the fluctuation of ground-water level in the piezometric boreholes, which reach the zone of concentrated flows, and the discharge of the spring through which the aquifer is drained. The coefficient of correlation is often higher than 0.95 and shows that the relationship is often close to a deterministic function. Because of it, it is very important to determine the position of concentrated ground-water flow as accurately as feasible, that is, to reach the zone of most intensive water transport characteristics.

HYDROGEOLOGICAL CHARACTERISTICS OF THE TREBISNJICA SPRING DRAINAGE AREA

Characteristics of the Watershed Divide

The Trebisnjica spring drainage area (Fig. 6) is between three drainage areas of the Neretva, Drina, and Zeta rivers and the drainage area of the Trebisnjica River downstream of the spring area. Three local erosion base levels determine the essential hydrogeological characteristics of this drainage area. They are the Gatacko and the Fatnicko poljes and the spring levels.

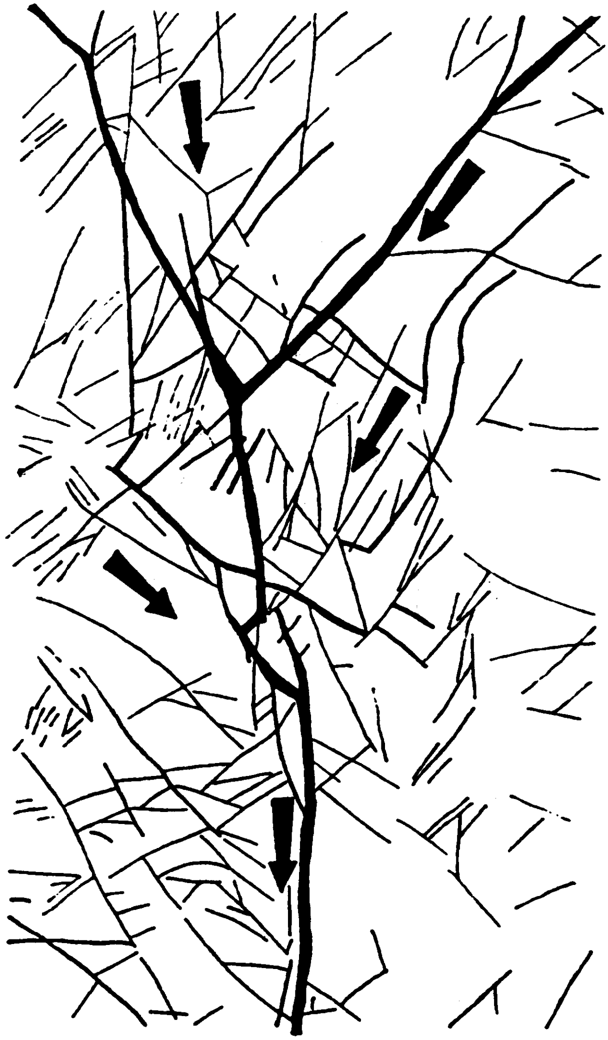

Figure 5. Direction of ground-water flow along the main fault system.

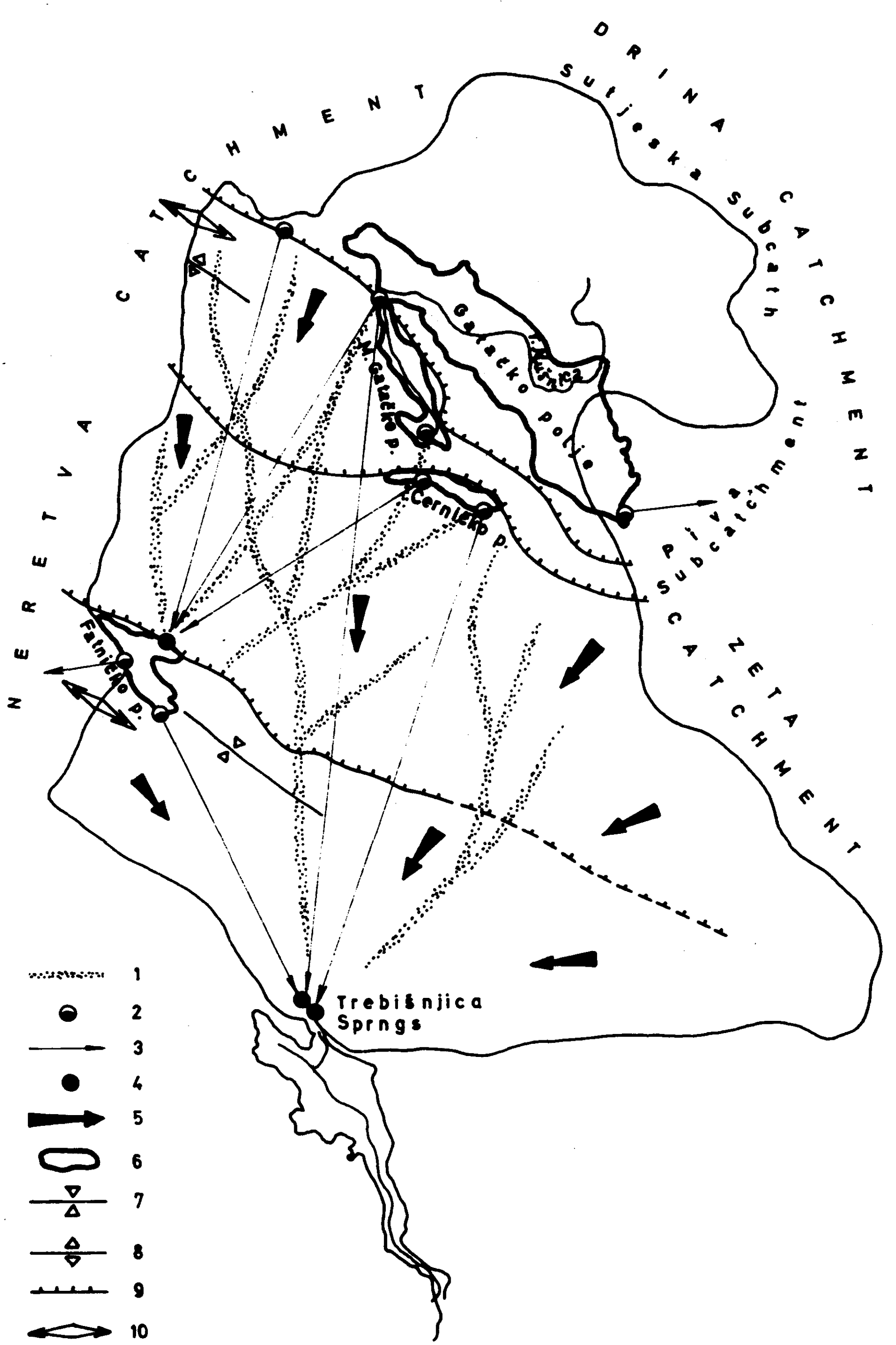

Figure 6. Simplified hydrogeological map of the Trebisnjica spring drainage area. 1. Zone of concentrated ground-water flow. 2. Ponor (swallow hole). 3. Underground water link established by dye tests. 4. Perennial or intermittent spring. 5. General direction of ground-water circulation. 6. Polje. 7. Syncline axis. 8. Anticline axis. 9. Reverse fault. 10. Zone of bifurcation.

Many investigations, such as detailed geological mapping, analysis of aerial photos, hydrological observations, determination of routes of ground-water flow, observations of a great number of piezometric boreholes, geophysical prospecting, and speleological investigations, have been carried out. The drainage area empties its water through the spring in a large cave. The eastern part of the drainage area empties water through separated source openings which are located in the vicinity of the cave spring. Both of these springs form the spring zone of the Trebisnjica River. On the basis of all the investigations performed up to now, the catchment area of the Trebisnjica River springs amounts to 1,144 km^2.

General Characteristics of Geological Structure

The Trebisnjica Spring drainage area covers part of a high karst geotectonic unit. Limestone and dolomite prevail over the clastic sediments. The basic structural characteristics of this unit correspond to the basic tectonic elements of the external Dinarides, namely the domination of folded structural elements, distinct linearity of the structures, direction of the Dinaric dip, and a dense net of faults. Formation of this area is connected to the Mediterranean geosyncline basin activity.

A negligibly small part of drainage area is made of Triassic dolomite. Nor are the Jurassic carbonate formations significant in the entire structure of the drainage area. Cretaceous sediments are the most developed stratigraphic unit. Cretaceous limestone with interlayers and zones of dolomite are continuous over the catchment. Only the northern part of drainage area is formed of the Upper Cretaceous clastic facies known as the Durmitor flysch. Both the Jurassic and the Cretaceous limestone and dolomite have been affected by karstification down to great depths. The Eocene flysch does not represent a significant lithostratigraphic formation; however, according to its hydrogeological role and location along the reverse fault at the all poljes in the catchment area, it has a huge effect on this aquifer. A large thickness of Neogene sediments has filled the tectonic depression of the Gatacko polje.

Evolution Process of the Trebisnjica Karst Aquifer

The evolution of the Trebisnjica karst aquifer is the result of complex tectonic erosional processes. It is schematically represented in Figures 3 and 4. Destruction of fluvial drainage systems was especially intensive in the Oligocene-Miocene period. This period is characterized by intensive folding. New tectonic dislocations featured by a dominant transverse component occurred in the post-Miocene period. It disturbed the homogeneity of hydrogeologically important structures. Dislocations and transverse fractures had a pivotal role in predisposing the development of the new erosion cycle. Fluvial processes gave way to karstification. The transverse fractures connect in the shortest way the cascaded bases of erosion. The capacity of karst erosion increased abruptly with the raising of the northern blocks. The Neogene sedimentation trench of the Gatacko polje lost the surface drainage and began to function as a local erosion base. Evolution of this part of the karst aquifer moved in two directions, eastward toward the Piva aquifer and southward toward the Trebisnjica spring zone. At the level of the Gatacko polje, two independent infiltration zones have been formed under the influence of these two erosion bases. One of most concentrated infiltration zones of this area was formed at the level of Gatacko polje.

A most favorable hydrogeological condition (ponor zone at level I in Figs. 3 and 4) was created. This resulted from water pressure of approximately 6 m during floods with the maximum inflow capacity up to 160 m^3/s. In the first stage of evolution of the karst aquifer, three step-wise independent karst aquifers developed between the infiltration zone at level I and the actual discharge zone at level IV. By adjusting to the present water outflow level, the separate karst aquifers between the Gatacko polje and the Trebisnjica River springs lost their independence. The karstification process spread over the entire rock mass, passing under the suspended flysch barriers, by connecting these aquifers so that the Trebisnjica aquifer formed as a unique karst system.

Location of the Dominant Water Transport Directions

The Gatacko polje prevails as an erosion base for surface water. The subsurface water flows through huge karst conduits forming the Trebisnjica springs drainage area. The long and deep north-south faults play a decisive role in this development. These tectonic systems developed preferential directions of ground-water flow. During the first stage of karst aquifer evolution, independent interconnections were created between the local erosion bases. The large elevation differences enabled intense development of vertical erosion which rapidly reached the lowest points of flysch barriers. They were transformed into the suspended barriers. Another preferential ground-water flow developed within this aquifer. This ground-water flow was predisposed by the most important transverse fault zone of north-south direction.

The important "tectonic crossing point" of this huge aquifer is shown in Figures 5 and 6. It is formed by two fault systems which cross each other. The fault of this rupture system, which stretches in the northeast-southwest direction, connects this crossing point with the infiltration zone of the entire drainage area. The fault of the other system connects this crossing point with northwestern part of aquifer. This second fault, in the southern direction of the tectonic crossing point, takes over the predisposed role of water transport (karst conduit system) of the whole aquifer. The tectonic zone which extends from this crossing point to the Trebisnjica River springs is characterized by a series of sinkholes, some of which are as deep as 100 m. Sinkholes are interconnected by a wide system of faults and cracks.

Basic Hydrogeological Characteristics of the Trebisnjica Aquifer

The described tectonic system predisposed the formation of the longest continuous ground-water conduit flow in this part of Dinaric karst area. It begins at the Gatacko polje and ends at the Trebisnjica Springs. Intensive water circulation through this system occurs in the wet season, that is in the period of high level of the water table. Straight-line distance between these two points is 34 km. Water tracers travel this distance in five days. During the dry season, the time of water travel is six times longer. The velocities of ground-water flow range between 0.9 to

14.0 cm/s, most often within 6 and 12 cm/s. The maximum water-table amplitude is 129 m within the region of the crossing point. The change of water levels in the period of raising and drawdown is rapid (Milanovic, 1986).

During dry periods only base flow occurs. The base flow is formed at levels of karstification base and along the dominant transverse faults. The base of karstification along these areas is of lower depth than the average depth of karstification.

Rapid saturation of the aquifer occurs during the wet period. This aquifer is characterized by a relatively small storage capacity but by exceptionally large transport capabilities. Both are also relatively small in comparison with the infiltration capacities of the ponor systems. The lack of aquifer storage space is compensated by the natural surface storage areas, that is the closed karst poljes depressions.

Characteristics of Spring Area

The spring area is formed in the bedded limestone of Upper Cretaceous age. Dolomite is significant in the lower parts of this spring area. According to the size and water yield of the spring area, two points of concentrated outflow of this aquifer dominate; they are the outlets of karst channels of high water transmissivity. These two outlets are separated by about 200 m. Even though they are interconnected and function as a unique zone of outflow from the complex karst aquifer, parts of the drainage area affect each of them differently. The main cave spring opening is affected more by the drainage area which extends toward the upper erosion base level of Gatacko polje.

The spring cave outlet bottom is at an elevation of 325.2 m above sea level. The cave opening is 4 m wide and 6.5 m high. Under natural conditions and during the summer period there is no outflow from the cave. Outflow of about 2 m^3/s occurs at the "aye" outlet approximately 200 m downstream of the cave. The maximal discharge of the spring zone is over 300 m^3/s. Hydrogeological investigations and geophysical prospecting, carried out in the background of the spring area, have proved that karstification has not descended down to the rock mass which is located deeper than approximately 300 m. Above this level, except for the karst channels, zones with rock porosity of up to 10% were identified. This noticeably exceeds the total porosity of the rock of the whole aquifer, which is less than 2%. Possible occurrence of karst channels under the base of karstification should not be excluded; however, they are very rare and, if they exist, have limited water transport capacities.

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HYDROGEOLOGICAL RELATIONSHIP IN THE CENTRAL DINARIDIC KARST

MIOMIR KOMATINA
Geozavod, 11,000 Beograd
Karadordeva, 48
Yugoslavia

INTRODUCTION

The central Dinaridic Karst has been studied and investigated by many national and international geologists since the beginning of the century. This remarkable terrain gave rise to widely known theories on the morphologic evolution of karst and on the ground-water distribution in this complex environment (Grund, 1903; Katzer, 1909; Cvijic, 1900 and 1926). The physiography of the terrain is largely influenced by large poljes, situated at five step-like bases of ground-water emergence and sinking. The investigation area of 19,000 km^2 is dominantly composed of structurally fractured and solutionally modified Mesozoic and Cenozoic limestone. This paper is concerned with a comprehensive approach to and detailed analysis of geological, geomorphological, and hydrogeological factors to identify drainage systems, karstic springs, and the preferential pathways of ground-water circulation within these units.

The large region of Dalmatia, western Bosnia, and Herzegovina, constituting the Central Dinaridic Karst bounded by the Zrmanja and Neretva Rivers (Fig. 1) has been studied and investigated for its geology and geomorphology by many geologists. It is for this typical karst terrain that numerous theories have been developed for the morphologic evolution of karst and the distribution of ground water in this specific aquifer system. These include the controversial theories of Grund (1903) and Katzer (1909) on water storage in karstified aquifers or on a step-like subterranean system and water flow only through karstic channels. A notable contribution to the study of large poljes was given by J. Cvijic, who named the territory described here.

The Central Dinaridic Karst terrain is characterized by a complex and specific hydrogeological relationship. The largest karst features developed are the poljes of Livno, Duvno, and Glamoc situated at five well defined step-like bases of water emergence and sinking (Fig. 2). The terrain was studied between 1952 and 1970 for its enormous water resource. Extensive hydrogeological investigation was carried out for the purpose of understanding the water supply, beginning with a complex study of the region (Komatina, 1975) and ending with detailed explorations in water reservoirs for the abstraction of ground water.

GEOGRAPHY

The terrain (19,000 km^2 in surface area) bounded by the Adriatic Sea, the Zrmanja and Neretva Rivers, the line through the source-springs of the Sana and Pliva Rivers, the town of Jajce, and the Vrbas River in the north consists of two

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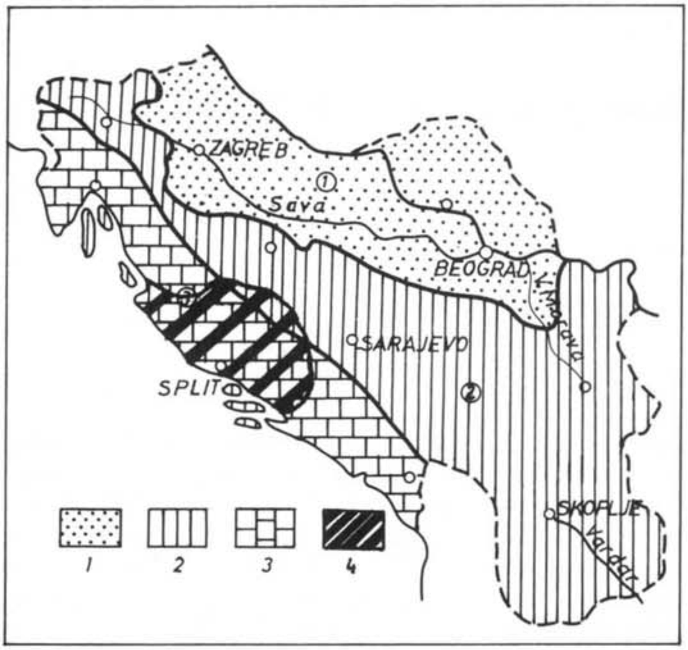

Figure 1. Geographic location of the study region. 1-3. Principal regions of Yugoslavia (1. Plain region; 2. Hill-mountainous region; 3. Dinaridic Karst); 4. Studied area.

Figure 2. The Plovuca River lost in Opaki Ponor, margin of Livno Polje.

morphologic units: 1) the less developed central Dalmatian Coast and 2) the hilly to mountainous area of the central Dalmatian littoral and mainland. The northern Dalmatian littoral is a smaller part of the terrain west of the Krka River; it is marked by a large karst plateau of the highest level of the Balkan peninsula (Cvijic, 1926). Mountain massifs are separated by large, long poljes. More than half the area north of Livno and Duvno Poljes lies at an altitude over 1,000 m. This is the 'West-Bosnian Height,' developed into an articulated arc. The highest massifs are those in the east: Cvrsnica (2,116 m) and Vran Mt. (2,074 m) separated by Dugo Polje.

The terrain is characterized by a relatively well-developed surface drainage, uncommon for most karst districts. Impermeable rocks provide for development of poljes and their surface streams. More specifically, surface-water drainage in a large part of the region is restricted to poljes, or sequences of poljes. The drainage of step-like poljes themselves is dominantly by underground flow in which a surface stream at a higher level sinks and emerges at the edge of the respective lower-level polje as a large karst spring, forming another surface stream. These streams typically have discharges on the order of 100 m^3/s.

The region of the Dinarides has the highest precipitation in south and central Europe. The amount of precipitation is directly dependent on the distance from the sea and on the altitude. Consequently, extremely low annual amounts of precipitation are found in north Dalmatian littoral (below 1000 m elevation) and the highest in the mountains of Cvrsnica, Vran, Ljubisnja, and Cincar (above 2,100 m elevation). Precipitation is the highest in the fall and winter, especially from October to December. The period March through May is rainy at the coast, and July and August are the driest months. The distribution of precipitation is more uniform in the mountainous climatic region where snowfall is more abundant.

GEOLOGY

The region of Dalmatia, western Bosnia, and Herzegovina has a complex stratigraphic column and an extremely diverse structural pattern. The thick complex of Alpine geotectonic products is represented by the dominant formation composed of carbonate rocks. Quite subordinate are clastic sediments, flysch and flysch-like deposits formed in the early Alpine cycle (Lower Triassic) and the late geotectonic stage (Eocene), and lake formations of molasse type from the Neogene. Mesozoic and early Cenozoic sediments were much folded and faulted during the major orogenic stages, Pyreneean and Savian, when the Adriatic platform arrowed the coastal belt and the crustal compression resulted in the development of diverse features. The features prevailing on land, north of Muc fault, Arzano, and Rakitno Polje, are those of crustal expansion expressed locally in a typical block structure. Two areas of specific geological columns and tectonic patterns are distinguished: 1) littoral belt and 2) mainland karst belt (Fig. 3).

The oldest rocks in the littoral belt are Cenomanian-Turonian dolomite, which forms the roots of numerous anticlines and the fronts of some thrusts. The dolomite is overlain with a series of bedded rudistic limestones of Turonian or Senonian age, large in thickness and extent, and foraminiferal limestones (Paleocene-Eocene). The Middle and Upper Eocene is developed in the flysch facies

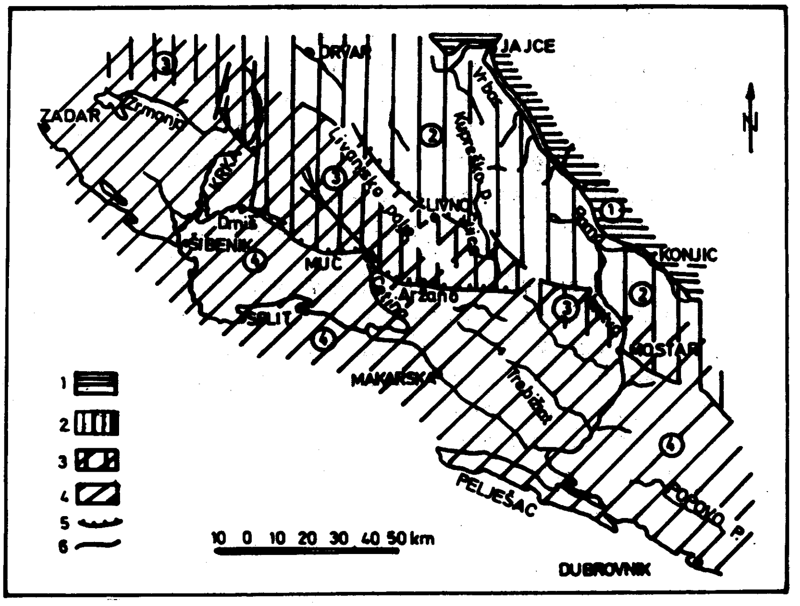

Figure 3. Geotectonic map of the Central Dinarides. 1. Inner Dinarides (eugeosyncline); 2-4. Outer Dinarides (miogeosyncline); 2. Crustal expansion zone (Inner Karst); 3. Transitional zone; 4. Crustal compression zone (littoral karst belt); 5-6. Bordering faults.

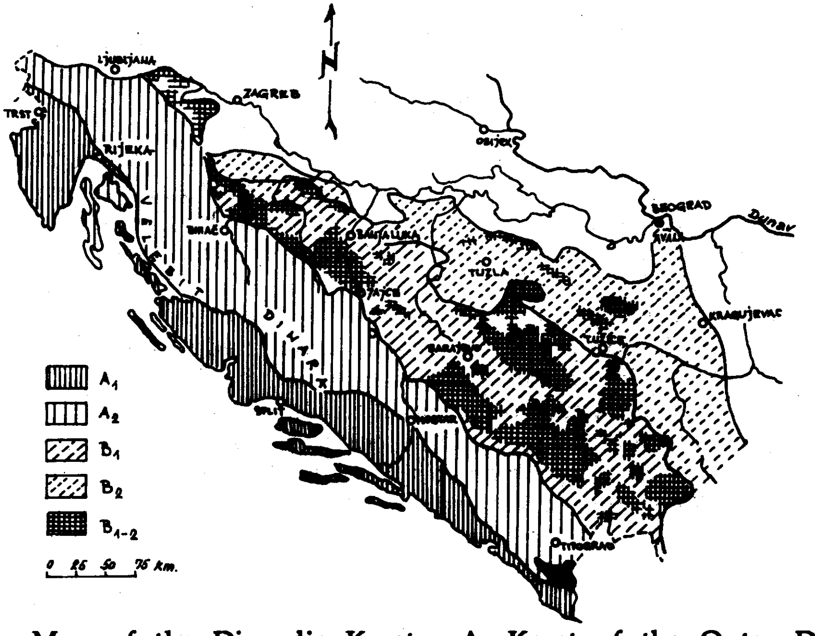

Figure 4. Map of the Dinardic Karst. A. Karst of the Outer Dinaridic miogeosyncline (A_1 - crustal çompression zone; A_2 - crustal expansion zone); B. Karst of the Inner Dinaridic eugeosyncline (B_1 - Karst and nonkarst region; B_2 - region of isolated limestone masses; B_{1-2} - extent of limestone and dolomite.

as a complex marl, sandstone, conglomerate, and sandy limestone succession. Folding is best developed in northern Dalmatia, where long folds extend parallel with the coast line, with associated systems of fractures and diagonal faults. Central Dalmatia and western Herzegovina are the areas of the 'littoral overthrust sheet', built up of Triassic and Jurassic limestones and dolomites of the Biokovo massif thrust over Eocene flysch.

Lithostratigraphic members in the mainland karst belt generally have a monoclinal dip. The older Triassic sediments form an irregular zone on the northeastern margin, from which the younger lithostratigraphic units extend to the southwest. The oldest rocks are flysch-like Werfenian deposits. The Middle and Upper Triassic is represented by a facies of massive dolomite. The Jurassic column dominantly consists of calcareous facies, and so does the extremely thick Lower Cretaceous interval. Upper Cretaceous limestones are developed northeast of Drvar and south and northeast of Livno and Duvno poljes, respectively. The structural pattern of the mainland karst belt is controlled by faulting. The area between Drvar and Bosansko Grahova, for instance, is characterized by parallel synsedimentary faults, with a monoclinal distribution of sediments between the faults. Other areas of western Bosnia are faulted into large limestone blocks, such as Vran block or the block of the Pliva source hinterland.

GEOMORPHOLOGY

The extreme development of the karst, in intensity and size of its features and in depth, resulted from favorable climate and morphology, great extent of vertical tectonic displacements, and the extremely suitable lithology and structural pattern. It resulted in a classical karst region, one of the type karst terrains of the Mediterranean and the world.

In consideration of the orogenic and epeirogenic sequence of events in the geosynclinal domain of the Outer Dinarides and the geological evolution of holokarst, Milovanovic (1965) stated that the Dinaric holokarst consisted of a combination of the recent karst and stratigraphically different palaeokarst, and that this complexity was basically controlled by the epeirogenic and orogenic dynamics which also controlled the effect of exogenic factors. He considered Dinaric holokarst a complex phenomenon, largely controlled by the lithostratigraphy and tectonic events in the region.

During the Neogene and Quaternary periods, large mountains and intermontane depressions were formed in a time of generally positive epeirogenic movements. These events provided for a continuous process of karstification which developed to great depths along the faults. The most suitable conditions prevailed in the Pleistocene, at the time of glaciation and interglaciation, when underground channels drained large amounts of cold water from the melted ice that gained erosive and corrosive power with the land uplift. The sea level during the early Pleistocene is estimated to had been 90-100 m higher than at the present time, which provided at the coastal-belt uplift stages for the development of a dense network of karst drains. This is how recent karst regions developed, and they primarily control the regional hydrology, that is, the distribution of ground water in the Dinaridic karst area (Komatina, 1968).

Karst Features

The geomorphological regionalization is coincident with the earlier given geological regionalization because the dependence of karst development on geological factors is more than evident (Fig. 4). Particularly well marked is the difference between the littoral and mainland karst belts, i.e. the zones of crustal compression and expansion. One of the major dissimilarities is the presence of large-size depressions, poljes, in the inland belt, and large, ground-water paths along regional fault zones, unlike those in the coastal belt.

Differentiated by their large size are poljes (Fig. 5). These features are not only most conspicuous in the central Dinarides, but are also the most complex karstic phenomenon. Poljes are commonly closed depressions, drained under ground. Most of these depressions are periodically inundated when the inflow from springs exceeds the discharge into sinkholes. Most poljes are structurally controlled, and their present shapes have been influenced by exogenic geological processes in a less resistant lithological environment. The largest polje, even by international standards, is Livno Polje.

Another karst feature is the karst plain; the most uniform is the Kistanje. In contrast to Kistanje, numerous karst plains around Kupres and Ravna poljes are covered with many sinkholes that form a distinctive cockpit karst. Next to poljes, the typical karst surface features in the west-Bosnian region are sinkholes. Ponors are associated with poljes, uvalas, and dry valleys (dolinas). Their position and distribution within a polje is primarily influenced by the nearest base of erosion. Ponors are always found on the side of the erosion base.

GENERAL HYDROLOGY OF THE REGION

The Central Dinaridic region, extremely complex in geology and hydrogeology, is an interesting and unique part of Yugoslavian hydrology. A complex relationship of rocks of various permeability, the high degree and depth of karstification in carbonate rocks, several step-like levels of ground-water drainage bases, and other factors are responsible for the extremely complex distribution and flow of water in this karst district of Yugoslavia.

The region is built up of impermeable rocks at the base, dominantly structurally fractured and solutionally modified Mesozoic and Cenozoic limestone. However, though scattered within the limestone mass, impermeable rocks also play a notable role in the ground-water distribution. Within impermeable rocks, most erosion bases are formed on which water collects from the surrounding limestone terrains. This concentrated water drainage caused the formation of a multitude of river systems and karst springs. Hydrogeological relationships north of Muc fault, Arzano, and Rakitno Poljes, within the inland karst, differ from those in the littoral strip. The difference is not only in the areal position and the relationship of rocks with different hydrogeological functions, but also in the distribution of ground water and its drainage pattern.

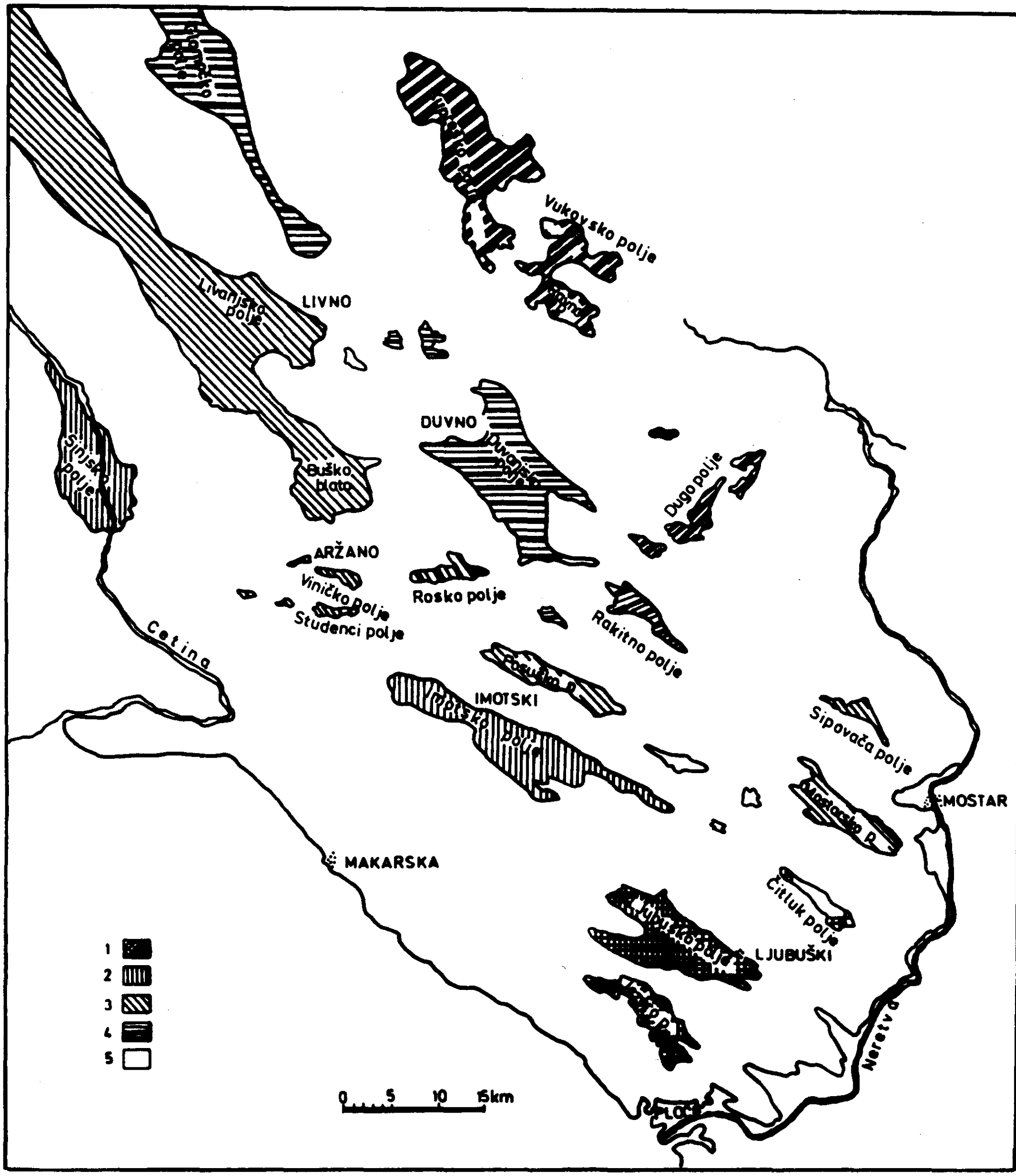

Figure 5. Poljes of western Bosnia, Herzegovina, and Dalmatia. 1-5. Identified levels and their altitudes (1. 25 to 80 m; 2. 196 to 320 m; 3. 571 to 716 m; 4. 858 to 978 m; 5. 1,078 to 1,278 m).

Relative to the intensity of individual factors, such as chemical composition and structural damage of rocks, or precipitation depth, the effect of karstification in limestone is not uniform (Fig. 6). Limestone, with the exception of thin-bedded marly limestone, is a permeable and discontinuous environment providing for flow of water along faults, systems of longer or shorter fractures, karst channels, and a variety of underground, not always interconnected, water pathways. Where widespread, dolomite may have a double hydrogeological effect, it is either virtually impermeable or it is permeable. The Triassic dolomite in the mainland karst zone and the Cenomanian dolomite forming axes of long anticlines in northern Dalmatia are impermeable. In contrast to these rocks, permeable dolomite lies in the frontal parts of overthrusts or alternate with limestone. Other impermeable rocks are Eocene flysch and Werfenian sediments. Like impermeable dolomites, these rock masses had an important influence on the distribution, orientation, and drainage of ground water. The greatest effect on the specific hydrogeological features of the terrain, including a number of step-like bases of ground-water drainage, was made by the different altitudes of impermeable Neogene-lake formations.

The distribution of ground water is influenced the most by recent cavities in limestone, developed for the most part some 100-200 m below the surface. Paleokarst features prevail downward, generally filled with bauxitic or other material, destroyed to some degree by recent tectonic events.

An evident relation between the faults intersecting limestone and karstification is expressed in the shallow subsurface and relatively deep zones of recent karst. In the former zone, rarely deeper than 20 m, the internal karst features of limited size are very extensively developed, connected, and conformable with the surface (Fig. 7). At depth, karstification advances along structurally displaced zones and faults, always to levels controlled by topography or the respective drainage base. This deep karstification is expressed in infrequent, more-or-less isolated, large cavities in nearly compact rocks (Fig. 7). It reaches a depth of no more than 200 m, as indicated by geophysical survey and permeability test data. Rock porosity varies from one place to another, being 8-12% in the subsurface, and 3-6% (effective porosity 2-3%) in deeper zones.

Ground water circulates through large channels (turbulent flow) and a system of fractures (laminar flow) within the zone of advanced karstification. The degree of hydraulic intimacy of collectors is dependent on their being in a subsurface zone or a zone of deep karst in a water-bearing environment; it is maximal in a submerged (in recent geologic time) subsurface zone of collectors, lessening where the epeirogenic uplift was greater. In the latter case, only two or three preferential directions can be marked on the surface by ponors or sinkholes developed along a fault or fault zone. The greatest concentration of underground streams is associated with the terrains of concentrated water replenishment, i.e. districts of large poljes.

Piezometric observation data have shown for numerous localities of the Central Dinaridic region that the difference between extreme water levels in one year usually was below 10-15 m, that is, the zone of horizontal circulation was of relatively small thickness. This does not exclude the possibility of a water-table

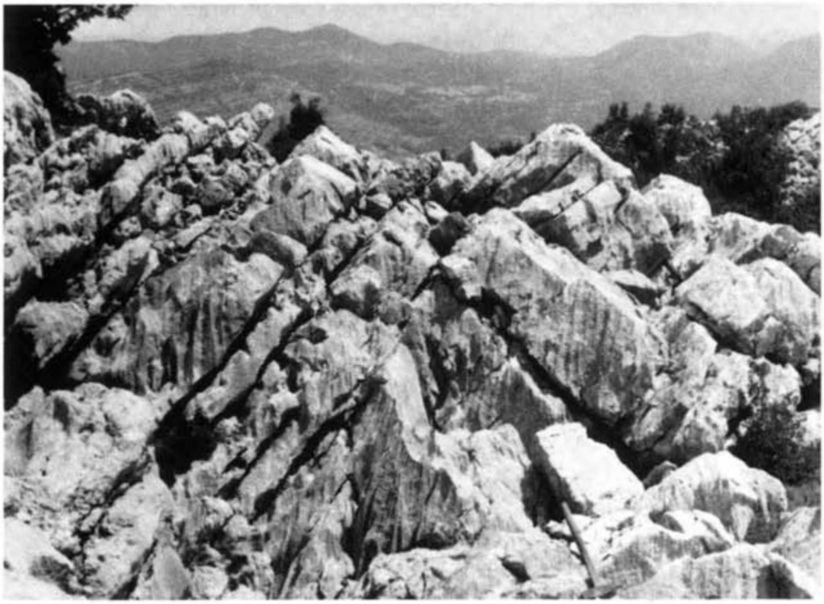

Figure 6. Photograph of karstified Turonian bedded limestone.

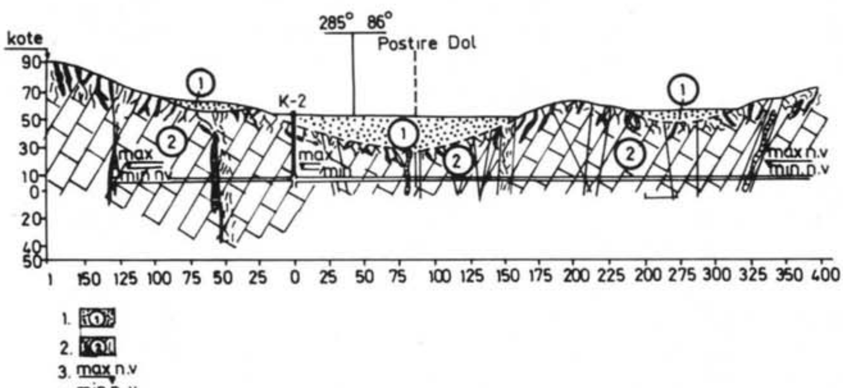

Figure 7. Hydrogeological cross section of intake structure K-2 in Postire Valley on the island of Brac. 1. Quaternary sediments; 2. Limestone; 3-4. Extreme positions of water table.

fluctuation up to 100 m, in zones of secondary (local) bases of erosion or in Biokovo hinterland, for example, where after storm rainfalls, water is retained in rocks forming a backwater. Vauclusian circulation of water is far less important than horizontal flow; it does not exceed the depth of a few tens to a hundred meters below the minimum water table. The deep-water flow is not given any greater importance because more than 80% of recoverable reserves in the analyzed drainage areas is controlled by respective bases of erosion. Note that the principal income in the ground-water budget is the percolation of ground water, which is very high in the Central Dinaridic Karst, ranging from 75 to 91% of the precipitation.

The velocity of ground-water flow, estimated using tracers in 55 localities of the studied area, varies within a range from 0.5 to 35 cm/s (average about 3.5 cm/s). In areas of undeveloped morphology, it is more of the order 1.0 to 1.5 cm/s, and in large poljes, 5-15 cm/s. That is why the fictitious ground-water gradient widely varies between 0.0043 and 0.047.

The concentration of underground flow paths into one or a small number of discharging outcrops is a hydrogeologic feature characteristic of the Central Dinaridic Karst. Through the geomorphological history of recent karst, preferential paths of ground-water flow advanced toward lower levels in respective bases of erosion, where the main paths are presently developed. A large number of drainage systems were thus developed, or hydrologic units of a lower order. In the area of 19,000 km^2, only 55 large springs have been registered as a consequence of the ground-water concentration. This means that each emergence of 4 to 9 m^3/s drains most of the replenished ground-water resource in a karst area of 340 km^2.

However, the picture of the identified drainage areas is not that simple. The drainage systems differ in pattern, size, altitude, and order. Each of them is specific in geological, geomorphological, and hydrogeological features, and in the distribution and discharge of ground water. Because the general direction of ground-water flow toward the principal gravity destination -- the sea -- is much better expressed than the opposite direction, one of general drainage pattern features is a marked asymmetry.

The watershed of the highest order is the boundary between the systems draining into the Adriatic or the Black Sea. The division of streams between the north and south was largely controlled by impermeable Triassic dolomite and subordinately the Neogene of Glamoc and Kupres Poljes. The Adriatic system drains an area of 14,860 km^2, whereas only a fifth of the Central Dinaridic Karst is a part of the Black Sea drainage system. This relationship is the expression of a principal karst feature in general -- to drain its waters primarily to the lower base of erosion.

Ground-water outcrops are numerous, with many intermittent springs, estavelles, and submarine springs. Water emergence or exurgence is the most important type of ground-water outcrop. Ground water emerges at the contact of the limestone and impermeable rocks. Cvijic (1900; 1926) referred to these outcrops as contact springs (Fig. 8). The degree of uniformity for contact springs may vary; it is more often relatively high (the ratio minimum to medium to maximum flow is

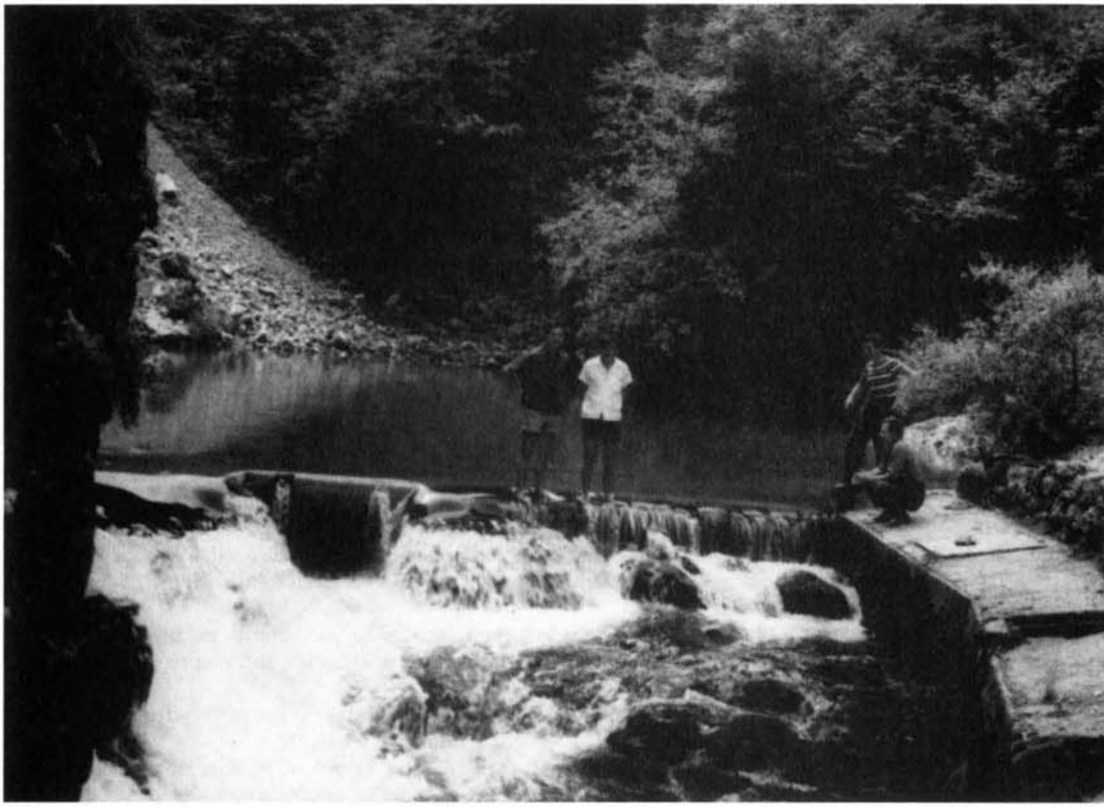

Figure 8. Bastasica contact spring at the town of Drvar.

1:2.5:4 for the Pliva spring and 1:2.2:3.3 for the Rama spring in Kovacevo Polje). Also notable, and not uncommon, is the group of siphonal springs termed so for the siphon-like upturned end of the discharging karstic channel (source springs of the Cetina, Opacac in Imotsko Polje, Klokun in the Tihaljina Valley, Crno Oko in the Neretva Valley, etc.) Exurgences, which are outcrops of lost rivers or are directly influenced by floods in poljes, are a specific case. These springs are extremely sporadic and intermittent (Rumin at Sinj, Ricina in Busko Blato, Bastasica near Drvar, etc.). The ten largest water exurgences, in the descending order of their average discharge, in m^3/s, are the following: Ruda spring in Sinjsko Polje, 20; Krupac in Kovacevo Polje, 19.2; the left Pliva source and Klokun, 14.0 each; Bug in Kovacevo Polje, 12.92; Grab in Sinjsko Polje, 11.0; Varvara in Kovacevo Polje, 9.47; Crno Oko, 9.4; Jadro at Split, 8.46; Opacac in Imotsko Polje, 7.38.

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KARSTIFICATION AND GROUND-WATER HYDRAULICS OF THE INTERIOR OF LARGE CALCAREOUS MASSIFS: THE CASE OF GIONA MOUNTAIN IN CENTRAL GREECE

PAUL G. MARINOS
National Technical University of Athens
Patision 42
Athens 10682
Greece

INTRODUCTION

Large continuous calcareous massifs occur in central Greece, the Peloponnese, and other zones of the Hellenic country. Exactly what state prevails in the interior of the karstic mountains at great depths and far from the areas where springs appear is still unresolved. This question arose with the construction of the large tunnel in Mount Giona, of the Mornos-Athens water conduit, and directly concerned problems of its construction and operation. The tunnel now traverses this mountain at a length of 14.6 km (in an approximately east-west direction, parallel to the coast), at an altitude of 377 m, beneath a covering of 1,500 m, and with its central sections 14-20 km from the shores of the Corinthian Gulf, where the ground water of Giona is drained and forms coastal springs (Fig. 1).

THE HYPOTHESIS

At the beginning of the project, the intense karstification of the surface of the mountain and the drainage towards the low points of the coastline led to the hypothesis that the tunnel would pass through karstic limestone, but more or less above the karstic water table (Fig. 2A). The tunnel was expected to be within the transfer or conveyance zone of ground water where serious danger of sudden inflows occurs during floods but where no permanent underground water exists. This water table would be considerably lower in areas of high permeability and easy discharge to the coast. It is possible that the water table would reach the tunnel or be a little above it in the parts of the calcareous massif which have fissures or only thin karstic channels, sectors in between the main karstic network. A few investigatory drillings, however, gave indications that the limestone is not karstified at depth but is only finely fissured. In such a case, the water table could be over 100 m or more above the level of the tunnel, which means perennial, but low-yield drainage would occur from the tunnel (Fig. 2B).

THE REAL SITUATION

The conditions which appear within and deep inside the mountain are more composite and, to a certain extent, combine these two types (Marinos, 1978, 1982). Beyond the endmost portions of the tunnel and the flysch, the limestone shows a short distance of karstification ranging from small to large channels (Fig. 3). From the entrance of the tunnel toward the interior of the mountain, the limestone shows no karstification. Its massif runs through a network of fine joints which display no enlargement. The karstified massif is thus restricted to a zone whose

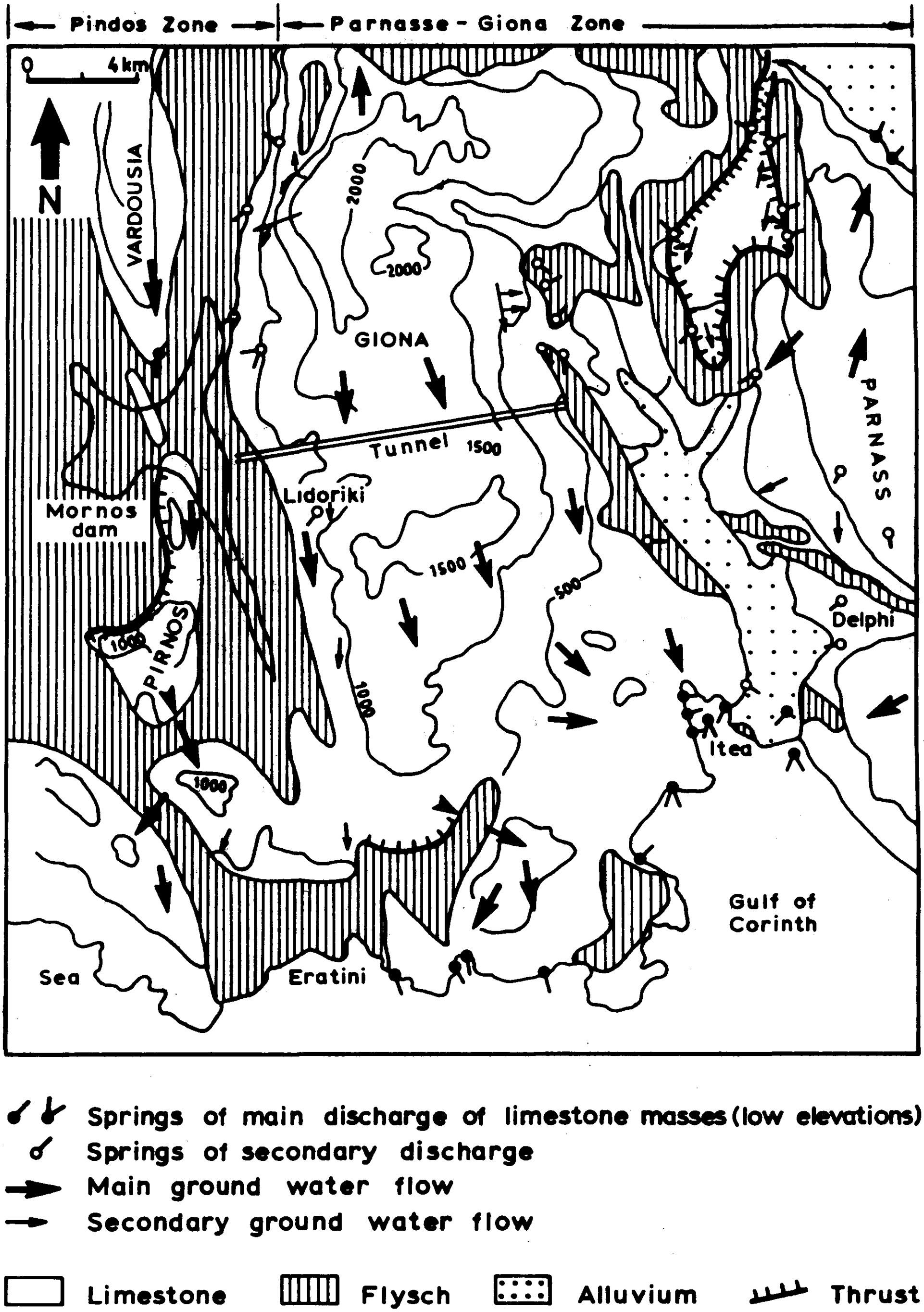
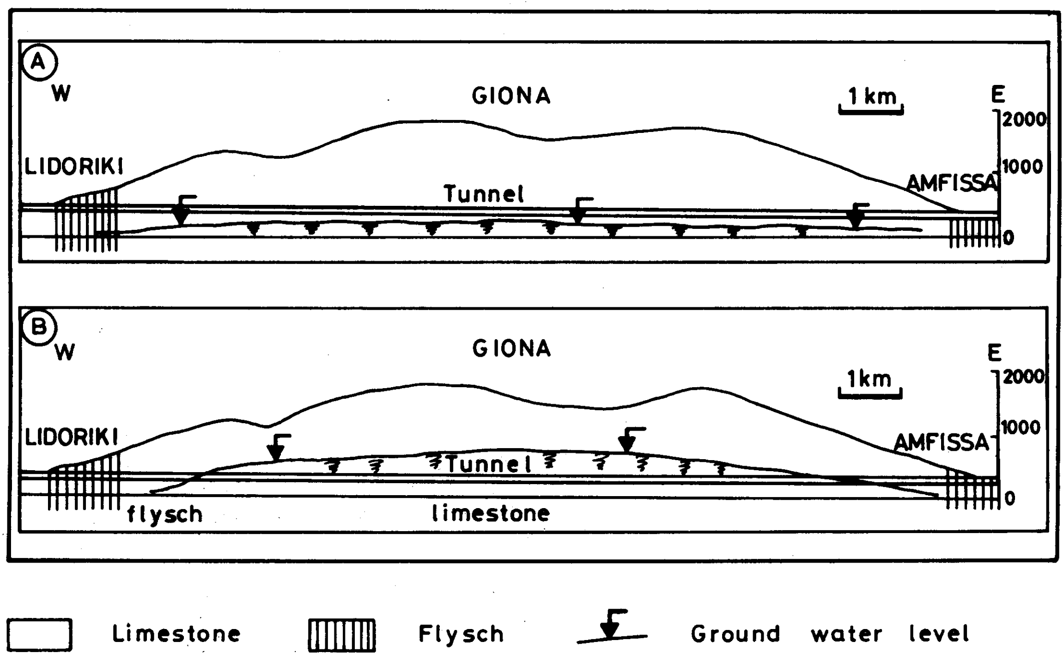


Figure 1. Hydrogeological map of Giona.



Hypothesis A: Karstic limestone in the interior of the mountain. High permeability (holokarstic situation).

Hypothesis B: Limestone not karstified at depth, but only fine-jointed. Low permeability.

Figure 2. Schematic of two hypotheses of ground-water occurrence in Giona.

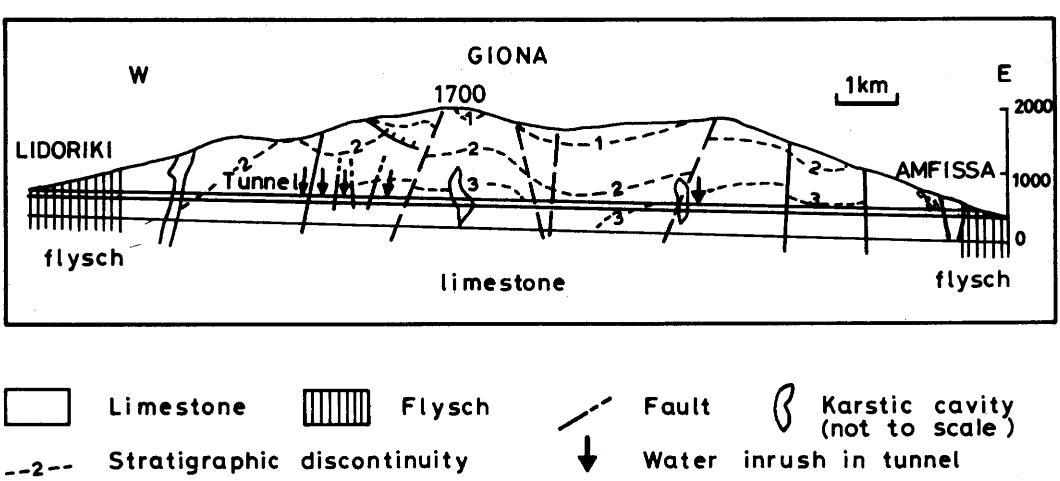


Figure 3. Longitudinal section of the Giona tunnel.

thickness does not permit it to reach the interior of the mountain, even in places of much greater altitude than that of the karstic springs which drain the mountain at its margins.

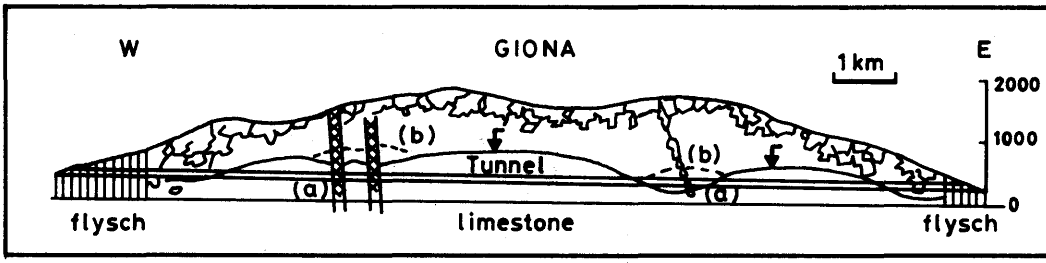
The water table where there is karstification is below the position of the tunnel and rises above it in non-karstified areas. In those areas the limestone has low to very low permeability and a flow that resembles a porous medium. Its drainage by the tunnel is barely perceptible and in the form of "transpiration" or drip flows.

Throughout its length in the interior of the mountain, the tunnel cut through only two karstic conduits (Fig. 3) which, because of their orientation transverse to underground flow, constitute no more than an exception and do not change the general non-karstic characteristics. These conduits, with a span of a few tens of meters, were met at 9.8 and 6.5 km from the western entrance of the tunnel. When the first conduit was drilled, water was freed under pressure, and then the flow balanced into small amounts. The second conduit was partially filled with clay, sand, and gravel without water, but with clear indications of underground flow. After a heavy storm on the mountain, a flood reached the gallery with a delay of only 8 hours. The water drained away within about a week. The active hydrogeological role of these conduits - the role of a zone of transfer of water to an underlying saturation zone - appeared once again when they refused to receive the water which the tunnel had been discharging to them. It was at the time of year of high rainfall which had raised the water table to near or above the level of the tunnel at that point.

This karstic tube forms an axis of drainage according to the model of a distinct discharge which significantly lowers the high water table of the massif. The difference is that in the interior of the mountain the appearance of such axes is extremely rare, and the surrounding massif is, in essence, predominantly non-karstic even in the form of fine channels.

The description of the interior of the mountain is completed by the presence of approximately vertical faults with or without a mylonitized zone, but which do not show karstic features or voids along their discontinuities. A few of these fault zones displayed water problems, especially in the sections between 9.3 and 10 km from the Amfissa side (Fig. 3). In all, more than 1,500 m^3/hr of water was entering the tunnel, of which approximately 400 m^3/hr was being supplied by a single fault in its slack mylonitic zone (a few meters wide). The fine-jointed limestone between the water-bearing faults was always of very low permeability, with drainage within the tunnel in the form of transpiration and drip flows.

The water-bearing faults increase the underground hydraulic heterogeneity of the interior of the mountain. These faults are fed by the karstic and highly permeable portions of the surface of the mountain. A water column is created which, according to the geometry of the faults and their discharge capacity, can maintain a significantly raised water table despite the high permeability at these zones. This column, according to its supply and yearly operation during periods of



Flysch

Karstified limestone zone of medium-high permeability

Fine-jointed limestone of low permeability only

Fault zone with limestone of high permeability due to fragmentation (not to scale)

"Lonely" karstic channels at depth

(a) Water table

(b) Level at high floods

Vertical position of water table not to scale

Figure 4. Underground hydraulic regime of Giona.

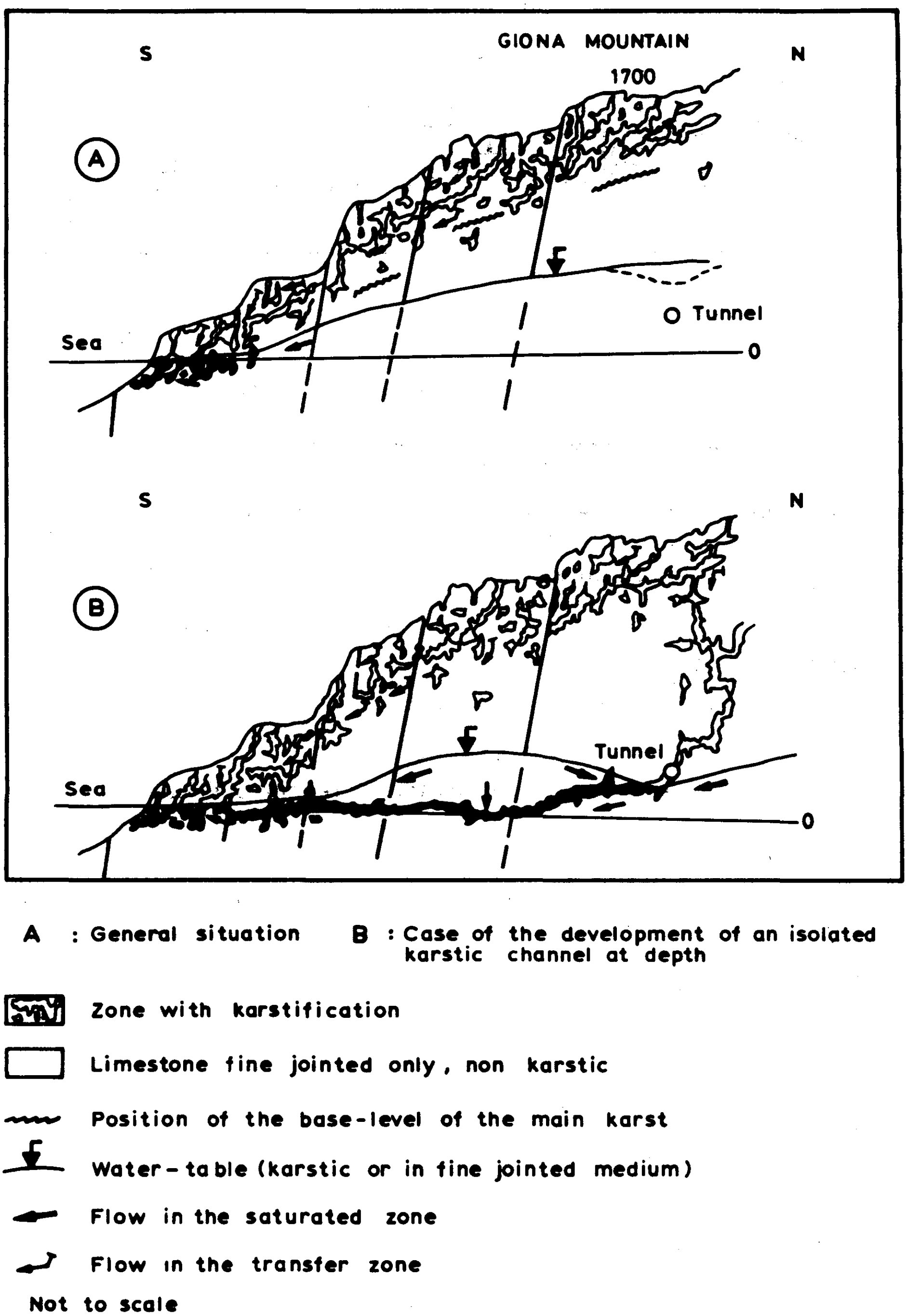


Figure 5. Underground hydraulic regime of two cross sections of Giona.

high supply and water table, fills the surrounding massif fine-jointed limestone and other times drains it.

CONCLUSIONS

The karstic and hydrogeological condition of the interior of the mountain appears as follows (Fig. 4 and 5). The interior of the calcareous mountain appears not to be karstified. Karstification seems to stop at a depth of a very few hundred meters and creates a karstic zone which proceeds in stages parallel to the surface of the mountain. The paleogeographic development of the area, with gradual surface erosion and leveling, and changes in the relative level of the sea, contribute to the formation of such an underground karstic geometry. Beneath the karstic zone, the limestone is not karstified but has a fine network of closed joints and low permeability.

The water table has low to medium gradients in the area directly uphill from the springs at the termination of the karstic zone, while further on toward the interior of the mountain it has steeper gradients. The concept of the base level of the karst, in its classical form, has no meaning except beneath the peripheral parts of the mountain and chiefly toward the side of water discharge.

Deviations from this generalized regime in the interior of the mountain, are a limited number of zones of high-permeability fault axes which have brecciated the surrounding limestone, or rare selective karstic channels. These zones or axes of high permeability receive the water which is conveyed to them by the karstic area of the surface of the mountain and, as isolated drains, move it toward the outlet at the sea. A reduced capacity of these zones to move great quantities of water and difficulty in its removal by the karstic drains of the coastal zone, creates a raised water table, often at high altitudes, which loads or drains the principal low-permeability limestone massif.

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CONDITIONS HYDROGEOLOGIQUES DE QUELQUES AQUIFERES KARSTIQUES DE LA CRETE (GRECE)

par Borivoje F. MIJATOVIC (*)

RESUME

Les problèmes abordés dans le présent chapitre concernent les ressources en eau souterraine karstique de l'île de Crète où les conditions hydrogéologiques et hydrodynamiques des systèmes aquifères karstiques sont d'une grande complexité, rendant difficile leur mise en valeur.

Toutefois, le contexte géologique existant permet d'envisager la possibilité d'une exploitation rationnelle de ces aquifères dans certaines régions de l'île, notamment sur le littoral septentrional.

ABSTRACT

The matters treated in this chapter are connected to karst hydrogeology problems at Creete Iseland, where we have just point out on decesive role of the carbonate rocks, namely on the hydrogeologic and hydrodynamic conditions in the course of development and the life of karst aquifers system.

Actually, the existing geological setting leads to cennection of certain possibility for rational exploitation of ground water from the karst aquifers within particular regions of this iseland, especially along the northern coastal part.

(*) Hydrogéologue - BELGRADE (Yougoslavie)

SOMMAIRE

1 - Introduction

La longue et légendaire histoire de la Crète a conféré une atmosphère de mystère à l'île tout entière, particulièrement à cause de son relief typiquement karstique. Berceau de la civilisation méditerranéenne, sa prospérité fut certainement favorisée dès l'origine de l'époque minoenne (entre 3000 et 1400 av. J.C.) par l'abondance des ressources en eau de ses aquifères karstiques, littoraux et continentaux. En effet, grâce à leur coefficient d'infiltration élevé, ces aquifères donnent naissance à de grandes sources dont certaines peuvent écouler jusqu'à plusieurs dizaines de m3/s. Par ailleurs, il existe aujourd'hui des possibilités nouvelles d'une exploitation plus rationnelle de certains de ces aquifères karstiques en rapport avec le contexte géologique de certaines région de l'île, notamment sur le littoral.

On s'est efforcé dans le présent chapitre de faire ressortir le rôle prépondérant des roches carbonatées dans la répartition des ressources en eau de la Crète en examinant plus spécialement les conditions hydrogéologiques et hydrodynamiques de deux des principaux systèmes aquifères karstiques de l'île : celui de la source Almiros d'AGIOS NIKOLAOS et celui de la source Almiros d'IRAKLION. Ces deux sources situées sur le littoral septentrional posent en effet de difficiles problèmes aux projets de leur captage en raison de leur contamination par l'eau salée de la Mer Egée.

2 - Rappel des conditions géologiques et structurales

L'île de Crète est caractérisée par une tectonique de nappes qui la rattache au domaine structural de l'arc égéen externe. On trouve de bas en haut (cf. figures 1, 2 et 3) :

- l'autochtone relatif ou zone de l'Ida, comportant plusieurs milliers de mètres de marbres noirs ou gris à silex, en bancs généralement décimétriques à métriques. Cet autochtone est intensément plissé en grands plis déversés vers le sud, et affecté par un léger métamorphisme.

- la nappe tectonique de Gavrovo-Tripolitza, la plus remarquable unité de l'ensemble des zones helléniques. Elle est connue depuis la Yougoslavie jusqu'en Crète et est constituée, du Trias au Lutétien, par des faciès massifs récifaux. Cette nappe est charriée en Crète sur la nappe de l'Ida.

- la nappe du Pinde-Ethia, représentée par des lambeaux épars, généralement très tectonisés et incomplet. Leur charriage sur la série de tripolitza peut localement s'observer dans de bonnes conditions.

- les nappes intermédiaires (nappes d'Arvi, de Miamou et de Vatos), représentées par trois lambeaux d'étendue variable, mais répartis dans toute la Crète moyenne et dont les séries, souvent mal datées, sont sans équivalent certain ni dans les Hellénides ni dans les Taurides.

- la nappe de l'Asteroussia, qui représente la partie supérieure de l'édifice crètois, est constituée de roches métamorphiques (marbres, gneiss et granulites à quartz). Les datations radiométriques de ces roches indiquent pour certaines un âge Crétacé supérieur.

La tectonique de failles qui a découpé la Crète, probablement depuis le début du Miocène, et qui s'est poursuivie au Pliocène et au Quaternaire, a isolé des fossés où les nappes furent préservées de l'érosion. En Crète cependant, les trois unités charriées sur la série de Tripolitza (la nappe de l'Asteroussia, l'unité d'Arvi et la nappe du Pinde-Ethia) sont en général bien érodées et représentées par des klippes éparses dans toute l'île.

Les figures 1, 2 et 3 précisent la répartition des nappes helléniques et des phénomènes tectoniques (d'après M. BONNEAU).
La paléogéographie des hellénides présente dans l'arc égéen externe une évolution caractéristique : la zone ionienne, au métamorphisme près, reste semblable à elle-même ; les autres zones évoluent d'une façon plus ou moins importante, certaines disparaissant.

3 - Constitution et karstification des aquifères carbonatés

Les unités hydrogéologiques dans l'île de Crète sont difficiles à reconnaître à cause de la complexité géotectonique sur l'ensemble de l'île. D'une matière générale on distingue surtout, d'une part les formations épimétamorphiques de l'autochtone relatif, et d'autre part la série carbonatée de la zone Gavrovo-Tripolitza : toutes deux représentent les plus importants aquifères de l'île vu la grande extension de ces ensembles carbonatés (cf. figure 1).

- en ce qui concerne les formations épimétamorphiques, une observation détaillée permet de distinguer (cf. figure 3) :
  . environ 350 mètres de calcaires en plaquettes, marbres noirs ou gris sombre, en bancs épais de un mètre en moyenne, avec des lits de silice blanche, surtout dans la partie moyenne de la série, et, vers le sommet, des intercalations de calcshistes rouges et verts.
  . des couches de passage au flysch, surmontées par un flysch métamorphique tertiaire.
  . des phyllades constituées essentiellement de schistes et quartzites dont l'âge triasique est prouvé paléontologiquement. Ces phyllades sont charriées avec la masse calcaire de la série de Tripolitza sur l'ensemble des calcaires en plaquettes.

- en ce qui concerne la série carbonatée de Gravrovo-Tripolitza, elle est constituée, au dessus de dolomies rubanées et calcaires du Trias et du Lias, par des calcaires massifs sombres et des faciès récifaux dont l'épaisseur est d'environ 300 à 400 mètres.

Dans l'un et l'autre de ces ensembles, la tectonique et la néotectonique ont nettement contribué au développement du karst, surtout dans la partie septentrionale de l'île. Il semble que c'est à la fin du Secondaire et au Plio-Quaternaire que les masses calcaires ont subi leur plus intense karstification. Durant le Quaternaire en particulier, les mouvements épirogéniques ont été très importants : l'élévation générale du niveau de la Méditerranée, commencée fin Pliocène - début du Quaternaire, s'est prolongée jusqu'au Moustérien, provoquant l'inondation partielle et le dépôt des sédiments du Calabrien, du Sicilien et du Tyrrhénien. Dès la fin du Pléistocène se sont manifestées les grades immersions attribuées aux failles et dislocations néotectoniques pendant lesquelles la Méditerranée orientale a pris l'aspect général qu'elle présente aujourd'hui.

4 - Complexité de la qualité des eaux kartisques

En fonction des mouvements néotectoniques et des oscillations du niveau de la mer, les eaux circulant dans les calcaires ont plus ou moins organisé leur réseaux d'écoulement en profondeur par élargissement des fissures et des diaclases de la roche : c'est ainsi que la karstification s'est développée jusqu'en dessous du niveau actuel de la mer, donnant naissance aux nappes karstiques littorales et conduisant à une contamination variable des eaux souterraines douces par l'eau de mer (karst submergé). Ainsi, sur la côte nord, qui présente le niveau actuel principal de drainage, plusieurs sources de très grand débit se trouvent très affectées par l'invasion d'eau salée : les exemples les plus typiques se trouvent dans les régions d'Agios Nikolaos, d'Iraklion et de Georgoupolis. Bien que la plupart de ces sources littorales se révèlent encore inutilisables notamment pendant l'été, leur intérêt économique ne saurait être sous-estimé.

Ce n'est heureusement pas que dans les parties contiguës à la mer que les aquifères karstiques sont développés en Crète. Il y a ainsi dans certaines vallées et dépressions surcreusées dans les calcaires et aujourd'hui en partie comblées par des alluvions et des dépôts récents, de vastes zones dont le volume de leur réserve souterraine en eau douce pourrait largement dépasser celui écoulé par leur source (karst barré). C'est notamment le cas de l'importante source du village Spili, au pied de la montagne Kedros (+ 1777 m.), dont le débit atteint 150 l/s en étiage.

5 - Examen de deux cas particuliers de sources littorales

5.1. Le système karstique d'Agios Nikolaos

La structure géologique du bassin versant de la source d'Almiros d'Agios Nikolaos apparaît très complexe à cause de la position centrale des phyllades métamorphiques de la nappe tectonique de Tripolitza (cf. figure 1). Ainsi, à l'est d'Uvala Drassi cette structure se complique : le Jurassique non métamorphique affleure seul au fond de la vallée et les phyllades disparaissent ; ceci semble lié à la présence d'une faille inverse transversale et responsable de l'effondrement de la partie littorale actuellement recouverte par les dépôts du Néogène et du Quaternaire.

La nappe karstique, contenue dans les calcaires et les dolomies du Jurassique et du Crétacé, se déverse en mer à partir de la source d'Almiros : son débit, en période d'étiage, est de 1,6 m3/s ; son débit maximal est de 3 m3/s ; sa salinité varie annuellement entre 2660 mg/l en Cl^- et 3100 mg/l. Ces variations de débit et de salinité sont très faibles et pourraient être attribuées à des conditions lithologiques favorables à une régularisation naturelle de l'écoulement souterrain. Mais elles indiquent aussi que l'invasion par l'eau de mer est bien avancée à l'intérieur de l'aquifère karstique, ce que confirment les observations faites sur plusieurs forages de reconnaissance dans toute la zone de l'arrière-pays de la source, jusqu'à une distance de 5 à 6 km de la côte : dans cette zone la salinité moyenne est en effet de l'ordre de 600 mg/1 en Cl^- (cf. figure 4).

Essai d'explication :
A partir d'un champ de 15 forages, on a pu contrôler simultanément la salinité et la piézométrie et délimiter les parties de l'aquifère constamment salée par rapport à celles qui sont protégées de cette contamination :

- la partie la plus contaminée est la partie centrale et sud de l'aquifère, du forage D2 (585 mg/l en Cl^-) au forage F7 (1500 mg/l), où un karst noyé est bien développé en dessous du niveau actuel de la mer. Dans ce compartiment de l'aquifère on observe un gradient hydraulique très faible, une transmissivité très élevée (de l'ordre de 10^{-2} m2/s) et une forte salinité mais pratiquement constante d'un bout à l'autre de l'année. C'est la région d'écoulement privilégié du bassin versant, qui communique avec les parties montagneuses de Dictea englobant le grand poljé karstique Lassithi. Du point de vue de la possibilité d'une exploitation des eaux douces, cette partie de l'aquifère n'offre pratiquement aucune chance ; et implanter des forages encore plus éloignés de la mer s'avère difficile à cause de conditions topographiques peu favorables.

- au nord du forage D2, tous les forages existants exploitent de l'eau douce. C'est donc une zone dont le contexte géologique est bien différent de celui de la zone sud. En effet, parallèlement au contact des phyllades et des calcaires en plaquettes, et au sud-ouest de celui-ci, existe une autre faille le long de laquelle apparaissent des phyllades. Entre ces deux accidents tectoniques affleurent des terrains du Jurassique et du

Crétacé, composés de calcaires et de dolomies, qui ont subi un faible métamorphisme. Ainsi se trouve délimitée une bande relativement étroite de calcaire, bornée au nord par des phyllades imperméables et disparaissant normalement au sud sous les roches néogènes également imperméables. Ce dispositif permet d'établir, tant au nord qu'au sud, des limites nettes et précises pour cette zone à eau douce. Il est fort probable que les phyllades jouent un rôle de niveau de base local avant que les eaux souterraines ne se déversent dans la partie de l'aquifère karstique submergée par les eaux saumâtres, sous les dépôts du Néogène.

Dans cette zone à eau douce, la transmissivité diminue en général du sud vers le nord (de Uvala Messa Lakonia à Uvala Exo Lakonia). A titre d'exemple, le débit de pompage du forage D2 qui est de 30 l/s, sans rabattement, traduit une transmissivité de l'ordre de 10^{-2} à 10^{-1} m2/s ; par contre, au nord, prés du contact entre les phyllades et les masses de calcaires jurassiques e la zone de Tripolitza, la transmissivité est très variable, de 10^{-3} à 10^{-5} m2/s : ceci est vérifié par le faible débit de la plupart des forages avec des rabattements très élevés (plus de 15 mètres).

Il est donc évident qu'en raison de la position différente des phyllades par rapport aux calcaires, au nord et au sud du bassin versant, la région nord-ouest de la source d'Almiros dont être considérée comme beaucoup moins perméable que la région sud où les calcaires miocènes jouent le rôle de collecteur privilégié de l'écoulement souterrain de la source.

On doit ainsi admettre que les caractères hydrauliques et chimiques spécifiques de la source d'Almiros découlent de la juxtaposition de deux sous-systèmes karstiques bien délimités : l'un caractérisé par une contamination très large des eaux souterraines par l'eau de mer et par les débits notables des forages existants ; et l'autre par l'absence de contamination directe et par les débits réduits des forages.
L'aquifère karstique de la source d'Almiros d'Agios Nikolaos présente donc, tant par sa constitution que par son fonctionnement, un système complexe dont l'étude et la mise en valeur exigent une meilleure connaissance de chacun des deux sous-systèmes situés au nord-ouest et au sud-ouest de la source.

5.2. Le système karstique d'Iraklion

La grande source d'Almiros d'Iraklion se trouve à 8 kms à l'ouest de la ville d'Iraklion, près du rivage de la Mer Egée. L'eau jaillit au pied du massif karstique de Keri qui constitue la partie nord-est du bassin versant de la montagne Psiloritis (+ 2456 m.). Son écoulement en surface (Almiros Potamos) rejoint la mer après 1 km environ de parcours au travers d'une petite plaine alluviale. C'est une source saumâtre dont on a essayé d'abaisser la salinité en relevant le niveau de son exutoire au moyen d'un barrage voûte : cette tentative, réalisée en 1971, s'est traduite par un échec complet qui montre ainsi la complexité des conditions hydrogéologiques et hydrauliques dans un karst littoral.

Du point de vue géologique, le bassin versant de la source est constitué par des calcaires en plaquettes, à l'ouest, et par les calcaires de la série Gavrovo-Tripolitza, à l'est du sommet de la montagne Psiloritis. Entre ces deux séries calcaires on trouve des phyllades dont la position tectonique complique les conditions hydrogéologiques dans l'ensemble du bassin versant : des recherches détaillées ont permis de préciser que les calcaires de la série Gavrovo-Tripolitza pouvaient se trouver superposés soit directement aux phyllades, soit directement aux calcaires en plaquettes (BONNEAU, 1977). Tous ces calcaires, épais et très tectonisés, sont bien kartifiés. A l'est du bassin versant, juste au sud de la source d'Almiros (cf. figures 5 et 6), un fossé tectonique tertiaire, comblé par flysch et marnes, a pour effet de concentrer le drainage souterrain dans une zone calcaire d'extension limitée au voisinage de la source (le massif de Keri), tandis que plus à l'ouest les limites correspondent à l'extension des formations carbonatés de la nappe tectonique de Gavrovo-Tripolitza (MONOPOLIS et al., 1969 ; BREZNIK, 1973).

La superficie de l'ensemble de ce bassin versant est estimée à 300 km2.
Les précipitations annuelles sont de l'ordre de 600 mm dans les parties basses, mais de l'ordre de 1400 mm en dessus de 1500 m d'altitude.

Compte tenu de son débit, la source d'Almiros d'Iraklion représente le plus important point d'eau de l'île de Crète :

- le débit moyen est d'environ 8 m3/s
- le débit minimal est voisin de 4 m3/s
- le débit maximal pourrait atteindre en grande crue près de 30 m3/s.

La teneur en chlorures varie en fonction du débit, mais dans le sens inverse :

- pendant les périodes de crue, l'eau de la source est presque douce, de 35 à 300 mg/l en Cl^-
- pendant la période d'étiage, la teneur en chlorures s'élève plus ou moins graduellement, et, à la fin de cette période, elle peut atteindre 5000 à 6000 mg/l en Cl^- (BURDON et al., 1964).

Le diagramme synoptique de la source représenté dans la figure 7, montre bien la gamme des variations de débit et de salinité de l'eau en fonction du régime des pluies sur l'ensemble du bassin versant.

Essai d'explication :
Dans une première approche, l'analyse de l'hydrogramme de la source parait indiquer que la contamination des eaux souterraines par l'eau de mer est bien limitée autour de la source, c'est-à-dire que le front de l'eau salée n'avance pas trop dans l'aquifère karstique de Keri. Mais lorsque on analyse les données de crue et celles d'étiage au niveau de ce système karstique, on peut constater qu'il existe un débit critique vers 10 à 11 m3/s, en dessous duquel l'influence de l'eau salée devient très importante. Ainsi, pour un débit de 12 m3/s, la teneur en Cl^- est de 100 mg/l ; pour 11 m3/s elle est déjà de 300 mg/1 ; en dessous de 10 m3/s la salinité s'élève rapidement pour atteindre une valeur moyenne de 5000 mg/1 pour un débit minimal de 4 m3/s (cf. figure 7).

On peut interpréter ces constatations de la façon suivante :

. la kartification, de même que le mode particulier de circulation des eaux dans l'aquifère de la source, sont dus aux mouvements tectoniques et néotectoniques, spécialement dans les parties des terrains karstiques contiguës à la mer, c'est-à-dire dans le massif calcaire de keri où la karstification se développe très en dessous du niveau de la mer (cf. figure 8). La situation de ce massif et l'existence du fossé tectonique tertiaire à l'est de la source, permettent d'envisager la possibilité d'un karst sous-marin dont les collecteurs doivent être largement interconnectés, atteignant une profondeur de 300 à 400 mètres au dessous du niveau de la mer : certains chenaux isolés pourraient même être développés à une profondeur de 600 à 1200 mètres au-dessous de ce niveau (BREZNIK, 1978). En conséquence, l'eau qui s'écoule par la source est saumâtre à cause du mélange direct avec l'eau de mer à l'intérieur de l'aquifère (convection mécanique). Les accidents tectoniques, bien visibles autour et en amont de la source (faille et fossé néogène), et les différentes formes karstiques superficielles (surtout les dolines) développées à la faveur des failles, conduisent à considérer le massif karstique de Keri comme le compartiment privilégié des écoulements souterrains d'eau douce et d'eau salée vers la source.

. il est très probable, sinon certain, que la plupart des conduits profonds du massif karstique de Keri se trouvent contaminés par les eaux salées (ceci a été mis en évidence par l'investigation chimique des eaux souterraines dans les forages de reconnaissance en période d'étiage). Mais il n'est pas encore possible de déterminer par quel passage se produit l'invasion marine dans la formation calcaire au-dessous du niveau de la mer : se produit-elle dans le massif de Keri lui-même... ou à partir du fossé tectonique en-dessous du dépôt néogène ? Il pourrait aussi se faire qu'une partie au moins de cette contamination provienne d'eaux salées fossiles emmagasinées dans l'aquifère au cours des périodes du Quaternaire ? Il faut bien souligner que les conditions paléogéographiques diverses et complexes qu'a connu cette région littorale permettent d'envisager beaucoup d'hypothèse et que l'on aura du mal à trancher en raison des difficultés bien évidentes d'accès au milieu à étudier en vue d'y collecter les informations nécessaires.

. l'absence de marnes imperméables -qui peuvent constituer ailleurs un écran permettant la conservation d'une nappe d'eau douce jusqu'au niveau de ses exutoires littoraux et parfois sous-marins- a provoqué un envahissement total par l'eau de mer de la partie aval du bassin versant de la source (aquifère de Keri) où le karst est développé assez profondément sous le niveau de la mer. Dans ces conditions il parait clair qu'en période d'étiage, lorsque la charge en amont de la source est trop faible pour refouler le biseau salé (ce qui correspond à un débit d'environ 5 m3/s) il se produit une convection mécanique entre l'eau douce et l'eau salée à cause de la contre-charge de l'eau de mer qui peut dépasser légèrement une hauteur de 10 mètres sur le site de la source suivant la relation $h_s = h_m \times d_s/d$ (avec h_s = contre-charge de l'eau salée à la source en mètres ; h_m = 400 mètres ; d_s = 1,03 gr/cm3 ; d = 1 gr /cm3) (cf. figure 8). A cet égard, les travaux de relèvement effectués en 1971 sur l'exutoire de la source ne pouvaient pas donner de résultats satisfaisants, le phénomène de mélange entre les eaux douces et les eaux salées étant impossible à éliminer sur le site de la source d'Almiros.

Solution envisagée :

La situation privilégiée de la structure du bassin versant qui limite la possibilité d'invasion d'eau salée à sa seule partie aval (massif de Keri) laisse un certain espoir de pouvoir exploiter en période d'étiage un important débit d'eau douce.

On peut envisager pour y parvenir de tenter d'abord de déterminer les limites de l'invasion d'eau salée en période d'étiage, puis de rechercher les points où les caractéristiques de l'écoulement souterrain et sa charge hydraulique pourraient convenir pour l'exécution d'un captage rationnel. A priori il semble difficile d'extraire de grandes quantités d'eau douce dans cet aquifère par forages d'exploitation ou puits de grand diamètre en raison de la dispersion des voies d'écoulement privilégiées liées aux accidents tectoniques et aux diaclases. Il nous paraît plus évident d'envisager un captage par galerie drainante en vue d'intercepter le plus grand nombre possible de chenaux, multiples et complexes dans l'aquifère de Keri.

Outre l'importance du débit assuré par ce type d'ouvrage, la qualité de l'eau devrait se trouver bonne, le niveau dynamique de la nappe karstique en exploitation étant forcément contrôlé par le niveau de la galerie (en dessus du niveau de la mer), ce qui minimise la possibilité de contamination de l'eau douce par l'eau salée.

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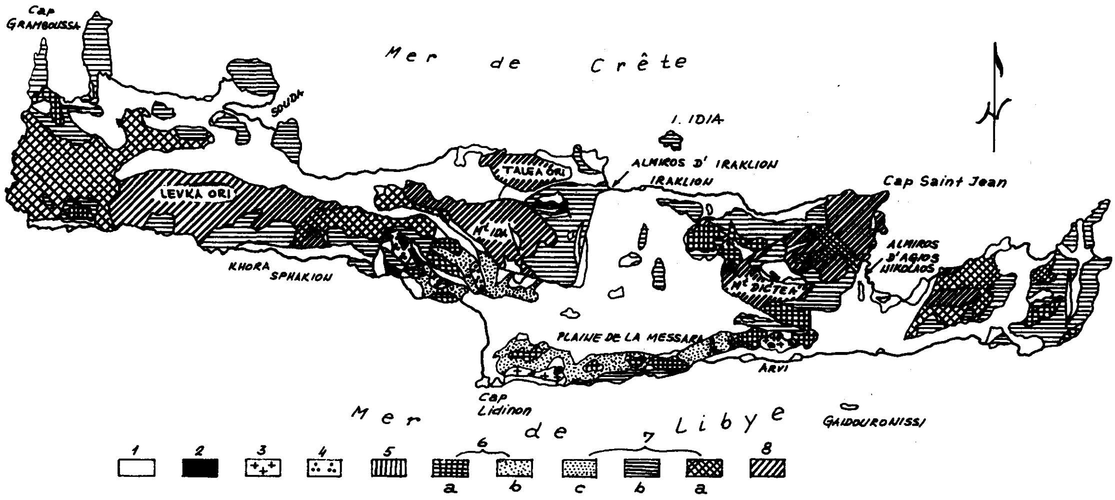


Fig. 1 - Carte géologique et structurale schématique de l'île de Crête (d'après M. BONNEAU, 1976)

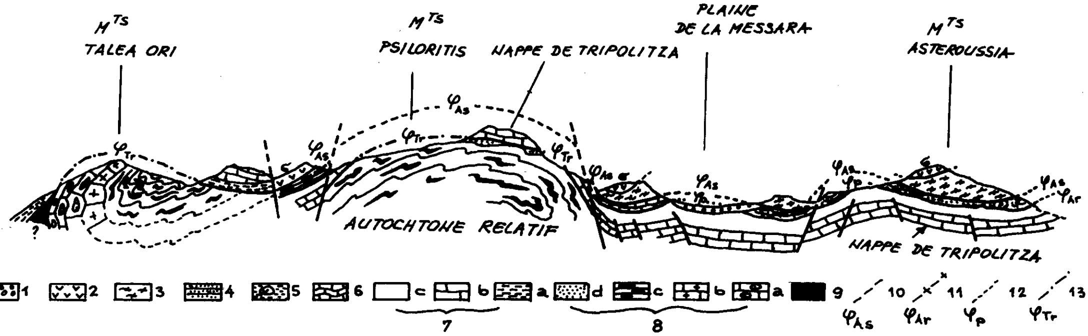
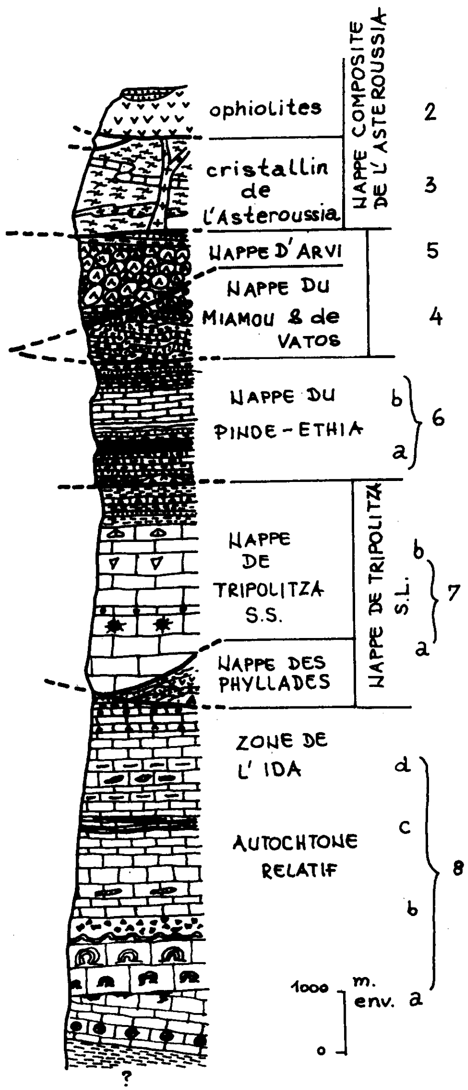


Fig. 2 - Coupe schématique d'ensemble de la Crète moyenne (d'après M. BONNEAU, 1973)



LEGENDE DES FIGURES

1 : Néogène et Quaternaire,
2 : Ophiolites,
3 : Nappe de l'Asteroussia,
4 : Nappe de Miamou (flysch + mélange),
5 : Nappe d'Arvi,
6 : Nappe de Pinde-Ethia
  a : calcaires et radiolarites : Trias-Eocène inf.
  b : flysch-Eocène)
7 : Nappe de Tripolitza
  a : phyllades
  b : calcaires : Trias ? -Lutétien
  c : flysch)
8 : Autochtone relatif (zone de l'Ida)
  a : calcaires à fusulines : Permien
  b : dolomies
  c : calcaires en plaquettes
  d : couches de Kalavros : Oligocène inférieur)
9 : Gypses
10 à 13 : Contacts des nappes tectoniques.

Fig. 3 - Diagramme structural et stratigraphique des unités tectoniques de la Crète.

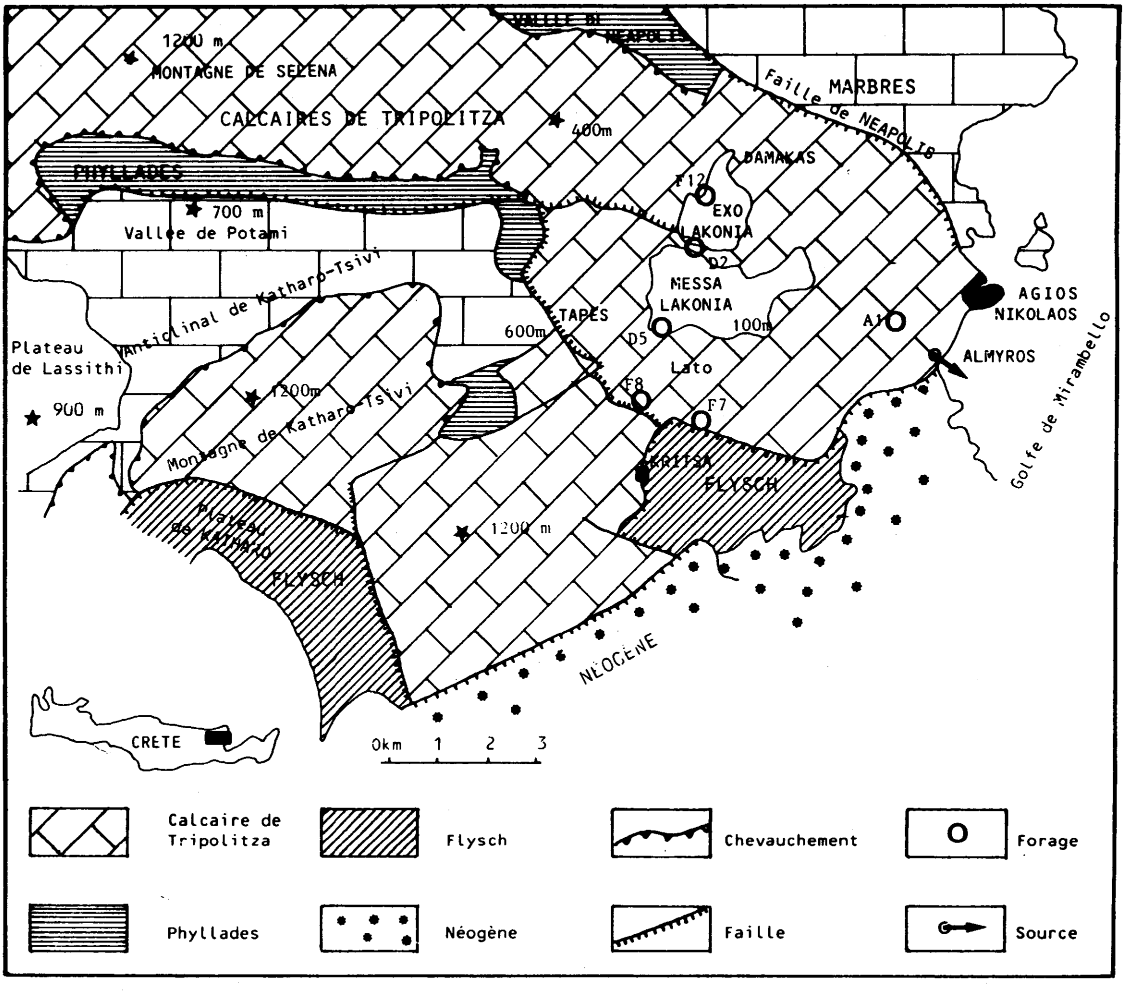


Fig. 4 - Carte géologique du système d'Almyros Agios Nikalaos (d'après C. BEZES et C. JOSEPH, 1983)

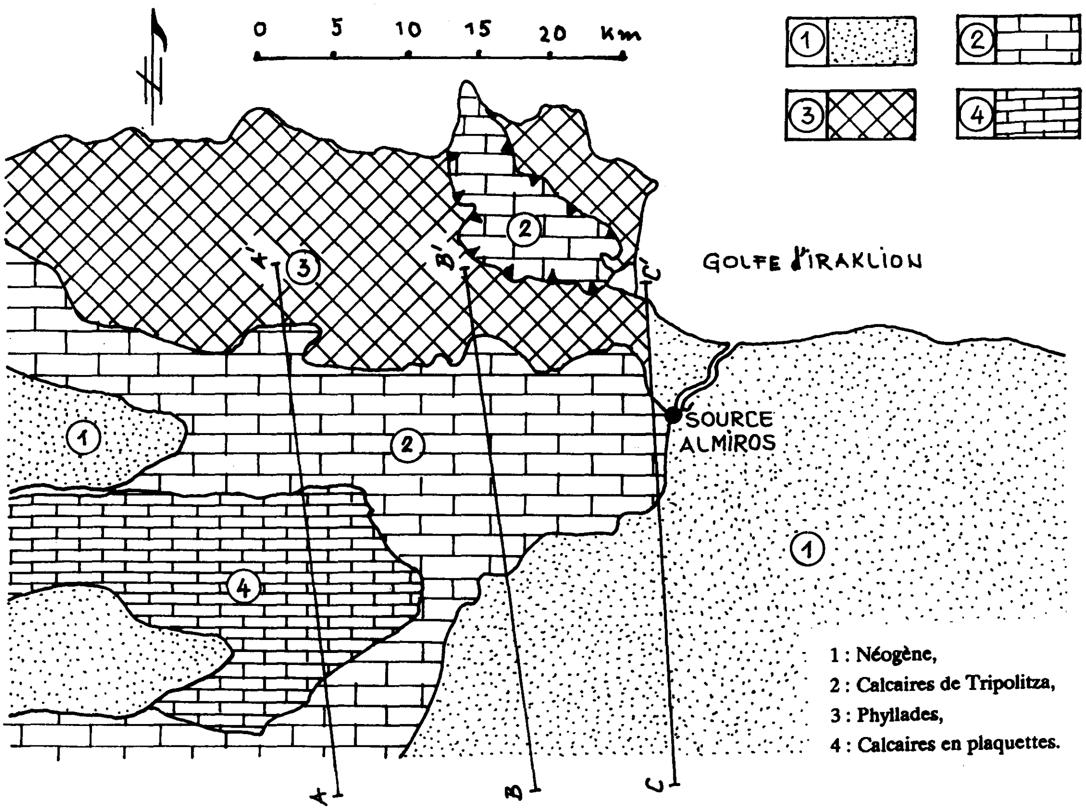


Fig. 5 - Carte géologique et structurale schématique des Talea-Ori (d'après D. MONOPOLIS et K. MASTOPIS, 1969)

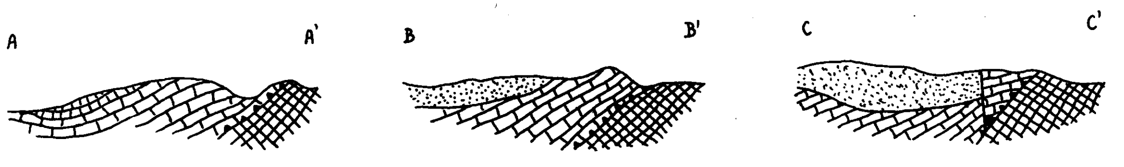


Fig. 6 - Coupes schématiques d'ensemble de la région karstique à l'ouest de la plaine d'Iraklion (d'après D. MONOPOLIS et K. MASTOPIS, 1969)

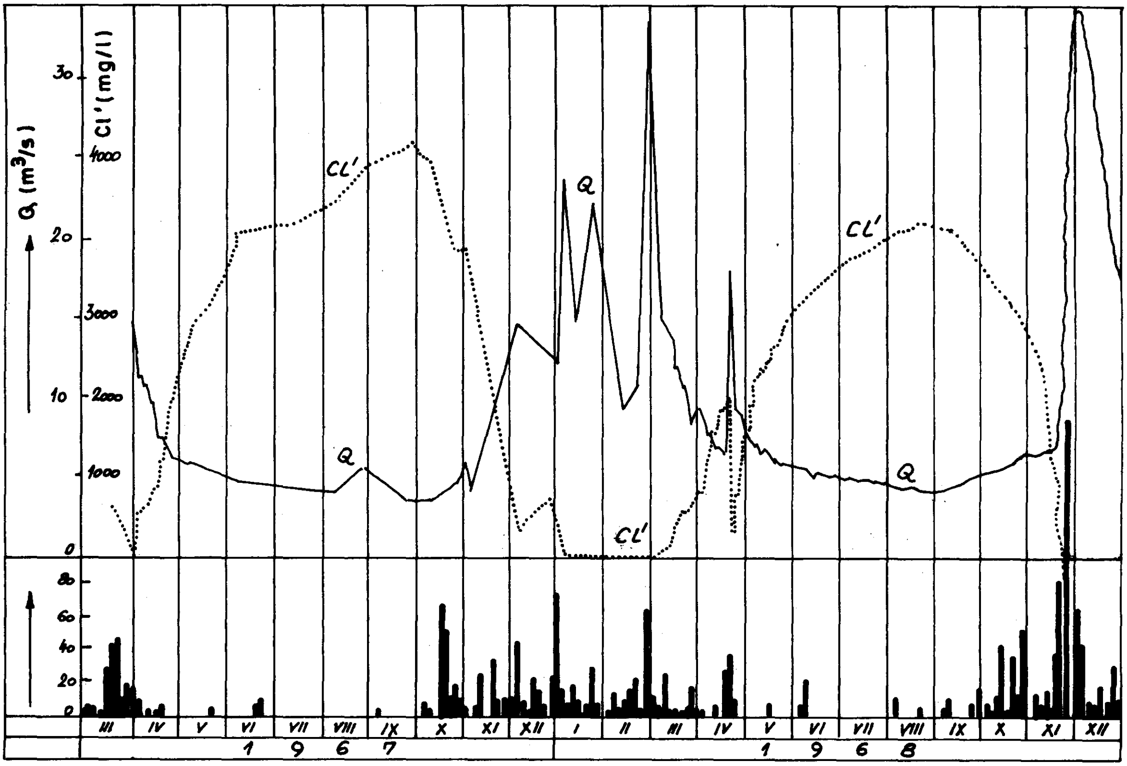


Fig. 7 - Diagramme synoptique de la source Almiros d'Iraklion : variations du débit et de la salinité de l'eau en fonction de la pluie (d'après D. MONOPOLIS et K. MASTOPIS, 1969)

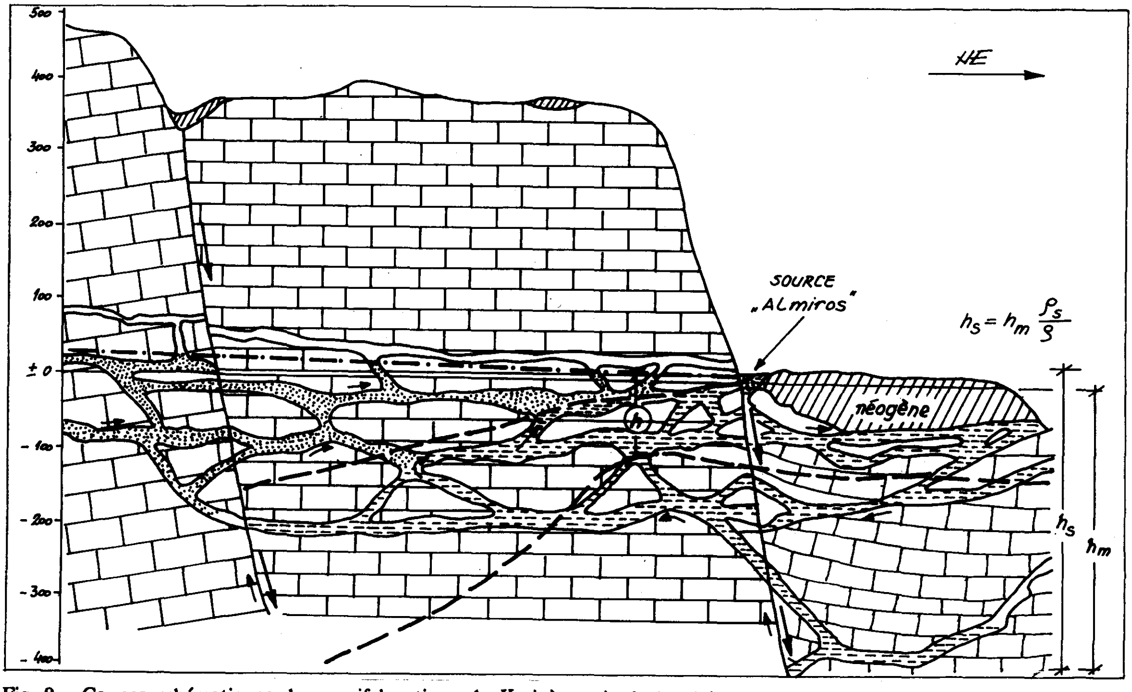


Fig. 8 - Coupes schématiques du massif karstique de Keri à partir de la plaine d'Iraklion et de la source d'Almiros (d'après D. MONOPOLIS et K. MASTOPIS, 1969)

TECTONIC INFLUENCES ON GROUNDWATER FLOW SYSTEMS IN KARST OF THE SOUTHWEST TAURUS MOUNTAINS, TURKEY

WILLIAM BACK
U.S. Geological Survey
Water Resources Division
431 National Center
Reston, Virginia 22092
USA

and

GULTEKIN GUNAY
Institute Karst Water Resources Research Center
Hacettepe University 06532
Beytepe, Ankara
Turkey

The Karst Commission first met in Turkey in July, 1985, during the International Symposium, "Karst Water Resources", which convened in Ankara for formal presentation of papers and followed with a field trip from Ankara through Konya and Antalya to Izmir. The Proceedings volume was published by the International Association of Hydrological Sciences (Gunay and Johnson, 1985) and a guidebook was prepared for the field trip. The latest meeting of the Karst Commission was held in connection with the "International Symposium and Field Seminar on Hydrogeologic Processes in Karst Terrains" in Antalya, October 7-17, 1990. Also during that period was the first organizational meeting of the International Geologic Correlation Program (IGCP) Project 299 under the leadership of Yuan Daoxian of the Karst Research Institute, Guilin, China. This chapter is excerpted largely from the guidebooks prepared for those Symposia and from the official report of the IGCP Project 299 (Yuan and Back, 1991).

About one third of Turkey is underlain by carbonate rocks that have been subdivided into four karst regions of which the Taurus Mountains are the most extensive (Eroskay and Gunay, 1979). They begin at the Turkish coast of the Aegean and Mediterranean Seas and extend eastward to the international boundary. The carbonate rock units are about 200 km wide along the Taurus Mountains which attain elevations of 2,500 m. These high mountain ranges, sharp peaks, deep valleys, and narrow gorges cause extremely rugged topography (Fig. 1).

Karst features of western Turkey bordering the Aegean and Mediterranean Seas dramatically demonstrate the tectonic, lithologic, and climatic controls on the occurrence, movement, and chemical character of ground water. The Taurus Mountains were formed by folding and overthrust faulting during the Alpine orogeny. This orogenic system also contains the karstified mountains of Greece and Yugoslavia, consequently, the western extension of the Taurus Mountains is submerged beneath the Mediterranean. The Aegean Islands have been connected among themselves; and Turkey and Greece have been interconnected at various

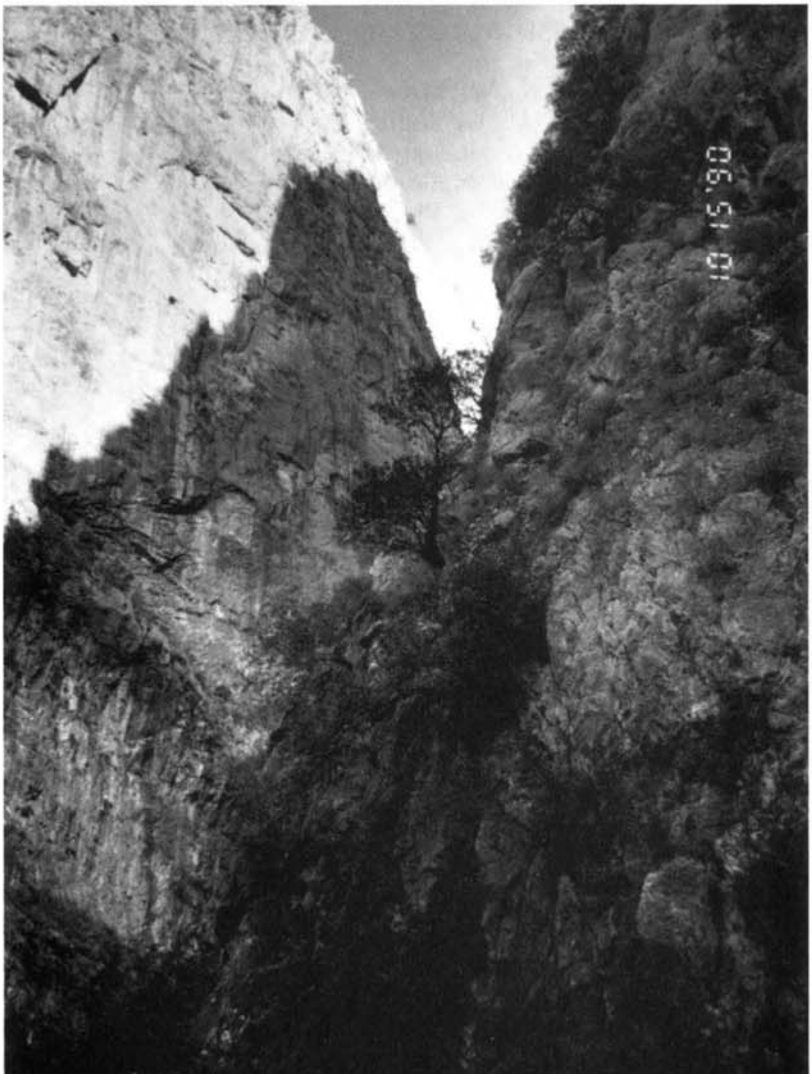


Figure 1. Many deep canyons such as Gökcesu are formed by dissolution of the limestone along faults and fractures.

times. Since the late Tertiary, western Turkey has experienced downward movement that has been largely controlled by faults. The highly crinulated and dissected Aegean coastline of Turkey is due to this tectonic subsidence (Fig. 2).

The epirogenic coastal movements continued from late Tertiary into the Quaternary period. The Quaternary tectonism is characterized by normal faults and, in this western part, by block faults. Many of the coastal bays are in structural grabens that extend inland as major valleys. At approximately the middle of the early Pleistocene epoch, structural activity changed the river systems. In the late Tertiary, the river drainage from the highlands was directed toward the north and the south. The early Pleistocene faulting diverted these rivers into a westerly direction. With uplift, karstification increased and the surface drainage of many lakes and rivers was transformed into subsurface drainage.

During the Ice Age the climate changed drastically several times. Although the Taurus Mountains are now free of ice, they were covered with glaciers during that period. In all of Turkey, the snowline during the last glacial epoch was 1,000 to 1,200 m lower than at present. Thus, the temperature in July would have been 6 to 8° C lower than at present. Those regions that were below the snowline were subjected to the effects of the peri-glacial climate during the Ice Ages. The climatic fluctuations also had effects in the lower elevations. During the glacial epochs the lakes expanded due to reduced evaporation. For example, the level of Lake Burdur (near Antalya) rose by 90 m, Lake Tuz (near Ankara) by 110 m. East of Konya, a lake that was 125 by 25 km has disappeared entirely. These climatic changes were controls on the extensive dissolution of limestone and precipitation of the travertine terraces of Antalya.

At the end of the last Ice Age, approximately 10,000 years ago, the temperature rose rapidly to its present value. As evaporation increased, ground-water levels declined and many springs ceased to flow. Freshwater lakes either turned into saline lakes or dried up entirely. The climate did not change much during the Holocene epoch, with the exception of the period of 5,500 - 2,500 BC, when it was somewhat warmer and more humid (Brinkmann, 1976). At the beginning of the Holocene, the people of Turkey changed from a hunting and gathering society to a planned agricultural society with farming and ranching. Even though the tectonic subsidence is continuing today, new land is being formed by deposition of sediments in many of the estuaries. The amount of sediments carried by the rivers has increased during the last millennia as a result of deforestation and agricultural practices. For example, the Menderes River (Fig. 3) near Izmir has an average delta advance of about 6 m/a (Brinkmann, 1976).

Much of the limestone, which ranges in age from Mesozoic to Holocene, has either been overthrust or deposited on formations with extremely low permeability. The combination of this extensive impermeable lower boundary of the regional karst aquifers along with the numerous faults associated with the tectonism are the major controls on the karst hydrogeology of this region. The Taurus Mountain region is characterized by abundant water resources, large hydroelectric potential, some of the world's largest karst aquifers, and the largest karst springs. These

Figure 2. View of the crinulated shore line forming many bays where the tectonic structures forming the Taurus Mountains plunge into the Mediterranean Sea.

Figure 3. View of the heavily braded Menderes River. The sediments are derived from erosion caused in part by agricultural activities during the past few centuries. The Menderes River (earlier spelling Maiandros) is the origin of the term "meander."

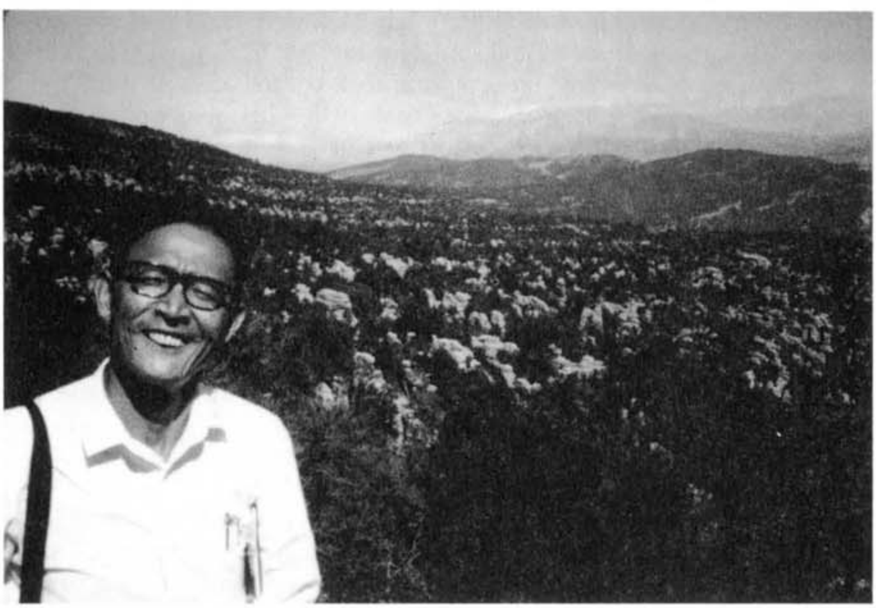


Figure 4. View of the deeply disserted karst formed in conglomerate. Daoxian Yuan, Leader for IGCP Project 299.

springs of exceedingly large discharge are formed by the channelized flow along the fractures by water that cannot penetrate into the deeper formations because of their low permeability; many of the large springs occur at this contact. The original heterogeneity of the aquifer formed by the tectonism has been increased by dissolution of the limestone. This dissolution results from infiltration of the great amount of rain and snow that occur in the Taurus Mountains, particularly at the higher elevations. If the transmissivity of the carbonate aquifers were more homogeneous, and the aquifers were hydrologically connected to the underlying sediments, storativity of the entire system would greatly increase, and diffuse flow would occur.

This area is characterized by abundant dolines, large poljes, coastal and submarine springs, sea caves, and large travertine terraces. One of the more unusual types of karst is that formed in an extensive conglomerate. (Fig. 4) The conglomerate is composed of cobbles of limestone with calcareous cement. The mode of transportation and deposition is not well understood. Participants on the field seminars had the opportunity to observe many of these features. Some of the major conclusions of the participants can be summarized as follows (Yuan and Back, 1991): (1) The formation of karst in southwestern Turkey is strongly influenced by the alternation of Jurassic and Cretaceous limestones with ophiolites and Tertiary flysch in both the horizontal and vertical planes, resulting from nappe structures. Many large springs, some of which are submarine, discharge through structural windows where carbonate rocks are exposed within ophiolite. 2) The general characteristics of this region reflect a combination of both humid and arid environments, a combination that does not occur in tropical climates. This unusual combination in Turkey results from the Mediterranean climate which is characterized by dry summers, wet winters, and a long period of snow cover in the mountains. Therefore, mechanical weathering is as important as chemical corrosion. 3) Because of lower water temperatures, coastal karst of this area displays effects of biogenic processes less than does that of the Caribbean or of the southeast Asia region. 4) The hydrogeologic conditions of the karst indicate two types of recharge: a) systems that are recharged by poljes on lower plateaus are characterized by high fluctuations of discharge, higher temperature, higher bicarbonate content, and extensive travertine deposition; b) systems recharged mainly from high mountains that are covered by snow for long periods are characterized by more stable regimes of discharge, lower temperatures, lower content of bicarbonate, and little or no travertine deposition.

The Antalya travertine deposits have formed as a result of the precipitation of carbonate minerals due to outgassing of carbon dioxide from water initially saturated with respect to the carbonate minerals (Herman and Hubbard, 1990 and this volume). The source of the calcium and carbonate ions is the limestone of the Taurus Mountains to the north of the plain. The travertine in this plateau has a thickness of about 300 m and extends over an area of approximately 630 km^2. These deposits exist as three separate terraces; one at the altitude of ~300 m, another between 50 and 150 m, and the third terrace is below sea level.

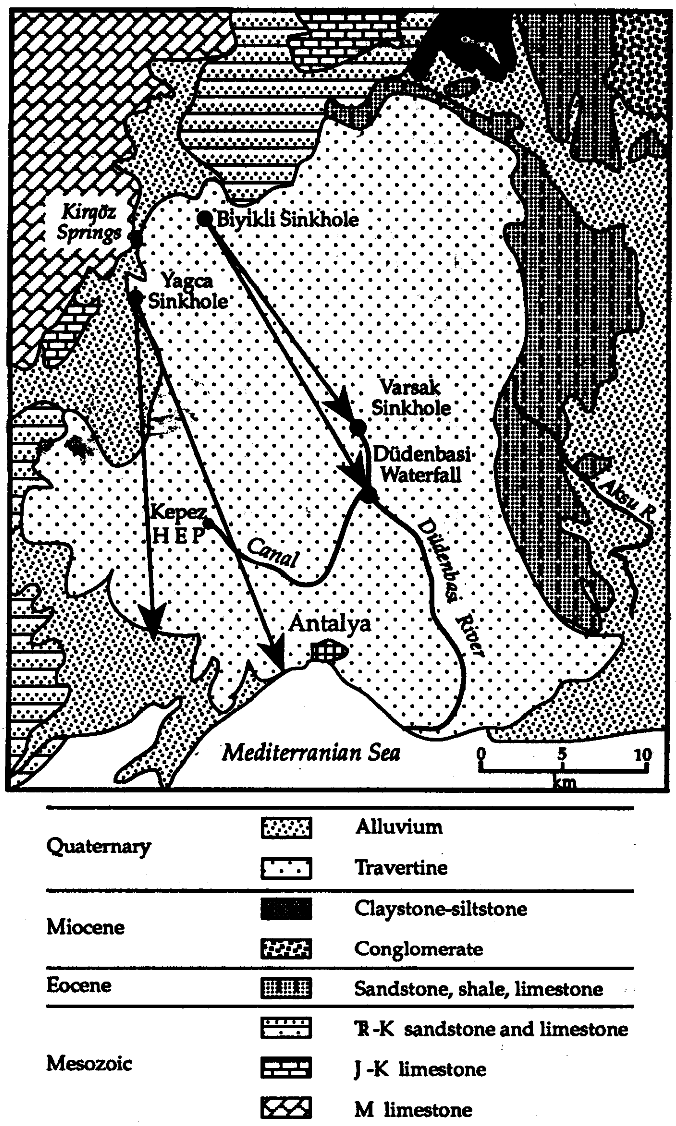


Figure 5. Map showing geologic control on the flow of surface and ground water in travertine terraces of Antalya.

A spectacular waterfall (Dudenbasi) on these terrace deposits has an unusual hydrogeologic occurrence. The water travels to the waterfall through three different pathways, even though the original source is one major group of springs (Fig. 5). These springs (Kirkgoz) are about 25 km from the waterfall and part of the water travels about 8 km and is lost into the Yagca sinkholes. Even though hydrologic connection exists between Kirkgoz Springs and Biyikli sinkhole, the sinkhole receives water only during high flood. Part of the water from Biyikli and Yagca sinkholes travels directly to the Dudenbasi waterfall; part of it discharges into the Varsak doline where the water enters on the upgradient side of the doline and discharges downgradient to continue on to Dudenbasi. The third source of water is from a canal that carries water from Kirkgoz springs to the Kepez power plant. After the water is used for production of electricity, it is diverted into other canals and reused for irrigation. The returned irrigation water is transported by canal to Dudenbasi waterfall. The water cascades over a cliff about 15 m high where an extensive cave system has developed. It is possible to walk in the open cave system in back of and beneath the waterfall. The cascading water forms the Dudenbasi River, which makes another waterfall where it enters the Mediterranean. The Dudenbasi waterfalls are a major tourist attraction in the Antalya area.

Another major tourist attraction in the area of the field trip of hydrogeologic significance is the spectacular travertine deposit of Pamukkale (Fig. 6 and 7). Pamukkale means "cotton castle" and is on the ancient site of Hierapolis which was a major city even before the Romans came. The travertine terraces are deposited from the series of geothermal springs that are brought to the surface along faults bordering the graben that forms the valley. The area has some potential for geothermal energy.

The source of the major thermal spring for the travertine terraces and ponds is a cave that had long been known as a road to Hades. The water from the thermal spring is 35°C, and originally the spring was used for the medicinal qualities of the water. The atmosphere in the cave could not support life, probably due to the high concentration of carbon dioxide gas and consequent lack of oxygen. The reputation of the city was further enhanced by the miracles that the priests would perform by using the cave. It was well known that the cave was dangerous, yet the priests could stay in for long periods of time, presumably by following the pathways where the upper chambers contained adequate oxygen, then emerge with no adverse reactions. That would then give them the aura of immortality and they could then exert their will on the people. In the second century AD the Romans diverted the warm mineral water into the buildings for the baths. The Roman bath house is now a museum and the thermal springs have been made available to tourists through large swimming pools associated with a motel.

Figure 6. Travertine terraces of Pammukale, with ruins of the ancient city of Hierapolis and the Roman Baths in the background.

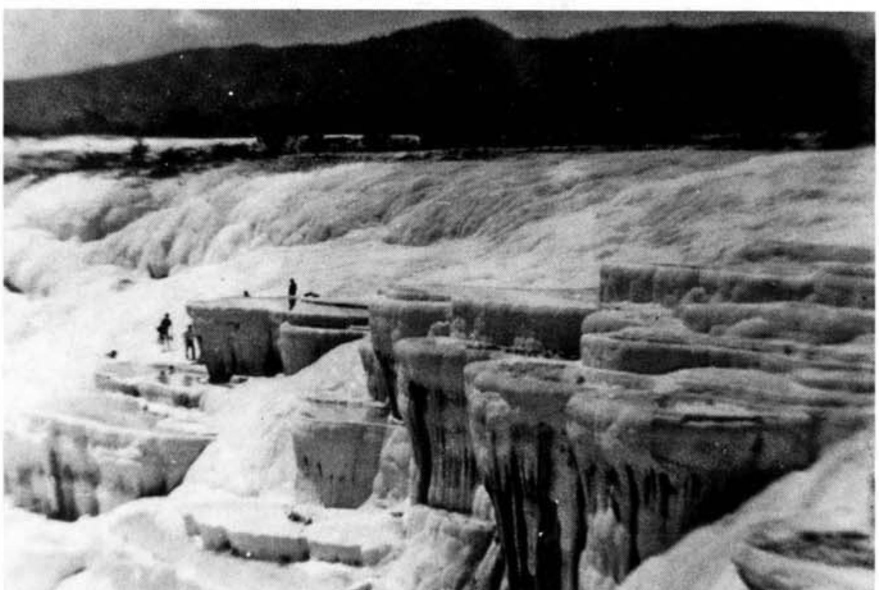


Figure 7. Close-up of the Pammukale terraces containing rimmed ponds of warm water.

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HYDROGEOLOGY OF KARST IN WEST GEORGIA

GURAM J. BUACHIDZE
and
TAMAZ Z. KIKNADZE
Vakhushti Bagrationi Institute of Geography
Academy of Sciences of the Georgian SSR
1, Z. Rukhadze Str.
380093 Tbilisi
USSR

INTRODUCTION

The economic progress of Georgia requires thorough hydrogeologic study of the areas underlain by soluble rocks. Construction of hydroelectric power stations, tunnels, railways, and roads, erection of industrial and civil structures, and development of mineral deposits within these regions are impossible without careful investigation of karst aquifers. Development of cave streams and conduit springs for energy is possible without requiring the construction of expensive dams. Water from karst aquifers is used for water supply, irrigation, and development of profitable fish culture. Oil and gas fields and mineral and thermal water resources are common in karst reservoirs. In studying karst, special consideration is given to dry and flooded caves, which may be exploited both for medical purposes (speleotherapy) and excursions.

Characteristics of the hydrogeology of karst regions, however, are poorly reviewed in the Soviet scientific literature, and, consequently, there is no single opinion as to the nature of karst water and its position within the available classification schemes for underground water. The principle difference of karst water from other types of underground water is not revealed completely. Characteristics are observed in the formation and dynamics of karst water and in the hydrogeological regime, chemical properties, temperature, and utilization. These distinctions, along with the methodical system of their investigation and the increasing demands for water, emphasize the need to study karst aquifers and stimulate a branch of hydrogeological science - that is "Hydrogeology of Karst."

A retrospective look at the history of the karstological science elucidates the struggle of two alternative hydrogeological hypotheses: that of A. Grund concerning single-level karst water in karst massifs and that of Martel-Knebel-Katzer regarding isolated karst streams. These hypotheses have their supporters even now, though recent investigations proved that under various hydrodynamical conditions both hypotheses may co-exist even within the interior part of the same karst mass (Dublyanski and Kiknadze, 1984).

The most beneficial area for demonstrating general hydrogeological characteristics of karst involves mountain karst regions of the Alpine folded belt; research in these regions stimulated the origin of karstology and speleology. The Great Caucasus presents a most interesting region in the above megaregion.

Therein several dozen large and minor karst masses can be regarded as standard and representative for study and establishment of general characteristics of processes of karst genesis. The past decades were important for fragmental and incidental investigation of karst within the above region, and the most recent 15-20 years are notable for the appearance of prominent special monographs dealing with the geology and hydrogeology of karst (Kiknadze, 1972, 1979), karst morphology (Tintilozov, 1976), and karst hydrology (Gigineishvili, 1979).

The total area of Georgia is about 70,000 km^2. The karst occupies nearly 11,000 km^2, or 16% of the total area of the republic, with the share of carbonate rocks of all varieties accounting for 10,200 km^2, that is, 93% of karst area. Suffusional karst occupies nearly 200 km^2 and pseudo-karst more than 550 km^2 (Kipiani, 1974). Carbonate karst includes the lithologically different karst types:

1. karst developed in limestone, dolomite, and dolomitic limestone of Upper Juarassic, Cretaceous, and Lower Paleogene, considered the true karst,
2. karst in calcareous flysch of Upper Jurassic, Cretaceous, and Paleogene, the flysch type, and
3. karst in calcareous conglomerates of Mio-Pliocene and Quaternary stages, the clastokarstic type (Kiknadze, 1983).

The hydrogeological structure and effect of different geographical-geological factors upon the development of the first type of karst is as follows. This type of karst in West Georgia, within the Great Caucasus Southern Slope, comprises an area 350 km long and 5 to 35 km wide, that is, an area of about 5,000 km^2 (Fig. 1). Two karst types are distiguishable in West Georgia, by their geotectonical structure and geological evolution: 1) the geosynclinal type of karst, developed on the Great Caucasus Southern Slope, and 2) the platform type, developed in the Georgian Block or the intermontane trough (Kiknadze, 1979, 1983).

LITHOLOGY AND HYDROGEOLOGY OF KARST

Lithological composition influences a most significant part of formation of hydrogeological and hydrochemical conditions of karst development. Several systems of aquifers and aquitards are developed within West Georgia (Kiknadze, 1983; Fig. 2). They include:

1. aquifers of the Quaternary deposits,
2. aquitards of the Upper Paleogene-Neogene, the upper, nappe water-confining layer,
3. karst aquifers of carbonate deposits of Upper Cretaceous, Turonian-Datian stage, Lower Paleogene, and Paleocene-Middle Eocene,
4. aquitard complex of the Lower (Aptian-Albian stage) and Upper (Cenomanian stage) Cretaceous (intermediate water-confining stratum),
5. (a) karst aquifers of carbonate rocks of Lower Cretaceous (Valanginian-Barremian) and (b) Upper Jurassic (Lusitanian-Tithonian stage), and
6. aquitard complex of volcanic-sedimentary deposits of Middle and Upper Jurassic (Bajocian and Callovian-Oxfordian, multi-colored suite of Upper Jurassic), the lower water-confining stratum.

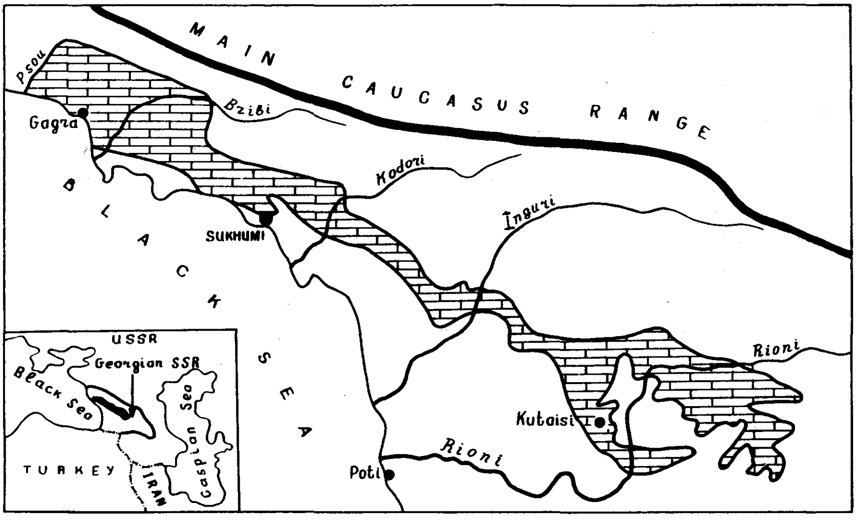

Figure 1. Karst terrain developed in limestone, dolomite, and dolomitic limestone in West Georgia within the Great Caucasus Southern Slope.

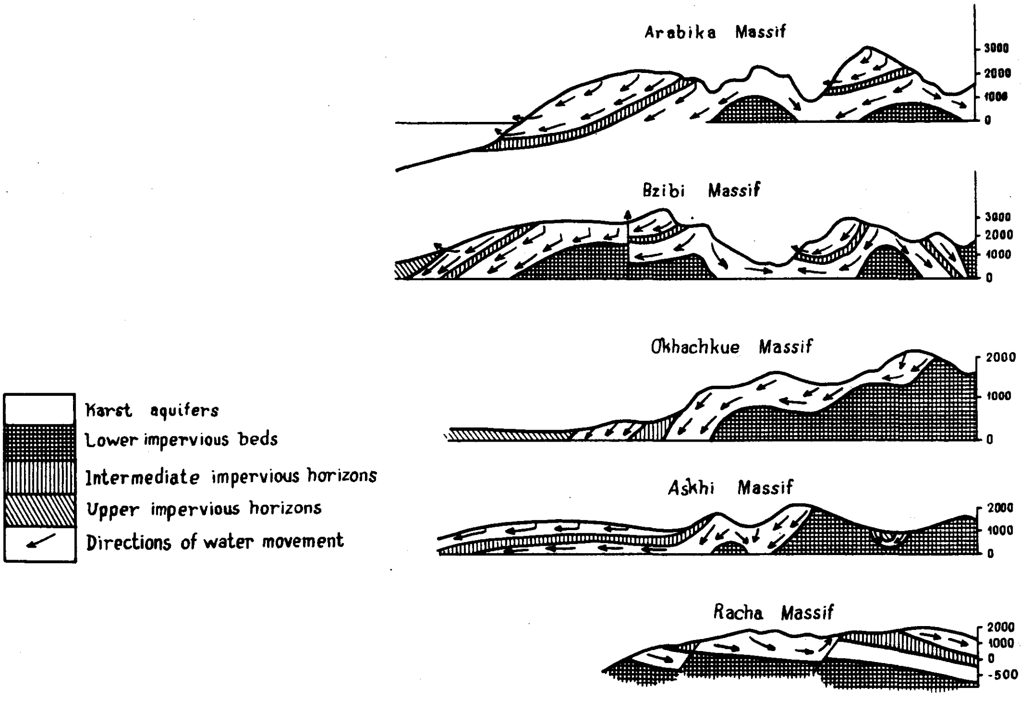

Figure 2. Cross sections of several aquifer-aquitard systems in West Georgia.

Karst Aquifers of the Upper Cretaceous-Lower Paleogene

Karst aquifers of the Upper Cretaceous-Lower Paleogene (#3 above) are composed of laminated, commonly marly and compact limestones which are badly fractured and shattered. This demonstrates intense karst processes within them and presence of ground water with fracture and solution circulation. This type of flow is common in the uplifted areas of karst masses, drained by river valleys. Somewhat deeper and southward, this flow is replaced by stratal-karst type flow. In Abkhazia, in the Aapsta-Kodori interfluve, this complex presents the primary karst aquifer and is characterized by abundant ground water in the zone of active water exchange. Water from karst aquifers flows through large streams, including the Kamani, Odishi, and Karasu, discharging from 0.6 to 7 m^3/s. Water temperature reaches 11° to 13°C with total dissolved solids concentration averaging 0.14 to 0.55 g/L and total hardness varying from 1.8 to 6.7 meq/L. Water composition is generally of Ca-HCO_3, Ca,Mg-HCO_3, and Ca, Na-HCO_3 type.

Karst Aquifers of the Lower Cretaceous Carbonate Deposits

Aquifers of the Lower Cretaceous carbonate deposits (#5 above), non-dissected Valanginian-Gotterivian with prevalence of Barremian stage, are vastly developed. This complex is represented by thickly laminated dolomitic and unstratified limestone, composing the karst massif of the whole limestone belt of West Georgia. In the western part of this belt, on the crests of alpine karst masses of the Gagra and Bzibi ranges, wherein denuded karst is common, surface runoff is absent. Many streams that drain this complex discharge at the Black Sea coast and its bottom, as well as in the river canyons. In Abkhazia these include the Reprua, Tsivtskala, Mchishta, Legveshara, Olori, Rechkhi, and Karasu; in the eastern part of the karst belt are the Verdzistava, Tsivtskala, Sharaula, Toba, and other streams. Their discharge ranges from 1 to 9 m^3/s, temperature regime is relatively constant at 6 to 18°C, water is poorly mineralized (0.25 to 0.5 g/L), and the total hardness is 1.6 to 65 meq/L. According to the chemical composition these waters belong to the Ca-HCO_3 and Ca,Mg-HCO_3 hydrochemical facies. In the south, toward the Georgian Block, this complex subsides below sea level, and is overlapped by young, thick deposits. Therein, ground-water circulation is somewhat hindered and results in relative water stagnation.

Karst Aquifers of the Upper Jurassic Complex

The karst aquifer of the Upper Jurassic carbonate deposits (#5b above) is composed of reef limestone and dolomitic rock, as well as marl of Lusitanian-Tithonian stage, and overlaps either the substratum of the Bajocian porphyritic suite or occasionally the Callovian-Oxfordian impervious zones. Karst areas in this complex occur in the northwestern part of Abkhazia, wherein, along with the Cretaceous deposits, they form the Arabika, Bzibi, and Khipsta karst massifs, and the entire Aschibakh-Lakorozi, Psheghishkha, and Dou massifs. These deposits are fractured and, thus, permeable, this being indicated by lack of surface runoff on highland plateaus. Alpine areas of these karst masses, at a height of 1,900 to 2,700 m above sea level, are denuded karst devoid of any cover. The karst aquifers of this complex intake abundant atmospheric precipitation and melt water and recharge. These are readily drained by rivers due to great dissection of the land surface. The flow regime of streams is rather variable and depends on intensity of atmospheric precipitation. The major streams of the complex are the Ghega waterfall, Lake Goluboe, and Djirkhva. Their discharge amounts to 2-4 m^3/s, with temperaure varying from 6 to 14°C, dissolved solids ranging from 0.18 to 0.40 g/L and total hardness of 1.2 to 4.2 meq/L; hydrochemical facies are of Ca-HCO_3, Ca,Mg-HCO_3, and Ca,Na-HCO_3 type. In the environs of Gagra, deep holes have tapped fresh water, which indicates deep dissolution of this complex.

STRUCTURAL GEOLOGY AND HYDROGEOLOGY OF KARST

The tectonic structure of the region presents a primary control on karst development. Surface and subsurface karst formations of West Georgia appear to be in close relation, both genetically and spatially, to tectonic environment. Intense linear folding, with asymmetrical folds, follows the general Caucasian trend. Planes of ruptures dip subvertically, which give a block nature to the tectonics of the region. Tectonic structures with a dense network of joints cause regular arrangement of surface and subsurface karst forms.

Within most karst masses of West Georgia, both the general contour and the minor karst forms bear structural pattern, that is, karst fields of the Tsebelda, Okhachkue, Askhi, and Racha. Therein the karst forms are developed mainly on the beds of synclinal folds. In this respect the Arabika and the Bzibi masses are exceptional, because they both are unusual for non-structural topography and nonconverging directions of relief forms and strike of tectonic structures. Maximum altitude of these masses above sea level, as compared with that of others, favored initial dissection of the primary structural relief. Highland regions of the masses then underwent Quaternary glaciation and the Wuermian glaciers of the valley type further altered their initial structural relief. Similar relationship to the tectonic structures is also traced in the river valleys which developed antecedent canyons.

Tectonic structure performs the primary role of development of karst water basins. In mountain karst, geosynclines cause nonconvergence of topographic and hydrogeological basin boundaries. The reasons for this lack of convergence include presence of tectonic shifting, mode of rock occurrence, block structure of the region, and the pattern of the top of the subterranean divide. All of these cause water to flow irrespective of the relief gradient (Kiknadze, 1972, 1979). Figure 3 shows examples of nonconvergence of topographic and hydrogeologic basin boundaries.

The above circumstance can be regarded as common for all the highland karst regions of the earth. To define the actual margins of karst basins we apply the complex method, involving comprehensive large-scale analysis of geologic-tectonic pattern of the region, location of subterranean divides, estimation of the course of karst water flow toward the discharge areas with employment of speleologic and tracer techniques, determination of the role of tectonic displacements at interbasin redistribution of ground water, calculation of water balance, etc. (Kiknadze, 1979).

Karst masses of the Great Caucasus Southern Slope are within the recent mobile tectonic belt. Therein occurs sequential bulging of karstifiable strata and water migration to great depths. Alteration of the stages of tectonic activity and repose generated cyclic processes of karst development. Every cycle of karstification begins with tectonic rise, develops karst intensely during stabilization, and proceeds until the following subsidence (Kiknadze, 1979). During the neotectonic period, the stages of active tectonic rise were rather prolonged as compared to relatively short repose intervals. This is indicated by the occurrence of highland karst cavities with deep, vertical sections such as Ilyukhin's system as deep as 1,240 m, the Arabika system (1,110 m deep), Pantyukhin's system (1,508 m deep), Snezhnaya-Mezhennogo (1,370 m deep), and others on the Bzibi mass. They are characterized by traces of polycyclic development, involving alternation of vertical and subhorizontal segments. Vertical areas, that is, interior pits and wells, are as deep as 20 to 300 m. Hence, the origin of karst aquifers, at the infiltration stage, was accompanied by intense and interrupted tectonic movements.

In origin and subsequent evolution of the karst water basins of the Great Caucasus, including West Georgia, one may differentiate several major cycles, coinciding with folding phases (Kiknadze, 1979).

1. The Styrian phase, preceeding the Middle Eocene, when the Great Caucasus mountain system developed. During this period depositional water was flushed by atmospheric water as the carbonate rocks were uplifted. At the same time, draining of the water-bearing formations of the folded structures became highly intensified.

2. The Attic orogeny, in Late Miocene, which brought about abrupt rise of mountains and then further deepening of the river drainage system along with tropical climate, marked the development of a new cycle of karst genesis. Within primitive karst basins water was drained by relatively shallow canyons of paleorivers. The Late Sarmatian development of karst processes of this cycle proceeded in Meotice as well, and when the height of mountains increased abruptly, appearance of new basins resulted within the median-mountain regions. Meanwhile, in the highland karst masses, hydrodynamic zones of vertically descending and subhorizontal circulation began their evolution.

3. The Rhodanian orogeny, at the delimination of Middle and Late Pliocene, caused continued rise of the mountain system and, hence, fall of temperature. By that time the river drainage had developed and activated the karst processes. During this cycle karst basins originated at the periphery of the Great Caucasus slopes.

4. The Valakhian phase, between the Pliocene and Pleistocene, caused the rise of the Great Caucasus mountain structures by another 1.5 to 2 km, and subsequent further fall of temperature and glaciation. At the same time, the Black Sea graben was developed. Intense contrasting movements caused highly convoluted ruptures and folding. All these factors favored deep

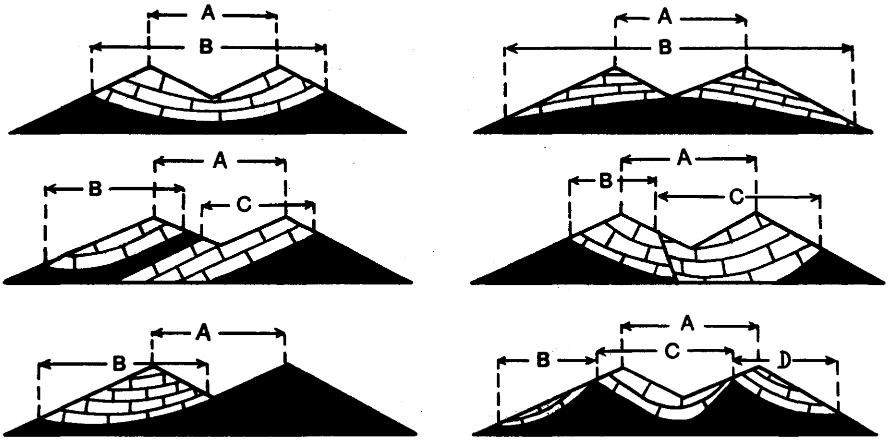

Figure 3. Schematic cross sections illustrating the nonconvergence of topographic and hydrogeologic basin boundaries.

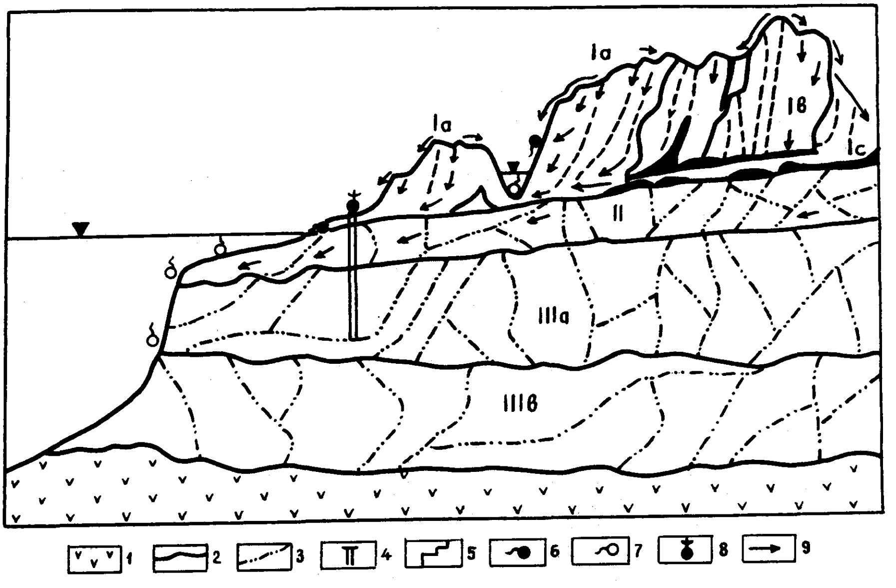

Figure 4. Cross section of karst in a mountainous area showing the zones of hydrodynamic zoning. The zones are further described in the text.

circulation of karst water. Within the denuded karst range, Wuermian glaciers overlapped the karst forms and closed the discharge areas of karst aquifers temporarily. However, after glaciation, aggressive glacial melt water favored intense revival of karst processes. Depression of the base level of karstification, intensification of erosion, deepening of valleys, and opening of structures gradually expanded the hydrodynamic zone with vertically descending circulation.

Thus, origin of karst water basins of West Georgia within the Great Caucasus Southern Slope began mainly during Mio-Pliocene, but these processes were in progress in Paleogene as well. However, the developed basins were eroded and preserved in the form of small fragments.

DYNAMICS OF KARST WATER

Water is recharged to karst aquifers mainly through infiltration and condensation in denuded and turfy karst or inflow and infiltration in mantled karst. The importance of these recharge sources varies for warm and cold periods and for valley, median-mountain, and highland karst environments. According to their dynamic properties, karst waters differ significantly from other ground-water types. Therefore, investigation into the environment, mode of occurrence, flow, and discharge of karst water is very urgent for these regions. But it is difficult to solve these problems for the mountain karst regions because application of estimation methods, commonly used in classical hydrogeology, is complicated and rather restricted in this environment.

Hydrodynamical Zoning

The general scheme of hydrogeologic zoning of karst water is still awaiting elaboration. The karst masses are subdivided into two hydrodynamic zones: vadose (pressureless) and phreatic (pressure) zones. Soviet hydrogeologists have worked out 3-to-9-stage schemes for karst water hydrodynamic zoning. The pattern of karst water zonation is determined by a whole complex of natural factors: that is, history of geologic development, tectonic-hydrogeologic and geomorphologic structure, hypsometry, thickness and spatial correlation of karstifiable and non-karstifiable strata, climatic conditions, altitude of a karst mass above sea level, and types of karst, that is, denuded, sodded, or mantled. Combination of the above factors in space and time determines the mode of ground-water circulation within karst masses.

Karst regions of the Great Caucasus offer beneficial material for elaboration of the scheme of hydrodynamic zoning. When developing the concepts suggested by the Soviet investigators, one of the present authors advanced a regional scheme of the karst water hydrodynamic zoning (Kiknadze, 1979). The scheme is demonstated by the example of the Great Caucasus, but it also allows for the characteristics of the dynamics of the karst regions developed over Europe. There are 3 zones with a number of subzones recognized in the cross section of karst masses of the mountain areas (Fig. 4).

I. The zone of gravity (free, active) circulation. In its upper part, karst water is formed by infiltration and inflow of atmospheric water as well as moisture condensation within karst cavities. This water flows along involved courses toward discharge areas, giving birth to insulated or concentrated flows. The thickness of the zone reaches more than 2 km, largely depending upon occurrence of karstifiable rocks and hypsometric level of the mountain karst. According to the pattern of karst water dynamics the following subzones are differentiated within this zone.

Ia. That of intermittent circulation of rain and melt water over the surface of karst masses. Circulation paths are rather short; streams velocity depends upon the surface gradient and amount of water. The water is highly aggressive to carbonate rocks, being especially active in the denuded-karst environment. Herein it directly affects the exposed rocks. As a result, various surface karst forms are developed. These are micro- and macro-corries, corrie fields, funnel sinks, depressions and cavities of various shapes and dimensions. In river canyons, cupholes and giant kettles, slope niches, etc., become rather common. These waters are characterized by low dissolved solids content from 50 to 150 mg/L, Ca-HCO_3 composition, and a temperature similar to mean annual.

Ib. That of subsurface descending circulation. Under the conditions of highland karst, rise above the regional and general base level of karstification it has somewhat a hypertrophied dimension, depending upon geologic-tectonic and hypsometric environment. Thickness of the subzone exceeds 1 to 2 km. Ancient tectonic movements and alternation of bulging and quiescent state affect the pattern of karst water circulation, that is, whether it is vertical, proceeds along the dipping canals, or is of a combined nature. Water of the subzone is formed by infiltration, inflow, and condensation. In the upper part it is much more aggressive than in the lower part. Total mineralization of water accounts for 200 to 300 mg/L with Ca-HCO_3 and Ca,Mg-HCO_3 composition, and temperature of 2 to 8°C. The subzone involves all the well-known subsurface karst forms: wells, holes, precipices, creating sometimes complicated karst systems and reaching great depths, such as, Saint-Bernard precipice in Savoyan Alps, France (1,602 m deep), Pantyukhin's precipice in the West Georgia (1,508 m deep), Snezhnaya in the West Georgia (1,370 m deep), Pier-Saint-Marten precipice at the France-Spain boundary (1,310 m deep), and Ilyukhin's system in West Georgia (1,240 m deep), with alternating vertical, dipping, and subhorizontal sections, subsurface rivers, lakes, and waterfalls.

Ic. That of subsurface subhorizontal circulation. It is just below zone Ib and within mountain karst, its thickness reaches 200 to 300 m. In certain cases this thickness is controlled by the range of many-storied subhorizontal canals. The water is Ca-HCO_3 and Ca,Mg-HCO_3 type with total mineralization of about 300 to 500 mg/L, and temperature within 6 to 12°C. Rates of water flow are much less than those in the previous subzone, though they increase abruptly during floods. Within the subzone, vast subhorizontal cavities are developed, with their lengths ranging from several hundred meters to tens of kilometers. Sometimes these cavities have siphonal canals within their discharge areas, such as Mchishta, Gega, and Krasnie caves in the Caucasus.

We may also come across subaqueous water discharge at the beds of river valleys, springs in the Khobistskali canyon in Samegrelo, with rather local head conditions.

II. The zone of relatively hindered circulation. It is located under zone I, usually in the center of karst masses. Ground water circulates under the head effect and may discharge through submarine springs such as in Gagra-Gantiadi and Sochi environs, depending on the position of the base water-resisting bed relative to sea level. Thickness of the zone reaches 200 to 300 m. The dissolved solids content of water ranges from 300 to 500 mg/L, and temperature is usually about 10-14°C with $Ca\text{-}HCO_3$ and $Ca,Mg\text{-}HCO_3$ composition.

III. The zone of deep circulation. It is below sea level, under zone II. Herein we may distinguish two subzones differing from each other by the karst water dynamics pattern.

IIIa. The subzone of relative stagnation, wherein karst canals are filled with water beyond the effect of the local base level of karstification. Karst aquifers may be tapped only by boreholes such as in Abkhazia, the Khasupse, and the Gagripsh. Water is fresh and poorly mineralized, from 400 to 600 mg/L. Temperature may reach 30-45°C, largely depending on the depth of occurrence. Hydrochemical facils are the following: $Na,Ca\text{-}HCO_3$, $Ca,Mg\text{-}HCO_3,SO_4$, and $Ca,Na\text{-}HCO_3,SO_4$. Submarine discharge is also possible.

IIIb. The subzone of stagnant regime. This subzone appears at even greater depths, which are already beyond the influence of the general base level of karstification. Deeply submerged carbonate rocks contain mineral and thermal water which, under favorable tectonic and hydrodynamic conditions, may recharge the above situated karst water zone and participate in karst-genetic processes, such as mixing corrosion. Hydrothermal karst phenomena may develop in this zone.

The proposed scheme of hydrodynamic zonation of karst regions in mountain countries offers objective reflection of hydrodynamic relations observed within karst masses. For the highland karst it becomes slightly altered due to abrupt increase of the thickness in subzone Ib, that of subsurface descending circulation. In the valley karst regions subzones of Ib-type usually prevail. Zone III initiates its development just when carbonate rocks subside below sea level and become overlapped by non-karstifiable deposits. Specific structural geologic and hydrogeologic conditions may bring about some deviation of the general scheme.

Direction and Rate of Karst Water Flow

To ensure solution of scientific aspects of karst-genesis and water-economy problems, which are rather acute in karst regions, it is necessary to determine hydrodynamic relations between individual drainage systems in the recharge areas. With this aim in view, speleologic and tracer techniques of investigation are applied.

Experiments revealed correlation of the Zhove-Kvara with Reprua spring, in the Arabika mass, of the absorbers in the Psirtskha upper reaches with the spring on the left-bank slope of the Tskvara in the Bzibi mass, of subsurface reservoirs of New-Aphoni cave with Vauclusian springs on the Manikvara in Gumista mass, of a subsurface stream in Tsivtskala cave and Amtkeli lake with Karasu spring in the Tsebelda mass, of the channel in the Turchu polje with the Tobi karst river in the Askhi mass. The recent experiments (August 1984) on the Arabika mass allowed determination of the direction and rate of ground-water flow from the crest part of the mass, the recharge area, and the periphery, the discharge area. The dye introduced into the subsurface stream in the Arabika (Kuibishevskaya) precipice at the depth of 500 m below the surface (absolute height of the entry is 2,180 m) was traced a fortnight later on the Black Sea littoral, just in the Kholodnaya Retchka and Reprua karst springs and in the well, located near the Kholodnaya Retchka. Thus, the estimated average rate of flow ranged from 1,100 to 1,200 m/day. Another dye, eosine, injected into the Ilyukhin's system (absolute height of location is 2,300 m) was recognized in the same springs by just slight traces. So, it was concluded that the Arabika mass contains the karst aquifer, which proves to be the deepest in the world whose depth reaches 2,300 m.

Tracer methods allowed the statement that hydrogeologic structure of the Arabika mass is not subordinate to rigid plicated structures, as it was previously suggested by one of the present authors (Kiknadze, 1972). Thus, for the present, only three major aquifers have been identified in the southern part of the mass (Klimchuk et al., 1987) instead of the complicated system of multi-karst water basins.

Tracer tests were accomplished several times on the Bzibi mass, in the Snezhnaya-Mezhennogo cave system, which is the deepest karst precipice of the world (1,370 m deep, 19 km long, with the volume of 1.74×10^6 m^3). In 1986, the experiment affirmed relation of the above precipice to the Mchishta Vaukluzian spring, altitude of the outlet is about 70 m. Rhodamine was introduced into the cave system at the depth of 700 m and after 5 days it was traced in the spring. The test proved that the west and east parts of the Bzibi mass present a single hydrogeologic system with well-developed sub-riverbed karst channels under the Khipsta and Eegri valleys (Dublyanski et al., 1987). Table 1 lists the results of the tracer techniques, accomplished within the West Georgia karst area.

Table 1. Rate of water flow in the karst of the Great Caucasus.

Karst massif	Extent of the flow under study* (m)	Rate of flow (m/day)	Date of testing
Arabika	3,500	4,900	August 1979
Arabika	17,400	970	August 1984
Bzibi	9,000	2,300	June 1974
Bzibi	28,200	3,100	August 1986
New-Aphoni	1,000	400	February 1972
New-Aphoni	900	1,300	March 1972
New-Aphoni	7,500	2,500	September 1974
New-Aphoni	4,200	2,100	November 1974
Otoyush-Chaama	5,500	11,000	July 1959
Tsebelda	8,000	1.760	October 1970
Tsebelda	10,000	1,900	October 1970
Askhi	4,600	12,270	July 1972
Okriba-Argveta	300	3,900	September 1970
Okriba-Argveta	300	2,600	September 1970
Dzirula	4,200	2,400	March 1973

* Estimated in a straight line, with karst cavity meandering factor correction (1.3).

Chemical Composition of Karst Water

Karst water of highland regions, such as the Great Caucasus and Georgia, are formed by mixing atmospheric inflow and condensation water with that of snow and ice melting within karst cavities. The dissolved solids content of rainfall, which fell during the warm period, varies between 7 and 80 mg/L. Rain and snow, which fell during the cold period, are of Na,Ca-Cl,HCO_3 composition with dissolved solids ranging from 4 to 80 mg/L. Condensation water has a no dissolved solids at the moment of its origination, but after flowing the first dozen meters it acquires mineralization of 70 to 140 mg/L and Ca-HCO_3 composition. Nearly the same dissolved solids is characterized for the infiltration water. Both snow and ice in the karst cavities have Na,Ca-HCO_3 composition with dissolved solids concentration ranging from 6 to 250 mg/L. Dissolved solids of ice changes from 6 mg/L for sublimated ice to 85 mg/L for hydrogenic ice, depending on its origin. Chemical composition of inflow water is formed at the surface. These are Ca,Na-HCO_3 waters with dissolved solids concentrations as great as 140 to 350 mg/L. Karst spring water has Ca-HCO_3 and Ca,Mg-HCO_3 composition with dissolved solids ranging from 180 to 370 mg/L. Composition of water of springs flowing out of various karst blocks may differ significantly, thus serving as a hydrochemical indicator of karst processes and the structure of karst systems. The chemical composition of karst water in West Georgia is shown in Figure 5.

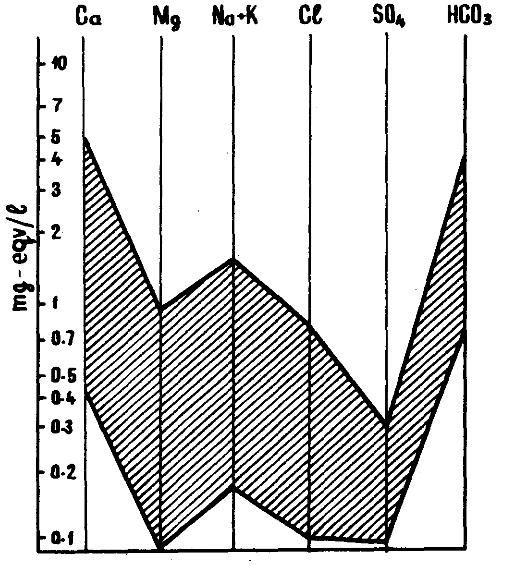

Figure 5. Chemical composition of karst waters.

All of the above illustrates that the chemical composition of karst water occurring in highland regions depends upon its genesis, discharge, rate of flow, temperature, and the composition of the rocks it contacts. These are mainly aggressive waters, whose dissolving capacity is maximum within hydrodynamical zone I, especially in its upper part. As for the lower zones herein, a major role is performed by mixing corrosion due to temperature contrasts and variation of composition of inflowing solutions (Dublyanski and Kiknadze, 1984).

Temperature of Karst Waters

The temperature of karst water depends on the recharge pattern, aerometric coefficient, and hydrodynamic zoning, as well as the morphology of karst cavities and local climate within them and the season.

Within the Great Caucasus of the Arabika and the Bzibi karst masses, wherein springs of hydrodynamical zone I are recharged by glacier and snow melt waters, the temperature of karst water is about 1 to 6°C. As for hydrodynamical zones I and II, herein temperature of water in caves and flows ranges from 1 to 13°C. In warm seasons karst waters are recharged due to infiltration, inflow, and condensation, while during the cold season by inflow and partially by infiltration. Depth of occurrence largely determines the different temperatures of these waters, which are 7 to 10°C the whole year, for example, the Ghega waterfall at 7.3°C, Goluboe lake at 8.1°C, Kholodnaya Retchka at 10°C. Water of the third hydrodynamical zone have different temperatures depending upon recharge, flow, and discharge patterns. There were cold waters trapped in the solutionally modified fault zones at a great depth, such as by the well in the Gagripsh canyon.

The majority of karst regions of the Great Caucasus belong to open heat-water-exchange geosystems. The presence of numerous cavities and karst fissures indicate rapid rock aeration and washout by quickly rejuvenated cold karst waters. In cold seasons, the fissured-karsted cavities are recharged by infiltration and inflow waters, the latter causing cooling of the upper part of hydrodynamical zone I. At the same time, due to incoming cold air and evaporation of moisture, the karst cavities get chilled, favoring origin of spelean ice varieties at their entry-side areas. In the highland karst cavities, along with seasonal ice accumulation, are perpetual accumulations of snow and subsurface alpine glaciers which cause continuous recharge of minor springs in the upper zone and general recharge of the karst mass. The cooling effect of karst development, especially notable within upper hydrodynamic zones, sometimes manifests itself at great depths as well (Dublyanski and Kiknadze, 1984).

CONCLUSIONS

Regional geostructural, geomorphologic, and climatic factors control karst processes of development and determine hydrogeologic and engineering geologic environments. Hydrogeological investigation of the Great Caucasus carbonate karst reveals considerable natural karst water resources, which prove to be highly useful for the national economy. Therefore, further investigation and reasonable utilization should be regarded as deserving proper consideration.

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KARST AQUIFER OF THE CRACOW-WIELUN UPLAND, POLAND

J. GLAZEK*
Warsaw University
Institute of Geology
Al. Zwirki i Wigury 93
02-089 Warszawa
Poland

A. PACHOLEWSKI
Geological Survey of Poland
Upper Silesian Branch
ul. Bialego 1
41-200 Sosnowiec
Poland

and

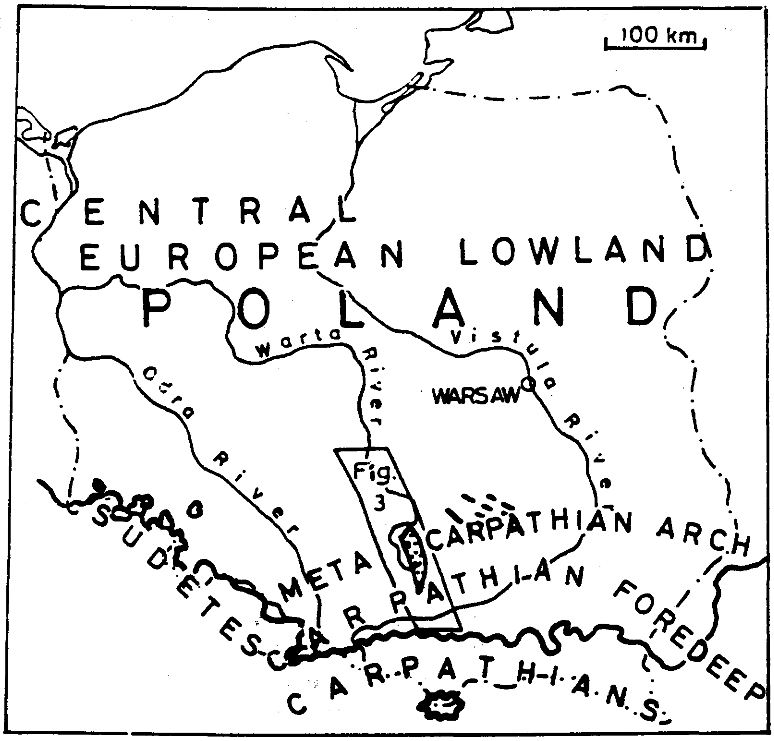
A. ROZKOWSKI
Silesian University
Department of Hydrogeology
ul. Mielczarskiego 60
41-200 Sosnowiec
Poland

INTRODUCTION

The Cracow-Wielun Upland, abbreviated CWU, is the most extensive and uniform karst region of Poland. It is a belt of Upper Jurassic limestone stretching from Cracow in the southeast to Wielun in the northwest on the northeast slope of the Silesian Upland. This upland is part of the Meta-Carpathian Arch separating the Carpathian Foredeep in the south from the Central European Lowland in the north (Fig. 1). The CWU is formed by a monoclinal belt of Upper Jurassic limestone gently dipping to the northeast (Glazek et al., 1972, 1982).

The CWU covers an area of 2,900 km^2 and forms an upland 160 km long and 10 to 30 km wide. The altitude ranges from 165 m on the Warta River in the northwestern corner to 504 m in the vicinity of Zawiercie. Residual hills (Fig. 2) separated by infilled karst depressions are the most characteristic feature of this upland. The depressions, with diameters of up to several kilometers and depths of 110 m are filled with kaolinized sandy-clay deposits of Tertiary age, with a cover of Pleistocene deposits, mostly outwash sand. More than eight hundred small caves are known in this area, but only two of them reach 1 km of aggregate passage length (Szelerewicz and Gorny, 1986).

**Present address*: Adam Mickiewicz University, Institute of Geology, ul. Maki Polne 16, 61-606 Poznan, Poland



maximum extent of glaciation

nunatak

Figure 1. Sketch map showing the location of the Cracow-Wielun Upland within the territory of Poland, as well as the general geomorphic situation of this Upland as a part of the Meta-Carpathian Arch between the Carpathian Foredeep and the Central European Lowland. A high-elevation region of the Meta-Carpathian Arch remained unglaciated during the Pleistocene.

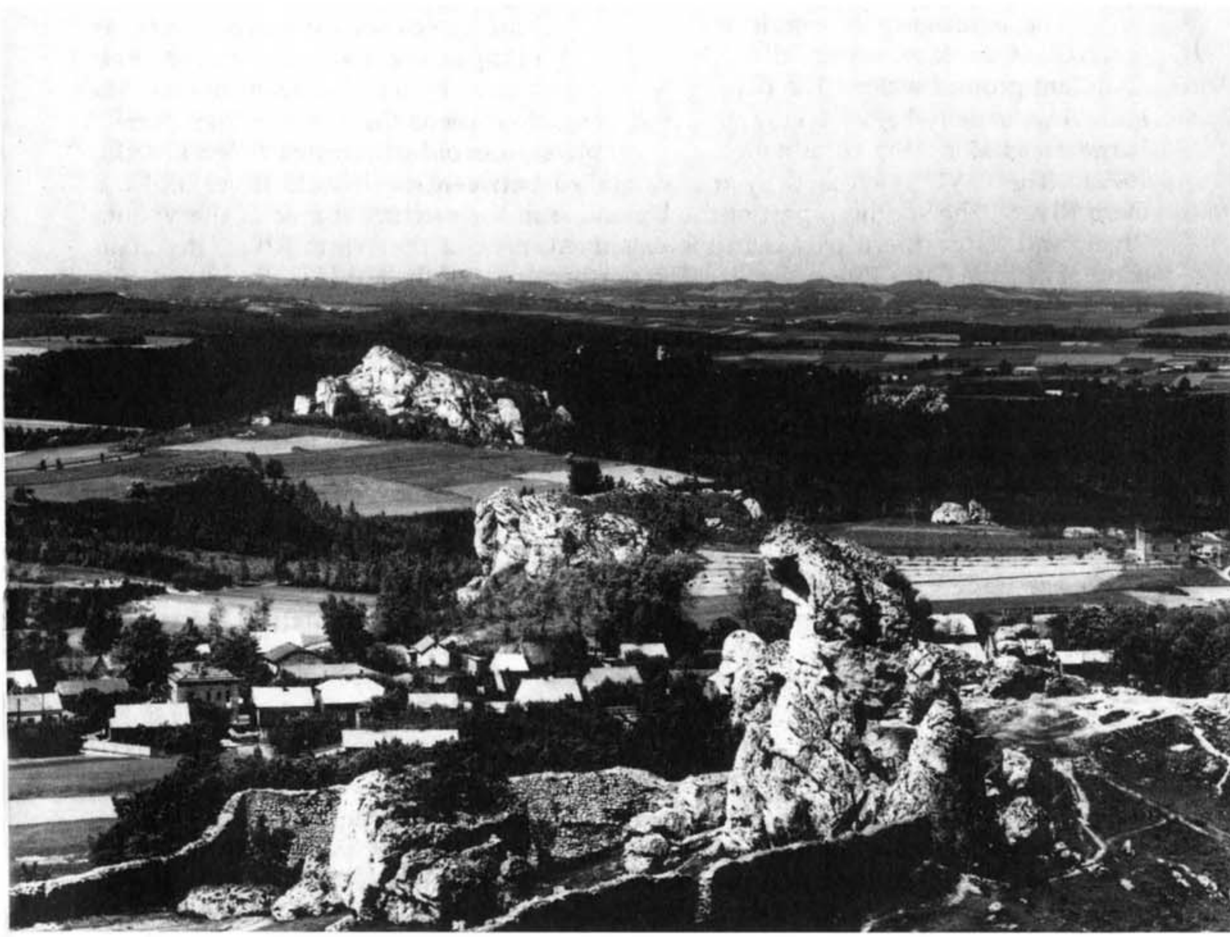


Figure 2. Typical landscape of Cracow-Wielun Upland. Residual hills of sponge cyanobacteria bioherm separated by depressions filled with Cenozoic deposits. Note: (1) the thick bedded detrital limestones of biohermal talus forming towers in the center (surrounded by walls of medieval castle), (2) the bioherm itself was to the left of these towers, (3) visible hills are case hardened and resistant to solution. Photo T. Zapasnik.

The hydrographic conditions of the CWU are typical of karst areas, such as scarcity of surface water, disappearing and reappearing surface streams, and abundant ground water. The density of rivers, counted by the "Schaefer method," is quite low, usually below 1 km/km^2, and nowhere exceeds the value of 2 km/km^2. Large areas along the culmination are completely devoid of streams (Kleczkowski, 1972). The CWU is crossed by the watershed between the Vistula River and the Odra River. The southern part of the Upland is in the catchment area of the Vistula River, and its northern part is in the catchment area of the Warta River, the main tributary of the Odra River (Fig. 1). The northern part of the CWU is dissected twice by gaps of the Warta River, southern one near Czestochowa and northern one near Wielun. The southern part of the CWU forms a secondary watershed between several lesser tributaries of the Vistula River (Fig. 3).

The mean annual temperature is 7.5° C; the mean annual precipitation is 750 mm in the south and 650 mm in the north. About 400-500 mm of precipitation fall in the summer, from May through October. The snow cover averages 80 days. During dry years the total precipitation amounts to 400-500 mm, and in wet years it ranges from 700 to 1,000 mm (Kleczkowski, 1972). A high runoff coefficient of 35.3%, the underground runoff consisting of 85% of total runoff, and the amount of infiltration varying between 28.4 and 48.3% of the annual precipitation are typical only of karst aquifers in Poland.

THE GEOLOGICAL FRAMEWORK

Geologic History

The CWU consists of a thick Upper Jurassic carbonate sequence underlain by older Jurassic, Triassic, or Late Paleozoic deposits and is partly covered by Upper Cretaceous and Cenozoic deposits of various permeability. The karst aquifer is composed of Upper Jurassic carbonates. The substrate of this aquifer is composed of Late Paleozoic rocks in the most southern part of the CWU. Further to the north, in the vicinity of Zawiercie, the karst aquifer is underlain by Triassic rocks. In the central and northern parts of the Upland, the base of the karst aquifer is composed by Lower-Middle Jurassic carbonates (Fig. 3). In the vicinity of Cracow, the transgressive Callovian deposits cover the karstified Givetian through Visean carbonate formations. Further to the northwest, the Callovian occurs over various continental deposits of Late Carboniferous through Lower Triassic age that are generally of low permeability. During the late Lower Triassic this area was inundated by the sea which deposited more than 300 m of shallow-water carbonates (Muschelkalk) that form a separate fissure-karst aquifer (Motyka and Wilk, 1984) and contain famous hydrothermal karst deposits of lead-zinc ores (Bogacz et al., 1970). This aquifer is in hydraulic contact locally with the Jurassic limestone aquifer. However, the regressive sequence of late Middle Triassic and continental deposits of the Upper Triassic age (Keuper and Rhaetic) covered Triassic carbonates with clay deposits that form an aquiclude dividing the karst aquifer within Middle Triassic rocks from those within Jurassic rocks between Olkusz and Zawiercie (Fig. 3).

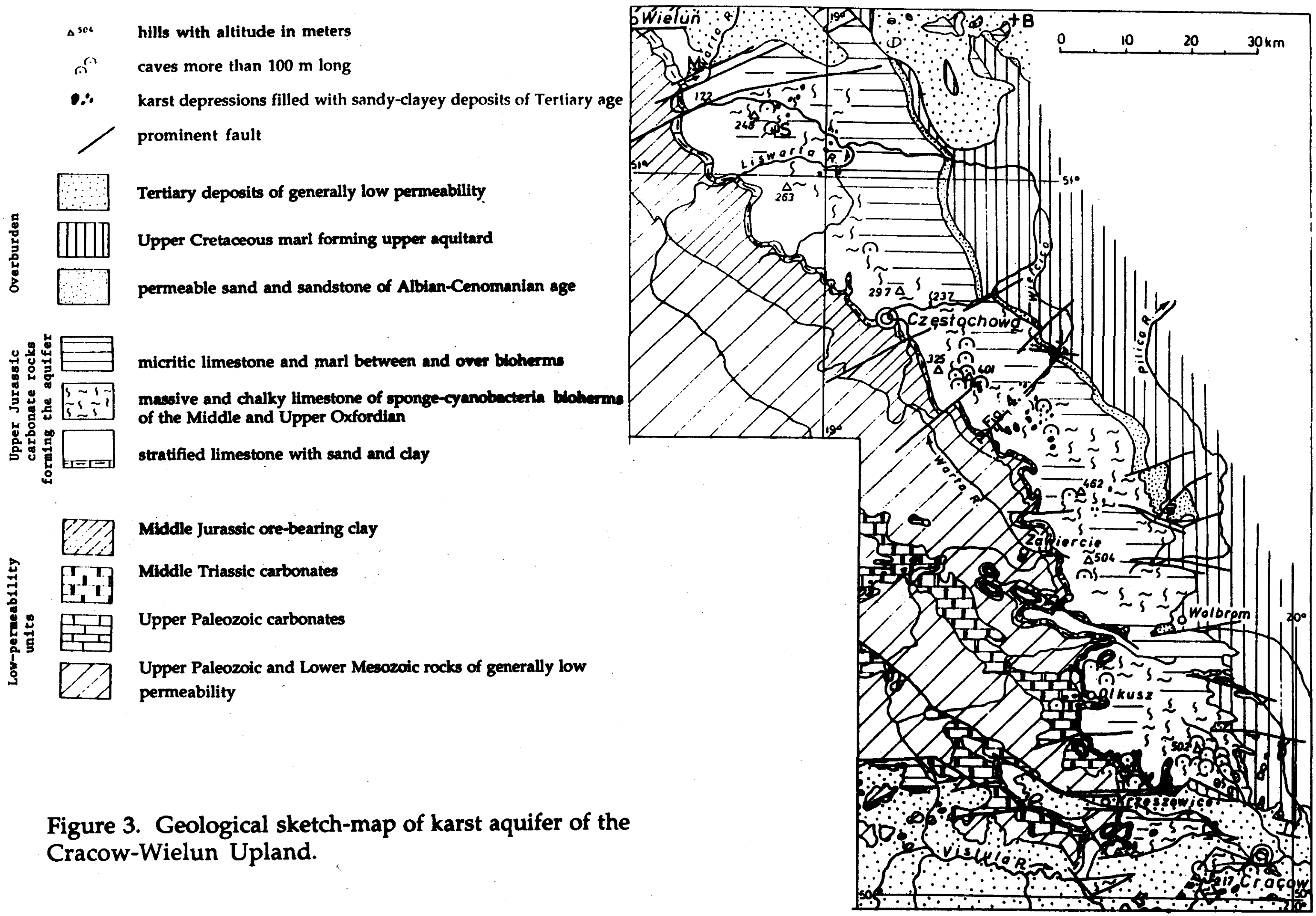


Figure 3. Geological sketch-map of karst aquifer of the Cracow-Wielun Upland.

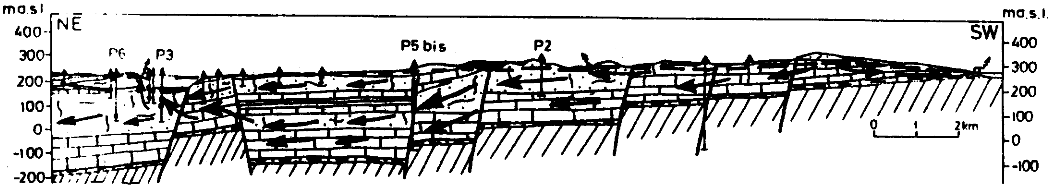
The early Jurassic sediments consist of freshwater sandy-clay deposits of generally low permeability. The area of the CWU was gradually inundated from the north by the Middle Jurassic transgression. Deposits of this transgression begin with sandstone of Aalenian and Early Bajocian age, forming the local aquifer, their thickness reaching 40 m in the northern part of the CWU and wedging out to the south. Later in the Middle Jurassic, marly clay and siltstone with siderite intercalations and concretions called "ore-bearing clays" were deposited. These "clays" reach thicknesses of 220 m in the vicinity of Wielun and gradually wedge out southward; in the area of the Wiercica basin (Fig. 4) they are 140-115 m thick and disappear completely between Zawiercie and Olkusz. The "ore-bearing clays" form an aquiclude beneath the karst aquifer in the northern and central parts of the CWU.

The detailed stratigraphic column of the Jurassic carbonate deposits is complicated (Kutek et al., 1977; Wierzbowski et al., 1983). The Jurassic carbonate sequence starts with thin, sandy, or marly deposits with stromatolites of Callovian-Lower Oxfordian age. Beginning in the Middle Oxfordian, the carbonate sedimentation was accelerated and strongly differentiated (Fig. 3). The sequence begins with sponge-tuberolitic layered limestone with subordinate and generally small cyanobacteria-sponge bioherms. They are overlain by strata displaying numerous large sponge-cyanobacteria bioherms and biostromes, massive or chalky limestone, interfingering with well-bedded, poorly-fossiliferous micritic limestone and marl, platy limestone, friable limestone, and marl. The advanced development of sponge-cyanobacteria bioherms resulted in uneven topography of sea floor and detrital limestone taluses around bioherms. The zones of advanced development of these accumulations appear related to tectonic elevations of tilted block edges in the basement of the Upper Jurassic. The Lower Kimmeridgian strata consisting of well-bedded micritic limestone, marl, grainstone, and lumachelles with limestone bioherms are preserved beneath the Cretaceous deposits only in the northern part of the CWU, where the thickness of Upper Jurassic carbonates exceeds 700 m. These rocks display complicated patterns of porosity and karstification.

Three general types of calcareous rocks may be differentiated:

1) Thin-layered marly limestone and marl in the bottom part of the sequence (Lower Oxfordian - early Middle Oxfordian) and in the upper part of the sequence (latest Oxfordian - Lower Kimmeridgian). Rocks of this type contain less than 90% $CaCO_3$ and in some layers only 65-70% $CaCO_3$. These micritic rocks are characterized by a low primary porosity and a limited secondary porosity along bedding planes, joints, and faults. These rocks are of low permeability and are slightly karstified along surfaces of discontinuity. They form aquitards within the Upper Jurassic carbonate sequence.

2) Layered micritic limestone, detrital, sponge-tuberolitic, platy, or friable limestone occur below and between biohermal limestone and contain allochems, clay admixture, flint, and micrite groundmass in various proportions. Friable limestone is micritic, primary microporous and contains no macrofossils. Detrital limestone with micrite, dismicrite, and sparite groundmass form fringes



Quaternary sand with intercalation of gravel, silt, and boulder clay

Tertiary clayey sand filling karst depressions

Upper Cretaceous marl with sand and sandstone of the Albian-Cenomanian age at the bottom

Oxfordian

marl

sponge-algal bioherms

bedded micritic limestone

Middle Jurassic ore-bearing clays with sand and marl of the Upper Callovian and Lower Oxfordian at the top

faults

springs

wells with caverns encountered during the drilling

potentiometric surface

ground-water flow

Figure 4. Hydrogeological section through karst catchment area of Wiercica river.

of bioherms and reveal a lesser primary, interparticle, porosity, but a prominent secondary, moldic, reactive, bedding plane, fracture, and channel porosity. Platy limestone is micritic, contains only sporadic macrofossils and in places passes into marly limestone. Sponge-tuberolitic limestone contains numerous sponges and other fossils, clay minerals, and diagenetic chert and reveals some interparticle and shelter, primary porosity and some moldic, reactive, bedding-plane, fracture and channel secondary porosity. Generally rocks of this type are pure limestone, over 90% $CaCO_3$, and reveal some primary porosity but more prominent secondary porosity. They are of intermediate permeability and karstifiability.

3) Biohermal limestone, massive and chalky limestone, represents similar sponge-cyanobacteria accumulations composed of unbedded or thick-bedded rocks. They replace each other laterally and vertically and essentially differ by diagenetic lithification. The chalky limestone is friable and microporous; its intercrystalline porosity reaches 30%. These rocks contain abundant flint as a result of diagenetic redistribution of sponge-derived silica. The massive limestone is hard and devoid of primary porosity as a consequence of early lithification. These rocks are devoid of flint and chert. These rocks are pure limestone, over 95% $CaCO_3$, and reveal well-developed moldic, reactive, fracture, vug, channel, and cavern secondary porosity; thus, they are permeable and strongly karstified. Within these rocks occur large caves and caverns with up to 20 m height. Also sinkholes and large karst depressions are more than 1 km in diameter and 100 m in depth. All these karst forms are filled partly or completely with Cenozoic deposits (Figs. 3 and 4).

The overburden of the karst aquifer is composed of Upper Cretaceous deposits in the northeast slope of the CWU (Fig. 3) and of thin Cenozoic deposits filling karst forms and mantling slopes of the Upland. The Cretaceous sequence deposited on the eroded top of Upper Jurassic carbonates begins with permeable sandy deposits of Albian-Cenomanian age and of variable thickness, from less than 1 m up to 150 m. The thickness patterns of these deposits resulted from synsedimentary block movements which accompanied subsidence (Marcinowski, 1974). These water-bearing sands and sandstones are in hydraulic connection with the underlying karst aquifer. As a result of Tertiary interstratal paleokarst, these sands filled sinkholes in Jurassic limestone near the surface boundary of these formations (Glazek and Szynkiewicz, 1980). Since the Turonian, this area was covered by thick deposits of Late Cretaceous age which form an aquiclude on the northeast slope of the karst aquifer.

A considerable uplift associated with block tilting was caused by the Laramian movements which started in the Late Maastrichtian. In consequence of these movements, the Upper Cretaceous deposits were eroded and intense karstification of Jurassic carbonates began. This karstification was blocked by Middle Eocene subsidence and transgression when the area was partly covered with thin detrital deposits preserved only locally near Wielun (M in Fig. 3). This area emerged during the Oligocene and was covered in places with lignite-bearing Oligocene - Lower Miocene deposits preserved only in the northern part of the CWU as outliers in karst depressions. The largest such deposit--Belchatow Mine--is marked B in Figure

3. The pre-Middle Miocene uplift resulted in the creation of erosional relief and karstification in the whole area under consideration. This erosional relief was covered by a Middle Miocene transgression from the south where the Carpathian Foredeep was formed. This transgression left sandy-clayey deposits of shallow-marine to coastal plain and alluvial plain environment. These deposits were eroded since the Late Miocene as an aggregate result of the Messinian Crisis, and the Pliocene uplift of the result of the Messinian Crisis, and the Pliocene uplift of the whole area (Glazek and Szynkiewicz, 1987). The karst sculpture being exhumed since the Lower Pliocene was "fossilized" with cave deposits containing Pliocene and Early Pleistocene vertebrates (Glazek et al., 1975; Bosak et al., 1982; Glazek and Szynkiewicz, 1987; Glazek, 1989).

Until the Middle Pleistocene and the continental glaciations, the whole area was drained into the Tethyan (Carpathian) basin and later into the Paratethyan (Fore-Carpathian) basin to the southeast. During the Middle Pleistocene, continental glaciations covered most of the territory of Poland with ice-sheets (Fig. 1). At this time the more elevated parts of the Meta-Carpathian Arch formed nunataks. The CWU was never completely covered by the ice-sheet. A concave nunatak was formed and covered with outwash sands (Rozycki, 1965) up to 410 m elevation. The less elevated limbs and the northern part of the CWU, north of Czestochowa, were covered with boulder clay during the maximum glaciation of the Middle Pleistocene (San Glaciation = Elster Glaciation). The two later glaciations (Odra = Saale I and Warta = Saale II or Warthe) reached the slopes of the CWU and left similar deposits down the slopes of the Upland. Since the beginning of the Upper Pleistocene this area has undergone denudation and further karstification. During the Middle Pleistocene the Baltic Sea depression was formed and a new river pattern consequent to this depression was created and the drainage to the Black Sea ceased (Dzulynski et al., 1968).

Tectonics

During the Alpine Diastrophic Cycle, the whole area of the CWU belonged to the stable continental margin of the European Continent facing the Tethys Ocean. This area was invaded several times by epicontinental transgressions and slightly affected by several tectonic events. These events were registered in alternating subsidence periods of extensional deformation and uplift periods of compressional deformation. The sedimentation of the deep neritic Upper Jurassic sequence of carbonate rocks was caused by extensional subsidence of several blocks which controlled the facies patterns and enabled accumulation of more than 500 m of carbonate sediments. This sequence was asymmetrically uplifted and eroded during the Lower Cretaceous. As a result of these movements, the Late Upper Jurassic rocks were completely eroded in the south (near Cracow) and partly preserved in the north. The angular unconformity between the Upper Jurassic and Upper Cretaceous sequences is on the order of 1°.

The next subsidence of Albian-Cenomanian age was accompanied by extensional tectonics along the eastwest faults, visible in outcrops and the occurrence of sandy deposits (Fig. 3). The subsequent uplift of the Silesian Upland with a NW-SE axis caused the erosion of the Upper Cretaceous overburden and was connected with Laramian movements. These movements are responsible for the present-day monoclinal structure of the CWU and the regional dip of strata of the order 2-3^{o} to the northeast. Local dips of strata are frequently much greater and inconsistent with this regional dip due to faulting and tilting of tectonic blocks.

An important faulting took place during the Neogene when extensional tectonics caused subsidence of the whole area and the transgression of the Middle Miocene sea to the north. This tectonic activity is responsible for a prominent fault pattern of E-W direction (Fig. 3) and the reactivation of some older faults. These faults opened during the Late Neogene uplift of the Meta-Carpathian Arch, and became water-bearing. In general, the fault pattern consists of dominant subparallel directions E-W and subordinate NE-SW and N-S direction. The joint pattern is more complicated, but subvertical joints of NW-SE and NE-SW directions dominate. This pattern is complicated in detail due to different relations of separate blocks and the believed dextral shift in the substratum of the Upland (Krokowski, 1984). Moreover, the joint pattern is strongly dependent on lithology. According to Gradzinski (1962), joints in the massive limestone are rare but prominent, average distance between them is 2 to 3 m; in layered limestone they are thin and dense, with average distance is less than 1 m.

Karst

The Upper Jurassic carbonate aquifer was affected by karstification during two periods, namely: 1) the Early Cretaceous and 2) the Cenozoic. Moreover, two older periods of karstification were differentiated in this area and affected the Upper Paleozoic and Middle Triassic carbonates immediately underlying the aquifer in the southern part of the CWU (Fig. 3).

This area was inundated until the Tithonian and probably also the Beriasian-Hauterivian when marly-clay sediments were deposited over the Kimmeridgian limestone in the basin. Thus the first period of karstification probably began during the Hauterivian and persisted until the beginning of Albian (Glazek and Szynkiewicz, 1980). During that time all sediments, including the Lower Kimmeridgian in the north and the Upper Oxfordian in the south, were eroded. But the karst macroforms of these ages are almost absent. Probably there were several specific causes for the insignificant extent of karst development during that period: 1) the low relief landscape until the Albian time was unfavorable for deep groundwater circulation; 2) the long persistence of clayey, impermeable, regressive sediments; and 3) the strong secondary silicification of exposed Upper Jurassic limestone on the pre-Albian surface (Glazek, 1989).

Unequivocal pre-Lower Cenomanian karst forms are known only in the Julianka Quarry (J in Fig. 3) where a hill of completely silicified Upper Oxfordian chalky limestone contains remnants of a cave that developed prior to silicification and submergence beneath Middle Cretaceous clastics. Nearby, on the other slope of that buried hill, a pocket filled with silicified limestone debris and covered with Lower Cenomanian clastics was described. A subtropical savannah-like climate has been suggested for this period (Glazek and Szynkiewicz, 1980; Glazek, 1989). This karstification unequivocally affected the porosity of Oxfordian limestone.

The final karstification period began with the Upper Cretaceous emergence, probably the Late Maastrichtian, and has persisted to modern times. This long period encompasses many phases that are more or less clearly separated by deposits from marine incursions or from glaciations in the Pleistocene.

The oldest phase is not represented by dated karst forms, however at Mierzyce near Wielun (M in Fig. 3) the sediments filling a buried valley cut into Middle Jurassic clay contain Late Eocene forams. This is evidence that the Middle Eocene transgression buried an uneven erosional surface of Jurassic rocks and prior to this transgression the Upper Jurassic sequence was completely eroded. It is believed that this phase includes: 1) accumulation of variegated sandy clay treated as the residue of the Upper Cretaceous overburden on the karstified surface of Upper Jurassic limestone; and 2) the residual hill landscape (Fig. 2) of Jurassic limestone treated as tropical karst mogotes.

The second phase of karstification began with the Early Oligocene regression and was caused by considerable uplift of the continental margin facing the Paleogene Carpathian basin. This phase was terminated by Middle Miocene (Badenian) transgression. A deep valley pattern was carved at that time and is preserved beneath the Badenian deposits in the southern part of the CWU. Many caves and deep sinkholes filled with the variegated sandy clay mentioned above, and vertical caves developed and were buried under Middle Miocene deposits. The karst channel and cavern porosity probably was developed in the whole thickness of Upper Jurassic limestone and filled-up with internal sediments partly derived from the surface in consequence of the Middle Miocene transgression.

The subsequent phase of karstification started with the latest Miocene regression (Messinian dessication) and was accelerated by the Pliocene uplift that caused the erosion of Miocene deposits and the exhumation of older karst features. This phase of karstification lasted until the Middle Pleistocene glaciation which probably blocked karst porosity with permafrost.

The karst relief and karst ground-water paths were mostly "fossilized" during the glaciations. However, there are some prominent effects of the meltwater and interglacial karstification. Erosive action of glacial meltwater is represented by the development of Szachownica Cave in the northern part of the CWU (Glazek et al., 1977; S in Fig. 3). This cave system has an aggregate passage length of about 1 km. It developed in a fracture network during the melt of the Wartanian Glacier on the top of a massive limestone hill, which was nearly buried beneath outwash sands.

The glacial epoch was a time of considerable reorganization of surface and subsurface outflow. The eastward outflow to the Black Sea along the axis of the Carpathian Foredeep was diverted and the northern outflow to the Baltic Sea appeared in the Middle Pleistocene. The present-day river network was then organized by captures and creation of gaps (e. g. the Warta River gaps). This change caused many buried valleys and the reorganization of karst ground-water circulation. This resulted in the filling of many cavities with Upper Tertiary or Pleisotcene deposits which are encountered in wells and caves.

The surface karst sculpture is diversified and depends on lithology. Cyanobacteria-sponge bioherms and their detrital taluses form many residual hills or towers, whereas layered micritic limestone forms a smooth landscape. It seems that the extreme karstification of biohermal limestone caused an almost complete dissolution of some bioherm cores (Fig. 2). The limestone towers are hardened by a crust in which primary porosity is filled with reprecipitated calcite from capillary solutions. These hardened crusts are destroyed below the outwash sands and the original level of the sand cover is frequently marked by a lower portion of tower walls devoid of crust. These crusts were probably created in the dry climate of the Messinian epoch, when the exhumation of hills from below thin Middle Miocene deposits was completed.

Numerous sinkholes and karst depressions completely filled with variegated Tertiary and/or brown Quaternary deposits are in all types of limestone. Less frequent are active sinkholes, but they are common in the Upland where the Cenozoic overburden is not thick. Frequently they are reproduced in a flat surface of outwash sands (Glazek et al., 1977).

Underground sculpture visible in caves and quarries is also influenced by lithology. About 95% of the caves are in biohermal limestone, while only 5% are in layered detrital limestone. The cave passages are developed exclusively along the joint network and seem to be independent of the present valley terraces (Gradzinski, 1962). In the layered limestone, karstification is disseminated along numerous bedding planes and joints; thus, there are many minute anastomoses along bedding planes and tiny networks of corroded joints.

Thus, both the surface karst morphology and the underground water paths are polygenic and developed in different phases of karstification. They were several times filled with sediments and re-excavated. In general, since the retreat of the Warta (penultimate) Glaciation, the exhumation of the paleokarst sculpture and of underground water paths is occurring. This exhumation started earlier in the southern part of the CWU and is controlled by the elevation above the local base level formed by major rivers (Vistula in the south and Warta in the north; Figs. 1 and 3). In these areas the southern part exhibits a more advanced process of exhumation, whereas the northern part is only slightly exhumed. This difference is well illustrated by the number of known larger caves (Fig. 3).

HYDROGEOLOGY

Hydrogeological Setting

The Upper Jurassic carbonates of the CWU form an extensive, generally unconfined aquifer. The preserved thickness of limestone ranges from less than 1 m close to the southwest erosional boundary of carbonate sequence to 200-550 m on the northeast boundary with Cretaceous rocks and to 700 m in the northeast corner of the area (Fig. 3). Investigations of numerous boreholes and geophysical measurements proved that karst channels filled with water and/or sediments occur in the whole Upper Jurassic sequence. In spite of different hydraulic properties of particular lithologies in this sequence, the karst aquifer in the Upper Jurassic limestone forms one flow system in the southern part of the region. The occurrence of marly intercalations in the upper part of the stratigraphic column makes possible the existence of several local water-bearing horizons in the northern part of the region.

The storage capacity of the bedrock depends on fracture and karst porosity (Rozkowski et al., 1985). Wells draining water from this aquifer are characterized by a large range in capacity from 0.4 to 567 m^3/hr at drawdowns from 0.11 to 26 m. The specific capacity of wells ranges between 0.1 to 416 m^3/hr m. The highest yield was obtained in karstified fracture zones and in the zones of hydraulic connection of the karst aquifer with the river valleys.

Three hydrogeological subregions (Cracow, Czestochowa, and Wielun) can be differentiated in the CWU. The aquifer in these subregions is characterized by differences in thickness, lithology, degrees of removal of karst-filling deposits, and, in consequence, recharge, flow, and drainage conditions. Optimum hydrogeological conditions occur in the central part of the CWU, that is, in the Czestochowa subregion where the specific capacity of about 50% of the wells is more than 10 m^3/hr m and hydraulic conductivity range from 1.5×10^{-5} to 6.5×10^{-3} m/s. In the southern subregion, Cracow, and the northern one, Wielun, specific capacity is lower and hydraulic conductivity ranges between 1.0×10^{-7} and 1.15×10^{-5} m/s.

The aquifer of the CWU is typical of karst-fissure, due to the occurrence of numerous springs of variable yield; even the yield of efficient springs decreases during the dry period, some of them even disappear. According to Kleczkowski (1972) yields of springs range from 0.5 to 187 L/s. Springs with high yields are in karstified fracture zones and on lithological boundaries. Springs with the highest yield, which are relatively few, occur in the Czestochowa subregion at the lithological boundary of massive or chalky limestone. Their yields vary between 40 and 187 L/s (Kleczkowski, 1972). The Cracow and Wielun subregions are characterized by the presence of many small springs with yields 2 - 7 L/s, while only few springs revealed higher yield.

The temperature of underground water is 7.2 to 11°C. Mineralization (TDS) varies between 80 and 500 mg/L, reaching 750 mg/L in the case of polluted waters. Waters of Ca- HCO_3 type dominate, with their pH varying between 6.8 and 8.1. The total hardness ranges from 1.5 to 6.5 epm and is close to the calcium hardness. The presence of dissolved CO_2 (4 - 40 mg/L) and O_2 (0.8 - 10 mg/L) testifies to open flow routes and an oxidized environment (Rozkowski and Pacholewski, 1988).

Hydraulic properties of the Jurassic limestones

Hydraulic properties of the Upper Jurassic rocks have been studied in the representative catchment area of the Wiercica River. The overall permeability (hydraulic conductivity) of the rock mass was determined by aquifer tests. Results of 42 tests showed the mean hydraulic conductivity to be 1.14×10^{-4} m/s with a minimum of 1.3×10^{-7} and maximum 6.19×10^{-3} (Pacholewski, 1982). Zone test pumpings revealed a decrease of permeability with depth. This may be explained by the diminishing karstification of layered limestones below the zone of maximum development of bioherms (Fig. 3) and by a lesser extent of exhumation of infilled karst channels in the deeper parts of the aquifer. The specific yield determined by a few pumping tests ranges between 4.1 and 8.0%, while the regional specific yield calculated on the ground of piezometric level fluctuations and the underground runoff equals 9.9% (Pacholewski, 1984).

The hydraulic properties of matrix porosity were studied in the laboratory. Drill cores collected from a few boreholes were examined for the open porosity coefficient, permeability coefficient and specific yield. The overall permeability and transmissivity of a karst aquifer is an aggregate value of matrix porosity, fracture porosity, and channel and cavern porosity. Thus, laboratory measurements with a mean of 1.7×10^{-7} m/s (range 9.6×10^{-10} to 1.7×10^{-6}) represent only a fraction of the total permeability, which in the case of the Wiercica catchment area is one thousand times greater. The matrix porosity of 9.29% (range 0.62 to 29.78%) was measured in the laboratory. The fracture porosity was estimated in the field as the ratio of the fissure volume to the total volume of rocks in the investigated portion of the outcrop (Liszkowska and Pacholewski, 1989). The specific yield mean for 90 measurements is 1.41% with a range from 0.1 to 6.55% The karst porosity was thus obtained as the difference between the overall permeability and the sum of the measured matrix and fracture porosity in different groups of limestone.

These investigations demonstrate the hydraulic heterogeneity and anisotropy of the Upper Jurassic rocks forming the karst aquifer. The karst porosity represents only a fraction of a total karst voids, which are partly filled with sediments, and therefore prolonged pumping and removal of this loose material results in a considerable increase of the estimated karst porosity. Thus the inner hydraulic system of the karst aquifer is composed of pores, fractures, open karst voids, and filled karst cavities which would be opened during pumping as a result of human intervention into the paleokarst system which was in a nearly stationary state (Glazek and Szynkiewicz, 1979).

Different hydraulic properties of particular rock types are responsible for the heterogeneity of the karst aquifer in the Upper Jurassic limestone of the CWU. The high storage capacity of the aquifer is due to matrix and karst porosity. The highest storage capacity was found in chalky limestone, while the lowest value occurs in layered limestone, although it revealed the greatest density of joints. Because of their low flow resistivity, fractures and karst conduits appear to be the main flow paths of ground water. They drain the rock complex, while matrix porosity plays a secondary role in the transmission of water due to the high flow resistivity. The greatest karst porosity occurs in massive limestone.

Similar results were obtained by test pumping in a few wells. Representative hydraulic conductivity of layered limestone were found to be around $3.2\text{x}10^{-5}$ m/s, while these of massive limestone ranged from $7.6\text{x}10^{-4}$ to $2.6\text{x}1^{-3}$ m/s, and of chalky limestone from $8.0\text{x}10^{-5}$ to $6.4\text{x}10^{-4}$ m/s. These figures show that the massive limestone, in spite of having the lowest matrix porosity, demonstrates the greatest transmissivity as a result of its highest degree of karstification.

Ground-Water Dynamics

The flat ridge of the CWU forms an elevated watershed area. Water flows from this ridge to the northeast and to the southwest (Figs. 3 and 4). The Upper Jurassic limestone of the CWU forms a shallow karst aquifer with a mainly unconfined flow regime. Confined flow conditions occur where the limestone is covered by Quaternary boulder clay or Tertiary clay and down the dip of the Jurassic strata, on the northeast slope of the Upland where the limestone is overlain by Upper Cretaceous marl (Fig. 4). Block-fault tectonics divide the aquifer into separate hydraulic units with separate, but connected, flow systems.

The potentiometric level ranges in depth from a few meters to about 50 m below the surface. Seasonal fluctuations range from a fraction of a meter to tens of meters. The recharge of the karst aquifer takes place in the whole area, directly where limestone crops out, or indirectly through permeable Quaternary deposits. The exception is at the northern and southern boundaries of the CWU, where the limestone is covered with the Tertiary sediments. Moreover, hindered recharge occurs in small areas where limestone is covered by Quaternary clay.

Daily measurements of precipitation, stream flow, and water levels make it possible to determine the response time of the aquifer to rainfall and to snow melt in spring. The retardation depends on the Quaternary cover and the extent to which the rocks are jointed and karstified. The delay of aquifer response to recharge is about 2-3 days in bare karst and 2-4 weeks when the limestone is covered with a sand layer. In places where the aquifer is covered by a several-meter-thick layer of Quaternary or Tertiary clay and/or till the response time to spring snow melting amounts to a few months.

Tritium measurements have demonstrated that the percolation time of atmospheric water through 16 m of Quaternary sand and loam to the karst aquifer can be a few years (Rozkowski et al., 1985). The measurements in the representative catchment area of the Wiercica River (200 km^2) revealed the response time of this basin to melting of snow in spring to be between 3 and 6 months. A tritium survey carried out in the Rudawa and Pradnik catchment areas, near Cracow, in 1968 showed the ^3H content in ground water to be from 30 to 306 tritium units. Considering the ^3H content in atmospheric water in the years 1961-1967, these values indicate various residence times. Hence, the residence time of water in this aquifer depends on the karstification, the density of joints, the length of flow paths, and recharge conditions.

Ground water in the CWU aquifer circulates along the karst conduits and joint systems. Tensional fault zones are the preferred flow paths. Considerable differentiation of flow velocities has been observed. The mean flow velocity within Wiercica catchment area is 0.12 m/hr, while in the fault zones it attains 2-5 m/hr (Pacholewski, 1982). Turbulent flow occurs only in karst conduits and fault zones, and the water flow in the region is generally laminar.

The development of the recent flow system took place after the lowering of base level in the Late Tertiary dessication during the Messinian and uplift during the Pliocene. Both local and intermediate flow systems were the result of distensional faulting and deep valley incision prior to continental glaciation. The regional flow system has developed as a result of the monoclinal dip of Jurassic strata (Rozkowski et al., 1985; Rozkowski and Pacholewski, 1988). The underground watershed is generally shifted to the southwest with respect to the surface culminations (Fig. 4). The karst aquifer of the CWU is drained mainly by intermediate and local flow systems. Regional runoff does not exceed 20% of total underground runoff. Intermediate flow systems are a few hundred square kilometers and are drained by the main rivers of the region. Recharge takes place mainly in hilly areas which is evidenced by the downward decrease of total head from 0.006 to 0.1 m/m. Drainage areas are characterized by an upward increase of the total head from 0.003 to 0.1 m/m. The configuration of the potentiometric surface roughly corresponds to the surface morphology of the area, hydraulic gradients varying from 4.0 to 11 ‰.

The local flow systems have a catchment surface ranging from a few to fifty square kilometers (Rozkowski and Stachura, 1971). The distribution of the total head field of the local flow systems points out that the lateral and vertical downward flows occur in hilly areas. The decrease of the total head in the hilly area is about 0.07 m/m. The spherical distribution of the equipotential lines and upward flow are typical of stream valleys. The formation of springs at these valley bottoms directs the water flow along the paleokarst conduits. The vertical upward flow is observed within the stream valleys even deeper than 50 m from a valley bottom. The drainage area of local flow systems occupies about 20 to 30% of the local basin surface. The distribution of the total head and differentiation of the values of total head gradients in the recharge and discharge areas in relation 7:1 indicates that the recharge area of the local drainage basin is not totally drained by a valley. This shows clearly that both intermediate and local flow systems appear in the area.

CONCLUSIONS

The Upper Jurassic carbonate sequence of the Cracow-Wielun Upland is a karst-fissure aquifer with typical relief, scarcity of surface water, and abundance of groundwater. The surface karst morphology and the underground water paths developed in different phases of karstification. The Upper Jurassic carbonates which make up the unconfined aquifer display a complicated pattern of porosity and karstifiability. The overall transmissivity of the karst-fissure aquifer is an aggregate value of matrix porosity, fracture porosity, and channel and cavern porosity. Karst conduits and fractures appear to be the main flow paths of ground water. The development of recent flow systems took place after the Late Tertiary lowering of base level. The coefficient of underground runoff amounts to 85% of the total runoff, while the effective infiltration ranges from 28.4 and 48.3% of the total precipitation.

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THE HYDROGEOLOGICAL CHARACTERISTICS OF POLISH KARSTIC AREAS

JAN MALINOWSKI
State Geological Institut
00-975 Warszawa
ul. Rakowiecka 4
Poland

INTRODUCTION

Karst occurs in a few physiographic-geological regions of Poland (Fig. 1) which are developed in the calcareous rocks of Devonian, Triassic, Jurassic, and Cretaceous Periods and in the gypsum rocks of the Tertiary. In some regions they are an important hydrogeological regulator of underground runoff and ground water flow. In other regions they are passive and help feed the main water-bearing layers. The hydrogeology of some karst regions has not been studied adequately. The most extensive hydrogeological study of karst has been achieved in the Tatra Mountains (area 1a in Fig. 1), on the Cracow-Czestochowa Upland (area 8 in Fig. 1), and Lublin Trough (area 5 in Fig. 1). The hydrogeological recognition of karst is considerably less in the remaining regions: Swietokrzyskie Mountains (area 6 in Fig. 1), Sudety Mountains (area 2 in Fig. 1), Nida Basin (area 7 in Fig. 1). Karst does not occur in Carpathian foreland (area 4 in Fig. 1), Sudetic foreland (area 3 in Fig. 1), Polish Lowland (area 10 in Fig. 1), or Silesia Upland (area 9 in Fig. 1) (Gradzinski et al., 1975).

KARST OF THE TATRA MOUNTAINS

Widespread development of karst occurs in the Tatra Mountains (area 1a in Fig. 1) in part of Poland and Czechoslovakia. The Tatra Mountains are a small structural element within the Carpathian Mountains of the alpine orogeny. They are composed of crystalline rocks, such as granite and gneiss, and Mesozoic-aged formations, including limestone, dolomite, sandstone, and conglomerate (Figs. 2 and 3). The area of the Tatra Mountains is 715 km^2, of which about 180 km^2 are in Poland. The Tatra Mountains are high, from 750 to 2,350 m, with rugged landform and abrupt slopes. In the years 1951-1980, mean annual precipitation on the mountain summits ranged from 1,724 to 1,937 mm and at the foot of a hill from 1,119 to 1,162 mm (Malecka, 1985).

Karst occurs in the Mesozoic limestone, and its elevation is within the range of 0-1,460 m (Rudnicki, 1967). These are surface furrows, vaucluse springs, karst funnels, and dry valleys. The underground forms include caves connected with a system of passages. The total length of caves amounts to 13 km (Kowalski, 1953). The biggest cave is 1,650 m long, the smallest one is 500 m. By 1981 about 400 caves had been registered. Caves are joined by a system of passages which are of a various length and height, however, connection between the distribution of caves, their main directions, and geological structure have not been fully mapped. Some caves are parallel to the valleys of large streams which have developed along berms.

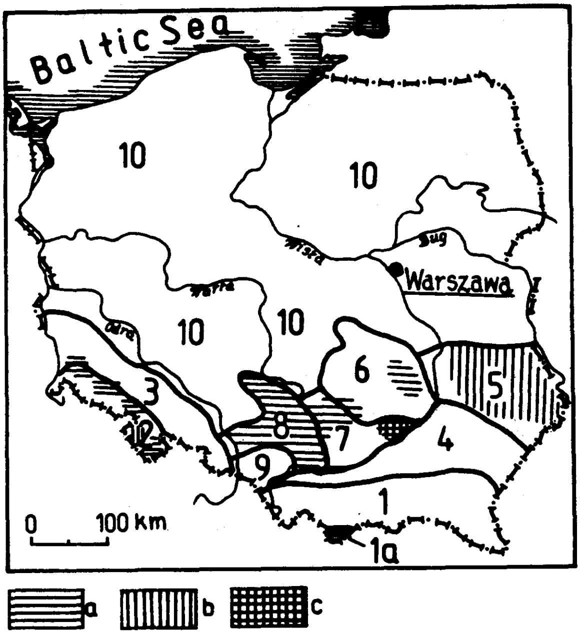

Figure 1. Geographic-geological regions of Poland. 1. Carpathian Mts., 1a. Tatra Mts., 2. Sudety Mts., 3. Sudety Foreland, 4. Carpathian Foreland, 5. Lublin Trough, 6. Swietokrzyski Mts., 7. Nida Basin, 8. Cracow-Czestochowa Upland, 9. Silesia Upland, 10. Polish Lowland, a-limestone karst, b-pseudo limestone karst, c-gypsum karst.

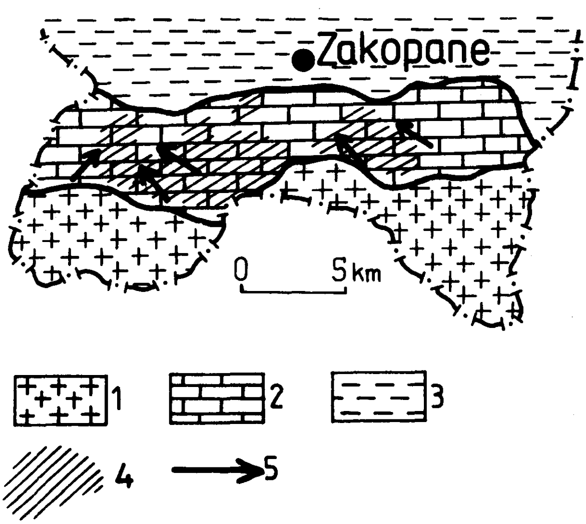

Figure 2. Geological sketch of Tatra Mountains according to Sokolowski (1959). 1. Granite and gneiss, 2. Limestone, dolomite, conglomerate, sand, and slate of the Triassic, Jurassic, and Cretaceous, 3. Flysch of the Tertiary, 4. Karst area, 5. Most direction of karst water flow.

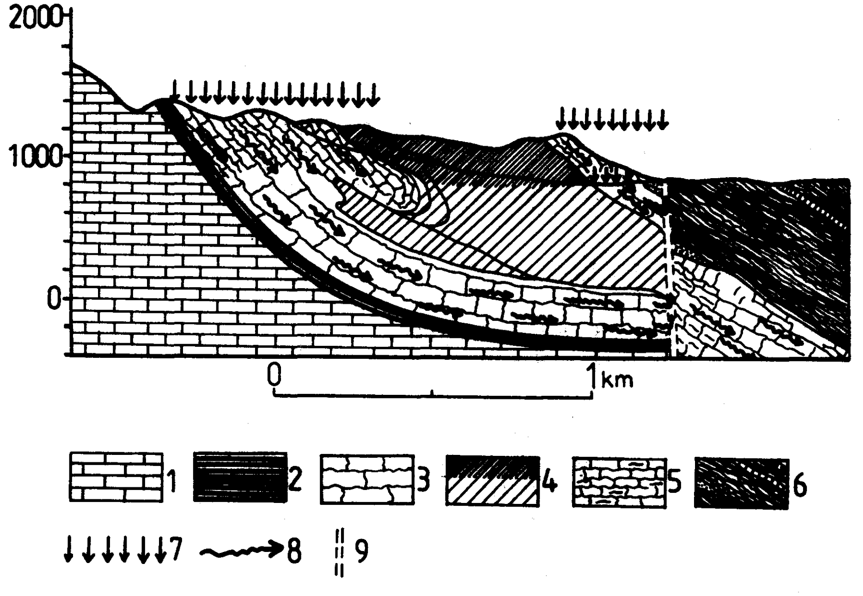

Figure 3. General cross section of the north zone of Tatra Mountains according to Sokolowski (1959). 1. Limestone, slate, and sandstone of the Triassic, Jurassic, and Cretaceous, 2. Slate of the Lower Triassic, 3. Limestone and dolomite of the Middle Triassic, 4. Slate and sandstone of the Lower Jurassic, 5. Limestone of the Triassic, 6. Conglomerate and slate of the Tertiary, 7. Infiltration, 8. Direction of water flow, 9. Faulted zone.

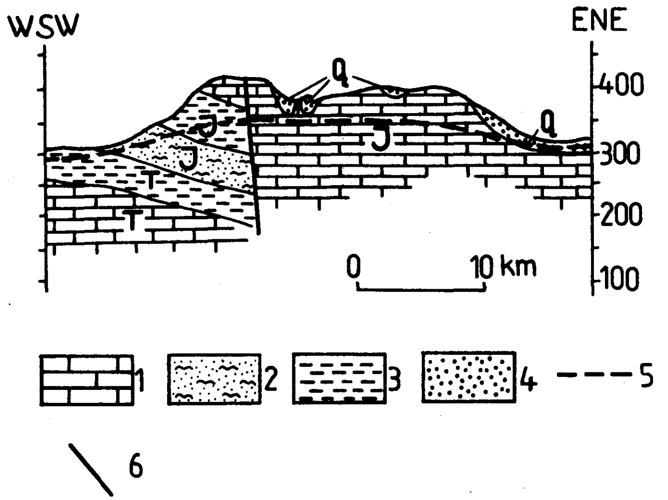

Figure 4. Schematic cross section through the Triassic and Jurassic rocks of the Cracow-Czestochowa Upland. 1. Limestone, 2. Sandstone and slate, 3. Slate, 4. Sandstone, 5. Ground-water level, 6. Fault. Q-Quaternary, J-Jurassic, T-Triassic.

The dynamics of water flow through the karst system have not been fully recognized yet. Glazek and Wojcik (1963) make a hypothesis that there are two zones of flow. A deep zone fed with precipitation in upper parts, and drainage is by springs in the northern edge of the Tatras. The shallow zone is able to gather water from direct infiltration, and drainage is by karst vaucluse at the bottom of the valley. That hypothesis is confirmed by Rudnicki (1967), who expresses the opinion that water of both zones can join in the outlets. An explanation of this problem seems to be the most important question about the hydrogeology of the Tatra Mountains.

KARST OF THE CRACOW-CZESTOCHOWA UPLAND

Cracow-Czestochowa Upland (area 8 in Fig. 1) is the second region in which karst phenomena occurs with great intensity. The upland covers an area of 10,258 km^2 in the Middle Polish uplands (areas 5, 6, 7, 8, and 9 in Fig. 1). Relief of the rolling upland is within the altitude of 300-500 m. The river system is weakly developed. Geological structure is simple. On the Palaeozoic base (Devonian and Carboniferous) lie fractured, homoclinal formations of both Triassic and Jurassic age which include limestone, dolomite, conglomerate, and sandstone (Fig. 4).

Karst has developed strongly in the carbonate formations of the upper Jurassic and less strongly in the formations of the Triassic. Karst in the formations of Jura period occur on the surface as sink holes, small valleys, and grooves (Kowalski, 1951-1954). However, at various depths there are caves, often large and connected with a system of passages. Karst originated in two cycles: Cretaceous, in which the erosion took place and was followed by the rising of big caves in which are various deposits. The younger cycle, Tertiary period, left smaller caves deeply lying with no deposition of sediments.

The presence of intensively fractured zones, along the main tectonic directions was favorable to the development of those processes. A karst system in the region is entirely exposed and there are favorable conditions for infiltration. The karst system occurs above the water table. In a few caves small lakes occur. Their depth was 1 m and water appeared only periodically. Caves, with the system of joining passages, are transit ways for the infiltrating waters which flow to the saturation zone, occurring at the depth of 20-80 m and in places to 200 m. Some investigators are of an opinion that a certain part of caves and passages occur below the water table. In such a case are favorable conditions for water retention what is proved by the existence of karst springs of high discharge over 200 m^3/s (Roskowski et al., 1985). In limestone and dolomite of the Triassic period, karst has developed differently. There are smaller caves in shell limestone not connected so intensively as in the Jurassic formations.

Motyka (1988) discovered in the southern part of the region 955 small karst caves with various dimensions. They play the role of internal drains in relation to the surrounding blocks of rocks. If such caves are filled with clay depositions, hydrodynamic barriers appear and they locally change the water runoff and even create local watersheds.

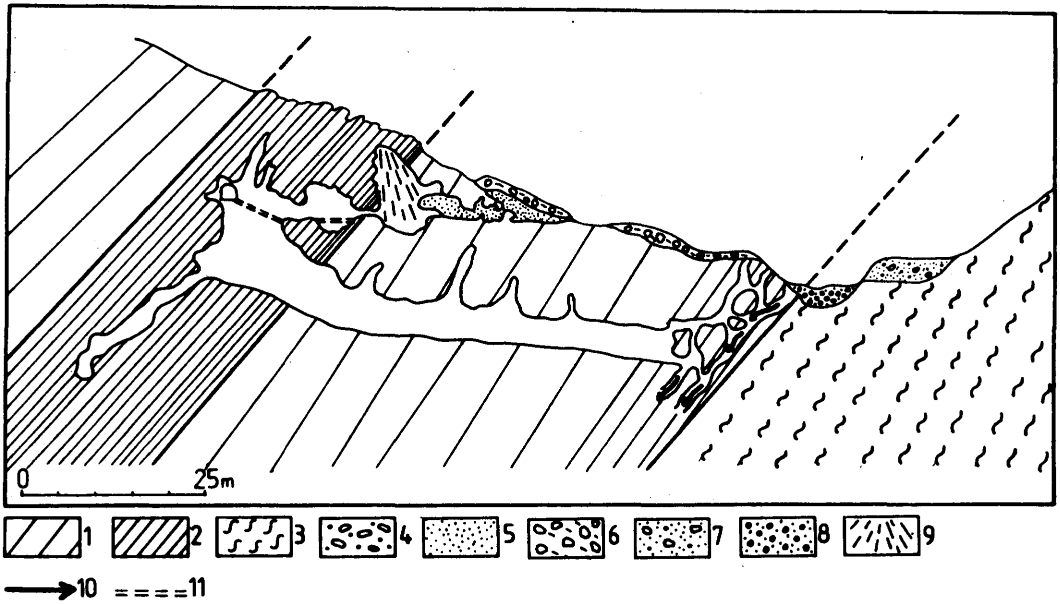

Figure 5. Cross section through the Bear Cave Kletno, the East Sudety Mountains, according to Kozlowski (1989). 1. Marble and dolomite, 2. Marble with calcite and dolomite, 3. Mica-schist, 4. Gneiss gravel, 5. Cave deposits, 6. Cover of weathering, 7. Gravel of older terrace, 8. Gravel of younger terrace, 9. Talus cone, 10. Direction of the river-water infiltration, 11. Level of tourist routes.

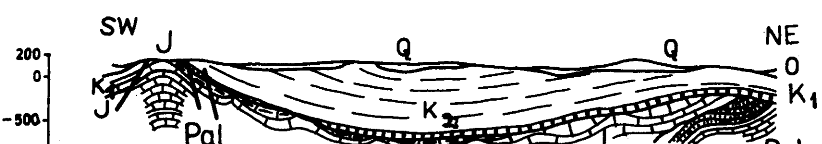

Figure 6. Geological cross section through the Lublin Trough according to Ksiazkiewicz et al. (1965). K_2-Upper Cretaceous, K_1-Lower Cretaceous, J-Jurassic, Pl-Paleozoic.

Due to the regular position of the water table in the whole area, numerous water intakes are being built. They achieve a considerable discharge of 30-120 m^3/hr, most often 70-120 m^3/hr and in the zones of intensive flows even 270 m^3/hr, at depressions ranging from 0.15 to 26 m. The water table occurs on ordinates at altitudes of 400-220 m, which is equal to the depth of 40-100 m from the level of the surface, and slopes in the direction of north-northeast, which is the regional direction of the water runoff.

The presence of karst system affects the water balance of small catchment areas. Obviously, it cannot be defined in what way the fractured karst system influences the water balance. Many investigators are of an opinion that control on water flows are hydrodynamic barriers which include impermeable rocks, such as siliceous rocks, or cave filling (Rozkowski et al., 1985; Motyka, 1988). Then the stream of water is divided in various directions, the hydraulic gradients usually increase and, water-flow changes are reflected in water balance of the catchment area. Considerable and violent inflows occur in the open-pit mines of brown coal. It is assumed that karst interstices act as drains gathering water, similarly to the carbonate formations of the Triassic period. Disturbing such interstices by mining produces significant water inflows. For example in the Belchatow brown coal mine inflows from the Mesozoic karst bedrock amounted to 32,790 m^3/hr in 1985.

KARST OF THE SUDETY MOUNTAINS

Karst features in the Sudety Mountains are typically developed (area 2 in Fig. 1) but have only local character. They occur in the crystalline limestone of Cambrian age in the form of lens among gneiss or other metamorphosed rocks (Fig. 5). Karst forms include swallow holes, vaucluse, and caves often with large dimensions. Hydrogeological role of karst is well demonstrated in the East Sudety Mountains. By systematical measurements it has been discovered that there are tight correlations between the river flow and the flow in the system of caves and passages through which water is continuously flowing. The excess of water outflows in vaucluse and where swallow holes occur, through which water feed the karst system (Fig. 5). Although the capacity of reservoirs is rather low, high precipitation causes the feeding to be permanent. Investigations carried out by Ciezkowski (1989) by means of dye tracers proved that water of karst systems are feeding various surface catchment areas, which are manifested by high values of the underground runoff.

Hydrogeological influence of karst phenomena in the Sudety Mountains is significant. In Sudety, 29 caves have been registered at the level of 260-280 m. The biggest cave, Kletno (Fig. 5) is 2 km long and 50 m high, is a tourist attraction, and the highest cave in the Sudety Mountains at 800 m.

KARST OF EASTERN POLAND

In the east, the karst intensity is less, and features have rather local character where their influence on the creation of hydrogeological conditions is low. They are best developed in the Swietokrzyskie Mountains (area 6 in Fig. 1). Karst phenomena occur in limestone of Middle and Upper Devonian. Karst phenomena in limestone of the Upper Jurassic period are weakly developed. The present stage is described as pseudo-karst whose structures are in the form of surface erosion and partly erosion between fractures of small intensity. In the Southern part of the Swietokrzyskie Mountains, gypsum karst occurs in form of small caves and interstices. All these forms occur above the saturation zone and their role in creation of hydrogeological conditions is passive. They are pathways of transit flow infiltrating the saturation zone. Water gathers in the fractured zones at the depth of 40-100 m and has considerable flows. It is visible in the discharge of ground water intakes, which usually amount to 70-120 m^3/hr and often 150-200 m^3/hr. Few karst springs have discharge ranging from 1 to 30 L/s (Czarnecka, 1975).

In the eastern part of Poland the pseudo-karst occurs in the Lublin Trough covering 10,000 km^2 (area 5 in Fig. 1) (Malinowski, 1989). That Trough is filled with formations of the Upper Cretaceous with thickness of about 1,000 m (Fig. 6). Karst processes started to develop on the surface in the formations of Maastricht, and probably in the last stage of the Tertiary period; they stopped in the glacial period. Karst forms were created as narrow erosion features of the top surface filled with subsidiary deposits of varying grain size. The erosion is partly attributed to flowing waters, which led to the development of deep valleys, 50-70 m deep and several kilometers long. The other aspect of karst activity is dissolution along the main directions of fractures.

The area of the Lublin Trough provides favorable conditions of infiltration because the cover of the Quaternary formations is thick and generally penetrable. Dewatering is intensive and the influence of precipitation can be noticed quickly. Favorable conditions of flow are reflected by high filtration coefficients. Values above 10 m/day are characteristic for the zones intensively fractured with karst scours. Existing scours, in the form of valleys, indicate underground flows form one catchment area to the other. By dewatering the central area of Trough, the Wieprz River (Fig. 1) shows in its mouth greater discharge than infiltration.

DISCUSSION

Ground water of the discussed karst areas have low mineralization ranging from 150-300 mg/L in Tatras (area 1a in Fig. 1) and Cracow-Czestochowa Upland (area 5 in Fig. 1) and a little higher at 500-600 mg/L in Lublin Trough (Fig. 6). Because of the development of industry, water is polluted which is expressed by the increase of total mineralization and of some anions, especially SO_4, Cl, and NO_3. More common is the increase of heavy metal contents Pb, Hg, Cd, and also other metals. The karst areas have been classified among the areas requiring special water protection in the first place. In these areas, all technical activity harmful for the environment has been limited.

Summing up the general hydrogeological characteristics of the karst regions, the following can be stated:

1. Karst regions of the Lublin Trough (area 5 in Fig. 1) and the Cracow-Czestochowa Upland (area 8 in Fig. 1) have been well studied. In those regions, intensive

exploitation of underground water has been conducted. In the fractured-karst systems, it is difficult to determine the amount of recharge and admissible exploitation. Investigations aimed at solving this problem are the focus of research work concerning balances of small catchment areas and observations of the ground-water table. In the Tatras (area 1a in Fig. 1) mountain region, the ground water is not utilized. This region is regarded as an experimental region in order to develop the research methods.

2. Ground water of karst regions is subject to pollution through surface infiltration, and karst areas are classified among most protected areas. Different actions in order to make the protection efficient have been undertaken.

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KARST AND KARST WATER IN CHINA

YUAN DAOXIAN
Institute of Karst Geology
Guilin, Guangxi 541004
The People's Republic of China

INTRODUCTION

According to the map of soluble rock in China, bare karst totals about 907,000 km^2 (Fig. 1). Taking into account both buried and covered karst, it covers about 344.3×10^4 km^2, approximately one third of China (Li Daton, 1983). Because of available land, energy, minerals, and water resources, hundreds of millions of Chinese live in karst areas. On the karst plateau of southwest China, solutional basins comprise the most important farmland. On the upper reaches and the main tributaries of the Yangtze River, many karst gorges are the most favorable sites for constructing hydroelectric facilities. In northern China, Carboniferous coal measures, currently the major energy resource in China, are on a Middle Ordovician karst aquifer. In the Sichuan Basin and the North China Plain, reservoirs of some of China's main oil fields are in buried karst. Important mineral deposits (cassiterite, quartz, bauxite, pyrite, manganese, realgar, lead, zinc, copper, gold, and diamonds) are in karst features of various origins, such as dolines, caves, solutional fissures, solutional pipes, and paleokarst surfaces (Yuan, 1981). Additionally, many caves and different types of karst landscapes are important for developing tourism industry. As natural resources in karst areas are increasingly exploited, water resources are in greater demand. Karst water in China totals about 203.97×10^9 m^3/a, about one quarter of the total ground-water resources in the country. Because of its uneven distribution, both in space and time, karst water has particular problems with respect to its quantity, quality, feasibility of exploitation, and protection (Yuan, 1985).

Controlled by a framework of soluble rock, karst water is stored and flows in thousands of karst hydrological systems that range in size from thousands to a few square kilometers. A karst hydrological system is a system with definite boundaries, with water and air circulating through it in dynamic equilibrium. The boundaries are configured by such factors as the stratigraphic relationship between soluble rocks and insoluble rocks, geological structures, and topographic features. A better understanding of karst hydrological systems is the key to the proper estimation and rational exploitation of karst water resources while protecting the environment. Each karst spring or underground stream may be a karst hydrological system, with the precipitation on its recharge area as input and its resurgence as output. There are thousands of karst hydrological systems in China. Systems in different parts of the country differ from each other because of local geological, climatic, and geomorphological features. The aim of this short article is to give some general ideas about the basic factors that form a karst hydrological system and their variability in different areas of the country.

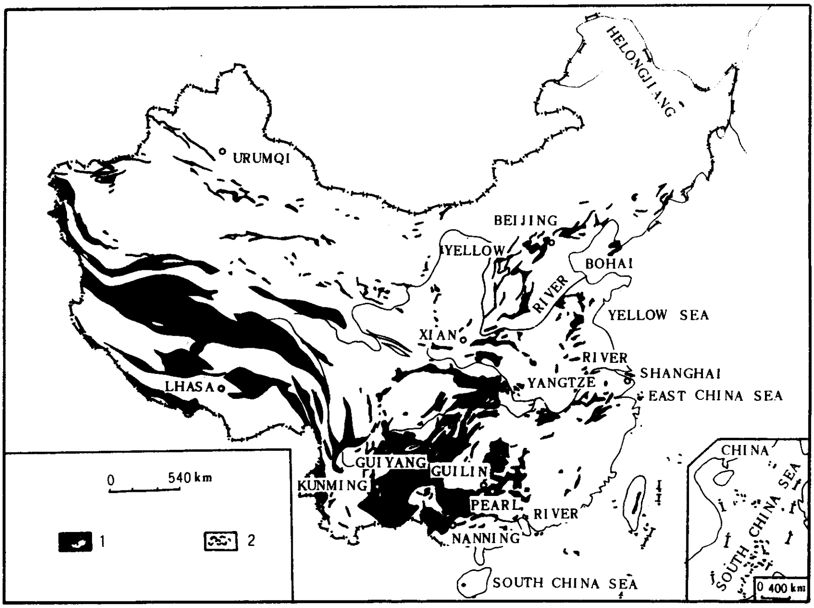


Figure 1. Map showing area of study.

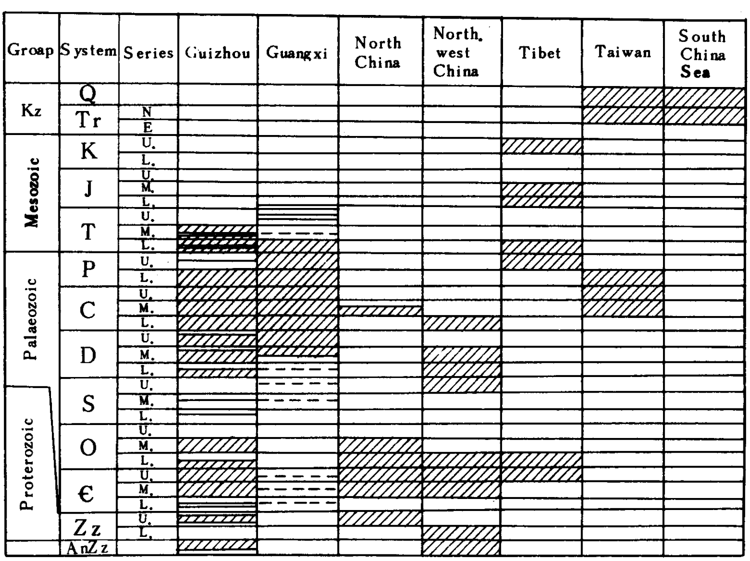


Figure 2. Chart showing stratigraphic distribution of carbonate rock in different karst areas of China. Hachured boxes indicate carbonate rock; blank boxes indicate non-carbonate rock.

GEOLOGICAL BACKGROUND

A karst hydrological system develops as circulating water and air react with soluble rock. Carbonates and evaporites are the two basic types of soluble rocks. Evaporites cover vast areas in China: many inland lakes of northeastern China, the Inner Mongolian Autonomous Region, Qinghai province, the Xinjiang Uygur Autonomous Region, the North Tibet Plateau where deep deposits of mirabilite, halite, and borax are found. In the red beds of Mesozoic to Paleogene age, which cover large areas in southern and northwestern China, are numerous deposits of gypsum and mirabilite.

Evaporites are mixed in with Middle Ordovician carbonate rocks in northern China and in Triassic and Cambrian carbonate rocks in southern China. Carbonates are the most important soluble rocks in China (Fig. 2). In the eastern part of the mainland, they are pre-Triassic, hard, compact carbonates. This compactness may explain the morphological differences between southern China's humid subtropical high tower karst and those of Central America and elsewhere. In the Guizhou karst, in southwestern China, carbonate rocks total 3,000 to 10,000 m in thickness from the Sinian (or Late Proterozoic) to Ordovician, and from Devonian to Triassic (Zhang Shouyue, 1979). This thickness and interbedding with non-carbonates, along with structural controls, such as folding and faulting, make the carbonate rock and its karst features widespread on the plateau where the dip is gentle (Fig. 4), but found only in relatively narrow belts where the strong folding and the steep dips predominate (Fig. 4). In the south-central part of the mainland (Guangxi, Guangdong, and the southern part of Hunan Province), the carbonates are stratigraphically continuous from Middle Devonian to Lower Triassic. Consequently, the karst is also continuous regardless of structure. In northern China, carbonate rocks extend from Proterozoic to Middle Ordovician with a total thickness of 1,000 to 5,000 m. These flat-lying rocks cover large areas and give rise to many major karst hydrologic systems, each with an area of hundreds to thousands of square kilometers.

Tibet is characterized by Jurassic and Cretaceous limestone from the ancient Tethys Sea. Permian carbonate rocks are also found in some parts of Tibet. The limestone outcrops on the summit of Jumulangma (Mt. Everest) are Ordovician. In Taiwan, Permo-Carboniferous carbonate rocks are found in the northern part of the Central Ridge, along the Su'ao-Hualian Highway. Cenozoic carbonate rocks are distributed on the islands in the South China Sea and some coastal areas of Taiwan and Hainan Island. Lacustrine freshwater carbonate rocks are reported in some places on the mainland. Some Mesozoic and Cenozoic conglomerates, with calcareous cement are also soluble and show karst features such as towers, caves, and underground streams. There is more dolomite in the Devonian and middle Carboniferous systems in southern China and in the Cambrian and Ordovician systems in both the south and the north.

There have been several phases of karstification during China's geological history. The most obvious paleokarst is in 1) the Caledonian Orogeny disconformity between the Ordovician limestone and middle Carboniferous Series in northern

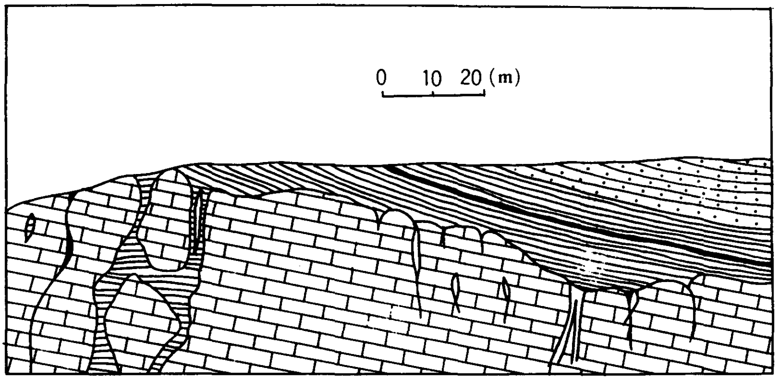
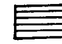
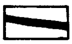


Figure 3. Cross section showing pre-Carboniferous paleokarst in northern China. This figure illustrates an example from Yaoxian County, Shaanxi Province. The Permian sandstone  overlies the carboniferous shale and bauxite  which contain the coal measures  and fill the paleokarst features.

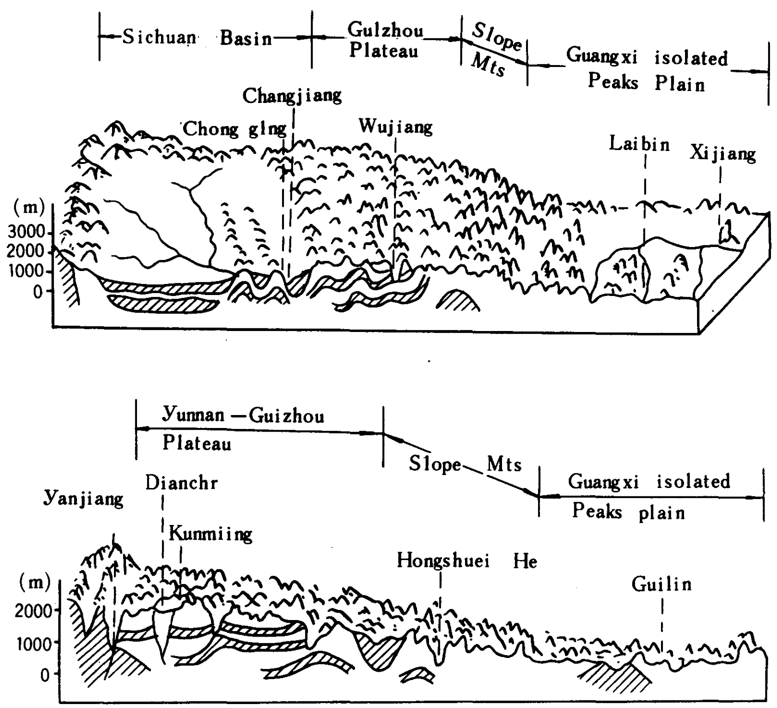


Figure 4. Schematic geomorphologic profiles across main karst regions in southern China. Hachured pattern indicates carbonate rocks.

China (Fig. 3), 2) the disconformity between the Lower Permian limestone and the Upper Permian Series in southern China, 3) the pre-Jurassic karstification in the Sichuan Basin, and 4) the pre-Tertiary buried karst in northern China. The effects of paleokarst can be noted in the various karst features in the carbonate aquifers, the yield of the oil and gas reservoirs, and the direct contact between important mineral deposits, such as coal, bauxite, and pyrite, with the karst aquifer, which complicates mining engineering.

China's karst can be classified into three categories: buried karst, covered karst, and bare karst. These types are characterized by the difference in the openness of the hydrologic system. Karst features that occur in soluble rocks underlying non-soluble bedrock are called buried karst. Hydrologic systems in buried karst areas are usually semi-open or semi-closed. The major buried karst areas in China are the Sichuan Basin and the North China Plain. In the Sichuan Basin, because of the subsidence during the Mesozoic Orogeny, karstified carbonate rocks of Proterozoic to Triassic age are overlain by hundreds to thousands of meters of Mesozoic red beds. Caves in Permian or Triassic limestone 4 to 5 m in diameter are frequently encountered in oil and gas wells at depths of 1,000 to 3,000 m. The North China Plain is a result of neotectonic subsidence. Carbonate rocks of Proterozoic to Middle Ordovician are buried under thousands of meters of Cenozoic strata. Caves found in deep wells are not rare. For example, in Huanghua County, Hebei Province (about 90 km to the south of Tianjin City), during drilling in Ordovician limestone at a depth of 1,000 m below sea level, a cave was encountered and a blow out followed, with a large quantity of stalactite pieces carried in the outburst of ground water. Karst water in buried karst usually has high total dissolved solids (100 to 300 g/L), and high temperatures (about 100°C), indicating that the system is relatively closed.

In covered karst, karst features are in soluble rocks overlain by unconsolidated sediment; the thickness of the loose overburden is usually 10 to 50 m. Covered karst in dolines, poljes, and solutional plains comprises the major arable land in the southern China karst. For example, in the Caoba area in southern Yunnan, 32,400 ha of farmland lie in karst basins on a plateau (Fig. 10). From a regional viewpoint, covered karst is usually a hydrologically open system.

Bare karst is the most conspicuous type of karst in China. Soluble rocks are exposed in bare karst; the karst features are subaerial or at least have a subaerial entrance. Major bare karst is in the uplift areas, and has generally a NNE trend in the eastern part of the country (Fig. 1) as controlled by the Neo-Cathaysian Tectonic System. China's topography, the result of differential neotectonic uplift, is characterized by a stepdown in altitude from west to east, from the Qinghai-Xizang (Tibet) Plateau (higher than 4,500 m above sea level), to the Yunnan-Guizhou Plateau in southern China and Shanxi Plateau in northern China (1,000 to 2,000 m above sea level), and then to the north China Plain and southern China low hilly land (0 to 200 m above sea level). This topography has a strong influence on climatic, hydrodynamic conditions and the development of karst. On the Tibet Plateau, where the strong uplift gave rise to a relatively cold and dry Quaternary, karst is not well developed, although there is an extensive area of carbonate rock. Minor karst forms such as pinnacles and natural bridges on the plateau are thought

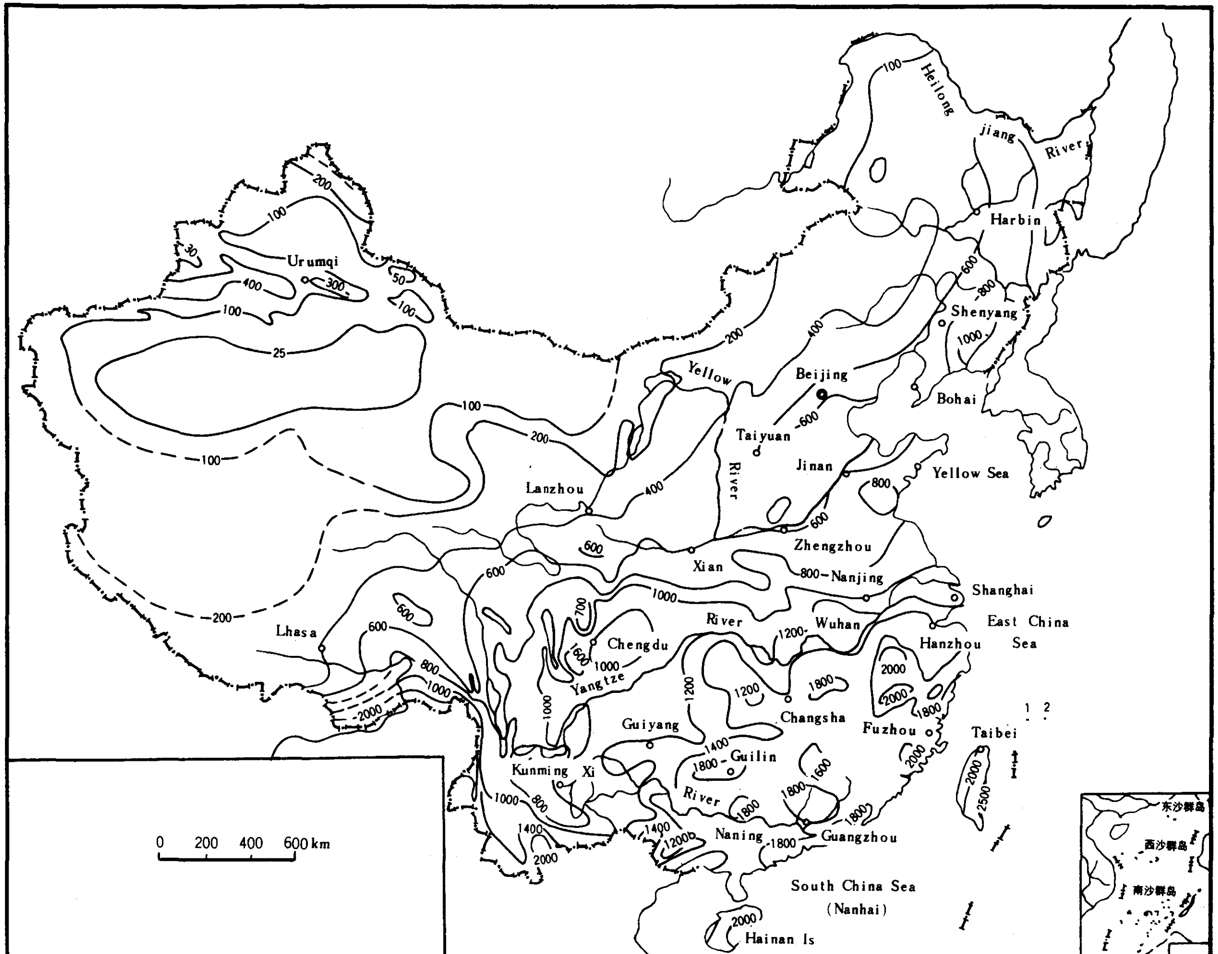


Figure 5. Map showing annual mean precipitation in China.

to be periglacial in origin and related to frost process. Calcareous tufa around some karst springs is found. Known caves in Tibet are less than 150 m in length. Whether there are Tertiary warm humid karst features in Tibet remains a controversy.

The Yunnan-Guizhou Plateau has had a humid climate after uplift. It is deeply dissected by the surface drainage system of the Changjiang River (Yangtze River) on its north side and the Pearl River (called the Hongshui River in the upper reach) on its south side (Fig. 4). The hundreds to thousands of meters of relief brings about an active hydrodynamic system which is most favorable for karst development. Many underground streams and cave systems occur in this area (Fig. 7). As a result of regional uplift, some higher cave passages, dated by paleomagnetism, ^{14}C, uranium-series, and cave fossil studies, have proven to be very old. For instance, fossils of the early to middle Pleistocene primate Gigantopithecus were unearthed in the Guangxi and Hubei Provinces, all from caves 80 m higher than the local plains. In southern China, in many cave passages some 20 to 30 m higher than local plain, speleothems are frequently found older than the limitation of uranium series dating (350,000 a BP). In downtown Guilin, a speleothem from Nan Cave about 23 m higher than the Lijiang flood plain, shows a paleomagnetic reserved inclination, which probably dates back 900,000 a, and gives an uplift rate of not more than 23 mm/1000 a (P. W. Williams, pers. comm., 1986).

In the bare karst around the Shanxi Plateau in northern China, well developed dry valleys are usually the recharge areas of major karst springs, which emerge at the boundaries between limestone mountains and plains or valleys (Fig. 6).

The eastern section of the southern China karst is a hilly lowland, less than 150 m in altitude. Because of the wide area of carbonate rock and favorable hydrodynamic conditions of lateral circulation and solution, resulting from relatively stable tectonics with little uplift, and the combination of both allogenic water from insoluble rock areas and autogenic water from the abundant rainfall, this area comprises the major solutional plain in China. As a remnant of long-term lateral corrosion, many isolated limestone peaks rise abruptly from the plain. This landscape is typical of that found near Guilin, and is an important scenic attraction, known as tower karst, or Chinese-type karst (Yuan, 1986b). Underlying the plain is a well karstified aquifer which is often in danger of causing surface collapse when overpumped.

CLIMATIC CONTROL ON KARST

Karst water is stored and flows in various surface and subsurface karst features. Without a basic knowledge of karst features, it is difficult to understand a karst hydrologic system. Many factors affect the development of karst landforms and can be divided into two categories: geologic factors, as discussed above, and climatic factors. The latter covers not only precipitation and temperature but other factors, including the intensity of runoff and subsurface flow, vegetation, biochemical processes, and aggressivity of water as affected by its carbon dioxide content (LaFleur, 1984). Climatic conditions, especially precipitation, are important

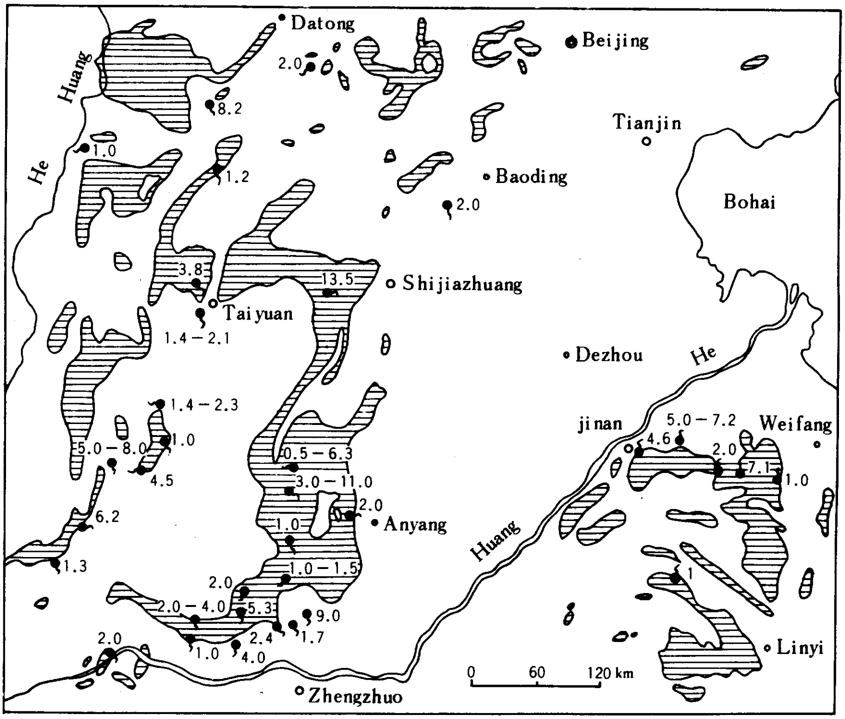


Figure 6. Map showing distribution of large karst springs in northern China (Geological Bureaus of Shanxi and Shandong, 1965). Number next to spring symbol is discharge in m^3/s. Hachured regions are underlain by Cambrian and Ordovician carbonate rocks.

in karst hydrological systems because of their strong influence on the development of karst features and as the main input to the system. Generally, China's precipitation decreases from southeast (more than 3,000 mm/a in eastern Taiwan) to northwest (less than 25 mm/a in southern Xinjiang Province; Fig. 5). Precipitation in the Guilin area (about 1,800 mm/a) and some parts of northeastern China (about 1,000 mm/a) is much higher than northern China (about 500 mm/a). Karst development in southern China is more intensive than in the north. Southern China karst is well known for its numerous underground stream systems; few large underground streams are reported in northern China karst, in spite of the presence of carbonate rocks with similar chemical and textural characteristics. The reasons for differences in the distribution of karst features include geologic and climatic factors.

To obtain a better understanding of the relationship between climate and karst processes, the limestone denudation rate has been observed since 1981. The observation points cover different climatic zones from the northeastern frontier to subtropical humid southern China, from the east coast to the desert margin in the west. The technique employs tablets made of Upper Devonian pure micrite-sparitic limestone from Guilin. The tablet is disk shaped, 42 mm in diameter, 2 to 3 mm in thickness, and weighs 9 to 16 grams. The limestone tablets are deployed in four locations at each observation point: two subaerial (0 cm and 150 cm above ground surface and two in the soil (15 cm and 50 cm below the surface). Each location has three tablets. All tablets are weighed annually, and the mean annual weight loss from a tablet for each layer is calculated.

The results from observation points in a north-south orientation show correspondence of mean annual temperature and precipitation. It shows a close relationship is apparent between limestone denudation rate and precipitation, although the temperature relationship is not clear. The denudation rate depends more on amount of precipitation than on latitude; for example, the denudation rate of Guilin is higher than at Guangzhou because of its higher precipitation, even though Guilin is farther to the north. The same relationship exists for the denudation rate of Yichun, Heilongjiang Province, northeastern China, and that at Beijing or Jinan, Shangdong Province. The contrast between subaerial denudation and subsoil denudation is varied; generally, in the humid regions, subsoil denudation is much higher than the subaerial rate, but in the arid or semi-arid regions, the situation is reversed. The subsoil tablets in Beijing even show small amounts of weight increase as a result of calcareous deposition on the tablets. The denudation rate of four observation points located along 35° north latitude including one from Japan (Akioshidai, collaborator: Dr. Yoshimura Kazuhisa), show a decrease from the east coast to the desert in the west as precipitation decreases.

The limestone denudation data sheds new light on the distribution of different karst features, which are essential to hydrological systems in karst. For instance, dolines and large cave systems are important to water circulation in a karst system, but they are mainly distributed in humid regions as a result of more intensive limestone denudation, and then are quite rare in arid and semi-arid regions except under some snowy mountains. Besides dolines and caves, karst features have a great deal of variation and can be distinguished into two major

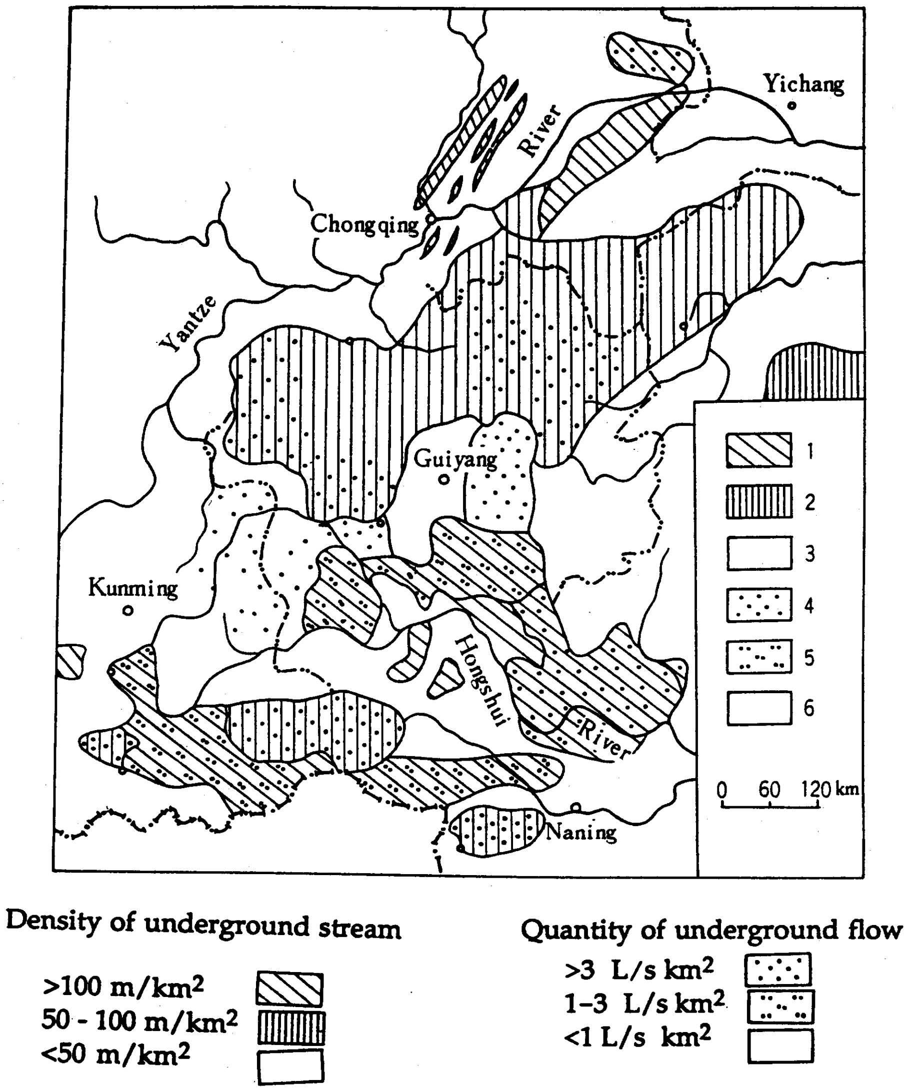


Figure 7. Map showing the density of underground streams and the modulus of underground flow in southern China karst.

categories, surface and subsurface features. These features in turn have different morphologies, despositional forms and solutional forms, each of which includes macro-forms, meso-forms, and micro-forms. Each climatic environment has a selection of karst features, known as the karst feature complex (Yuan, 1988), which are typical of a specific climate zone (Table 1).

Limestone denudation observations indicate that different climatic factors are not equal in importance and that precipitation plays a major role. Precipitation is a reflection of atmospheric circulation, which is influenced by a number of geographic conditions, including the distance from the coast and surface relief, rather than just latitude. This assertion is verified by the different trend of the isoline of the annual mean precipitation in different parts of China (Fig. 5). In central China, its trend is nearly latitudinal, but in northeastern China, it has a northeastern direction, whereas in southeastern China, it is orientated to the east-northeast. Precipitation isolines are sometimes parallel with mountain ranges, and sometimes with the coastline. This can explain the distribution of different complex karst features, and their relevant hydrological, biotic, and environmental characteristics in China (Table 2). For example, karstification in northern China is less intensive than in the south, as a result of lower annual precipitation (less than 500 mm), and at the same time the influence of hundreds of meters of loose cover on the carbonate rocks in northern China can hinder the infiltration of water and the development of endokarst. Farther north there are large karst features, where the annual precipitation is more than 800 mm. For example, the Xajiaweizi underground stream, 70 km to the southeast of Shenyang, northeastern China, is more than 2 km long, large enough for tourist cruises.

THE KARST HYDROLOGIC SYSTEMS

Thousands of karst hydrological systems have been investigated in China: large karst springs in semi-arid northern China and underground streams in humid southern and northeastern China. Each type of system has different characteristics of input, output, and interior structures.

In northern China, there are about 150 large karst springs each with a mean discharge greater than 0.1 m^3/s. Their total discharge is more than 200 m^3/s. About 60 have mean discharges greater than 1 m^3/s (Fig. 6). Most of them have been closely studied because of their importance to industry, agriculture, domestic water supply, and tourism. They originate mainly in the Cambrian and Ordovician carbonate aquifers. The boundaries of each karst system have been determined, and the discharge, water table, and hydrochemical characteristics of the systems have been monitored for many years.

In northern China, the Cambrian and Ordovician carbonate rocks dip gently. It follows that most of the karst hydrological systems in northern China cover areas as large as hundred to thousands of square kilometers. Towers and dolines are rare in northern China. The carbonate rock mountains are generally normal-shaped. Dry valleys are well developed because of the presence of underground drainage systems. In addition to direct infiltration on carbonate rock outcrops, the loss of surface runoff in the dry valleys usually comprises an important part of input to the

Table 1. Karst features developed under different climates.

Karst Features / Climatic environments		Arid, Semi-Arid	Humid		
			Arctic and Alpine	Temperate	Tropical Subtropical
Surface Features	Macro form	Normal shape Mts.	Normal shape Mts.	Hill-doline, stone knob-doline	Tower karst
	Meso form	Dry valley	Dry valley, shallow doline	Dish shape doline, solutional basin	Polygonal doline, solutional basin, stone forest
	Micro form	Individual karren	Sharp karren	Sharp karren, sub-soil round karren, kamenitza	Sharp karren, sub-soil round karren, kamenitza
	Depositional form	Limestone scree, case-hardening, caliche, calcrete	Limestone scree, stream tufa	Stream tufa, few outside stalactites, sometimes limestone scree	Terra rosa, abundant outside stalactites
Sub-surface Features	Meso form	Wind erosion rock shelter and through cave. Scarcely any cave except those under snow mts.	Large caves under snow mts. or modern glacier	Large cave system; its features depend on hydraulic conditions	Large cave system, with features depending on hydraulic conditions
	Micro form	Scallop, pothole in caves under snow mts.	Little scallops, potholes and notches	Potholes, scallops, notches, rock pendants	Minor solutional forms
	Depositional form	Small secondary carbonate deposit in caves under snow mts.	Glacial varve, collapse deposit, little secondary chemical and cold biotic deposit	Collapse and alluvial deposits, secondary chemical deposits, and temperate biotic deposits	Abundant secondary chemical deposits, collapse, alluvial, tropical and subtropical biotic deposits

Table 2. Types of karst systems and their characteristics.

Characteristics / Types	Mean annual precipitation (mm)	Mean annual temperature (°C)	Karst Features	Biotic Features	Main Environments	Distribution
Humid Tropical Jungle Karst	>1000	>22	Tower karst, polygonal doline, large cave system	Tropical jungle, percentage of forest cover 80%, nutrient soil, underground biotic community	Deforestation, soil erosion, drought, flood, surface collapse, soil cracks, water pollution, and damage to scenic attraction	Southern part of Guangxi, Yunnan
Humid Sub-tropical stony karst	>1000	16 - 20	Tower karst, hill doline, large cave system	Subtropical forest, percentage of forest cover about 10%, bare rock, underground biotic community	Deforestation, soil erosion, drought, flood, surface collapse, soil cracks, water pollution, and damage to scenic attraction	Karst of Guangxi, Guizhou, Yunnan, Hunan, Hubei, Sichuan
Arid, semi-arid karst	<500	11 - 20	Normal shape mts., hills, dry valley, caves are rare, except under snow mts.	Scattered grass vegetation, mostly arid plant, forest cover less than 10%	Plantation, water resource, water quality and engineering problems of evaporites	Shanxi, Hebei, Shandong, Henan
Humid temperate karst	>800	0 - 15	Normal shape mts. hill, shallow dolines, large cave system	Mainly grass pasture ecological environment, and also forests	Plantation, water pollution, surface collapse	Liaoning, Heilongjiang karst
Arctic and alpine karst	Variable	0 - 10 (alpine) <0 (arctic)	Tufa dam, large cave system under ice sheet	Vertical zonation of vegetation in alpine, permafrost vegetation in arctic	Permafrost, protection of scenic attraction, e, g, tufa dam	Tibet, Yunnan, Sichuan alpine karst

big karst springs. For example, from Baiyangshu to Luanliu, along a distance of 4 km on the Taohe dry valley, the loss of surface water is 0.44 m^3/s, which becomes a part of the discharge of Nyangziguan karst spring, the largest of this kind in northern China. The loss of Zhuozhang River at Xiwangqiao, north of Changzhi City, Shanxi Province is 6.15 m^3/s. The loss from Zihe River in Shandong Province and the Fenghe River in Shanxi Province are remarkable, and it is an important source to the karst springs. No underground stream has been identified from the semi-arid northern China karst although fossil cave passages hundreds of meters long at higher altitudes are found in some semi-humid locations. The interior structure of the Cambrian-Ordovician carbonate rock aquifer in northern China is generally regarded as a fissure karst aquifer with diffuse flow rather than conduit flow.

In southern China, underground streams are numerous. According to preliminary statistics (Yang Lizheng, 1985) there are 2,836 underground streams in an area of 300,000 km^2 in Guangxi, Guizhou, Yunnan, Sichuan, Hubei, and Hunan Provinces (Fig. 7). The minimum discharge of these underground streams totals 1,482 m^3/s, and they have a total length of 13,900 km. Most of them were identified by intensive cave exploration and tracing during a regional hydrogeological survey (mapped at a scale of 1:200,000) in the 1970's. Some typical streams have been studied in detail, with monitoring networks set up in recent years.

Tectonic effects on the karst of southern China are more complex. Guangxi karst is in an ϵ tectonic system; eastern Yunnan is in a longitudinal fold belt; eastern Sichuan, northeastern Guizhou, and a part of western Hubei and Hunan comprise a tectonic belt of strong folding in a northeast direction. These tectonic features, along with the interbedding of insoluble rock in some areas, make the size of the karst hydrologic systems in south China range from a thousand to just a few square kilometers.

The development of underground streams is also affected by topographic features. There are more underground streams on the northern and southern sides of the Yunnan-Guizhou Plateau, where the plateau is dissected by deep gorges of Yangtze River and Hongshui River (upper reach of Pearl River) drainage systems (Fig. 7). This setting gives rise to favorable hydrodynamic conditions for underground stream systems with a steep water table gradient to develop. The largest underground stream system known in southern China is in Tisu, Duan County, Guangxi. This system is on the northern side of Hongshui River Gorge and is developed on a northwest strike anticline composed of Devonian to Carboniferous limestone limited on both flanks by Middle Carboniferous dolomite. This system has a total underground course of 241.1 km comprising a 57.2 km long trunk and 12 branches. Its catchment area is 1,004 km^2, and its mean discharge is 38 m^3/s, ranging from 4 to 545 m^3/s.

Dolines are common in southern China and are the typical input location for karst hydrological systems. For example, thousands of dolines, ponors, and skylights are in the catchment area of the Tisu underground stream. The rainfall infiltration rate through an enclosed depression of this kind is generally high, as there is no surface drainage system. The data from some typical southern China