RIVER DYNAMICS OF INTERMITENT/EPHEMERAL RIVERS

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ABSTRACT.

Different cases of relative situation between aquifers and rivers are described, with emphasis in ephemeral and intermittent rivers. Principles of aquifer recharge, flow and discharge of aquifers related with ephemeral rivers are discussed, as well as existing methods to reduce uncertainties on aquifer recharges and transmission losses. Environmental implications of intensive use of aquifers, aquifer recharge in ephemeral rivers channels of imported or reused water and vulnerability to groundwater contamination in arid zones with ephemeral rivers are analyzed, as well as aspects of wetlands flow and contamination. Other environmental aspects including depletion of flow to wetlands due to groundwater pumping are commented, as well as the behaviour of ephemeral streams in karstic areas and several water resources management aspects.

INTRODUCTION.

Most research methodologies and experience in environmental flow assessment (EFA) are mainly related to permanent rivers with constant flow that in most cases have more or less important base flow components. In arid and semiarid areas, and in other situations common in Mediterranean climates there exist ephemeral and intermittent streams with non permanent flow (Chow 1964, Dingman 1994, Fetter 2001). Those streams with non dependable flow in most or even in all seasons can not support any important designated direct use, and even then only when they have water of adequate quality. Equally they can not support almost any kind of fish not seem to make much sense to apply to them similar EFA or water quality standards as applied to perennial streams. The IUCN decided to organize an expert workshop to discuss the main aspects of this topic. This work intends to be a discussion paper with emphasis in hydrological aspects, to induce an interchange of ideas. Therefore it should be considered as preliminary before being contrasted by the rest of the workshop participants. To begin with it seems worthwhile to analyze the different river aquifer situations.

DIFFERENT RELATIVE SITUATIONS OF RIVER AND AQUIFERS

Permanent rivers

Relative situation between aquifers and rivers is crucial for the existence of continuous or intermittent streams. In Fig 1 a classification of different river-aquifer situations slightly different from those previously given by Sahuquillo (1987) is proposed. Permanent rivers can be *gaining* or effluent when they drain their base flow from a connected aquifer, Fig 2. This is the most common situation in humid temperate climates. Although the river can recharge the aquifer during floods when river stage is above the water table, most of the recharge water can only remain in the aquifer during a relatively limited time before turning back to the river again. This is called *bank storage,* which we will not discuss in this paper. Groundwater head is always above the river stage. A *losing* or influent river (see Fig 3), can be connected to the aquifer and loses water to it. Although the existence of a water table below the river bed can be due to natural conditions, intensive ground water pumping can transform the river stage from gaining to losing trough intermediate progressive situations of increasing deeper groundwater heads. Most permanent rivers may have a less pervious bed due to the settling of silt or clay sediments in suspension. This silt layer is modified by floods and subsequent deposition of suspended material of stream water. That dynamics can be transformed by modification of floods and solid flow caused by upstream dam construction or water diversion.

There exist situations where a "*perched"* river is disconnected from the aquifer as in Fig 4. Existence of a semi pervious layer in the river bed reduces the flow towards the aquifer. To allow the existence of less flow than the one that would exist in the case of absence of the silt layer, the flow under the river should necessarily be a non saturated flow. The situation depicted in figure 4 can be natural or induced by man activities, if after reaching the situation of figure 3 pumping in the aquifer augments and the water table descends below the deeper point in the river channel. Once this situation is reached, vertical losses from the river to the aquifer would remain practically constant in spite of the water table descent. This effect has been denominated *"the shower effect"* by hydrologist, i.e. no matter how much you stoop while taking a shower you won't get more water. A perched stream can be permanent if the river remains fed from upstream with a sufficient continuous flow.

Intermittent streams.

Intermittent river flow can occur due to regional ground water head oscillation up and down due to differences in accumulated groundwater recharge during wet and dry periods. River dries when groundwater head is below the river bed (see Fig 5) and drains the aquifer when water table is above the channel bed during wet periods. Extreme situation are in *polges* in karstic regions, which are submerged valleys in wet periods that dry when water levels in karst go up.

There exists a different situation that causes seasonal intermittency in rivers draining low pervious basins. In this case rivers are permanent during wet season and in dry periods river flow ceases due to the capture by evapotranspiration in the riverside fringe of the meagre flow provided by the low pervious materials of the banks. The point is that the contribution from groundwater is meagre. They are *intermittent* streams but due to a different cause that the intermittence produced by water table oscillation commented above when describing the situation depicted in figure 5. This is the case of low fissured crystalline or metamorphic rocks; after a few weeks with scarce, if any, rainfall, their baseflow is minimal or there is only underflow trough the alluviums. In the two following figures, the first one, figure 6, is the hydrograph of a small river flowing to the Mirror Lake in a wet area in New Hampshire in North East USA in fissured granite. Having an important recharge of more than 300 mm/year and with a rather high base flow component in winter it forms a permanent stream with flow all the year round, except perhaps during the latest periods of the driest years (Mau and Winter 1997). But in most cases of basins in crystalline or metamorphic rocks in less humid environments the situation is more likely the case of figure 7. Figure 7 is a hydrograph of the Guadiamar River in the province of Seville in Spain. It drains slates that are more fissured at surface providing a relatively high continuous base flow in rainy winters. On the contrary, in late summer and fall mainly in dry years the river dries because the meagre flow drained from the exhausted "aquifer" is captured by evapotranspiration in the riparian fringe (Aliaga et al 2004). This is the situation of low pervious aquifer weakly connected to a gaining river in wet season and no flow in dry season. The no flow is in reality a non significant flow that is captured before reaching the river or disappears as underflow in the sands and gravels in the alluvium because groundwater table is always above the river stage.

In figure 1 *intermittent* stream appears twice for two conceptually different situations. One *intermittent* river can be successively gaining or losing, is the situation explained in figure 5. The other *intermittent weakly connected* river produce hydrographs similar to that reproduced in figure 7 where the river can gain water in humid situation and stop gaining or rather gaining irrelevant or small quantities of water in dry periods.

Ephemeral rivers

Ephemeral rivers flow only in response to specific heavy rainfall events. Hydraulic head in aquifers is always below river bed and usually well below it. They lose water recharging the aquifer. Existence of ephemeral rivers is the norm in arid and semiarid basins. Due to the much reduced aquifer recharge groundwater heads are low and river bed is above aquifer saturated zone all over their length except downstream, at the aquifer boundary or at specific sites where aquifer thickness is reduced, or it is absent. Downstream aquifer outflow feeds wetlands, salt marshes or flows to a lake or sea under ground surface. Figure 8 represents a downward flow from the river runoff produced by a heavy rainstorm.

Turning back to figure 5, in wetter climates as recharge augments ephemeral reaches in the river will be displaced upstream. For the same case if aquifer hydraulic conductivity is enough high water table gradient will be lower and the point where river situation changes from losing to gaining will displace downstream and groundwater in river heads can remain below ground surface. If water table slope is reduced, more and more downstream reaches would appear disconnected from the aquifer. Ephemeral river reaches can be found in steepest basins even with less pervious soil or rocks materials in less arid environments. Different losing or gaining river reaches can exist in adjacent contiguous reaches of the river due to changes from upstream to downstream in river channel slope, to changes in hydraulic conductivity of the aquifer and changes in aquifer recharge. That situation is very common in carbonated formations all over the Mediterranean basins. Even in humid climates headwater reaches are ephemeral in their steepest headwater stretches that shortly after rains remain dry except when water runoff caused by heavy rains circulates (Granato et al 2003, Barlow and Dickerman 2001).

In most cases a simple inspection of hydrographs or flow tables of stream gage stations provide a clear interpretation of river aquifer relationship. Figures 6, 7 and 9 provide a clear picture of it for gaining, ephemeral and intermittent streams. The case of intermittent alternative gaining or losing stream needs more insight and analysis than a simple look.

GROUNDWATER RECHARGE, DISCHARGE AND FLOW RELATED TO EPHEMERAL STREAMS.-

In arid climate direct groundwater recharge of aquifers by rainfall is usually very small. Recharge in arid zone is mainly produced trough stream loses in piedmonts, alluvial fans and ephemeral rivers, as rainfall recharge is exceedingly low or non-existent. Losses along ephemeral rivers are also call *transmission losses*. Losses in piedmonts, alluvial fans and headwaters between mountains and the main channel in those basins are equivalent to the otherwise named *mountain front recharge.* The two important processes included in the so called *transmission losses* are seepage, or aquifer recharge and evapotranspiration. The rate of seepage depends on type and hydraulic properties of the channel material, channel geometry, wetted perimeter, circulating flow and sediments, depth of groundwater, etc. Evaporation losses depend on meteorological conditions and transpiration of riparian vegetation that is supported by those losses. Accurate estimates of the transmission loss rates are required to determine aquifer recharge in addition to other water planning supply issues. It is a widely studied topic in arid regions of the United States of America, South Africa, India, Israel, Saudi Arabia, Australia and other countries. Studies concerning transmission losses have been made not only in short lived ephemeral streams, but also in losing perennial rivers with or without upstream dams, because rivers are considered necessary mechanisms to convey water resources to points of demand. Nevertheless, losing river mechanism in an ephemeral stream is a transient process produced by an infiltration pulse whose value is infinite for time cero and decreases as saturated front descents. The process is very different to a steady or quasi steady infiltration in a losing permanent river. The difference is conceptually showed in figures 4 and 8.

Losses in a permanent river can be localised along the river bed more easily and precisely measuring the flow in different points and localizing flow drainage through temperature, gravity, chemical and isotopic determinations. Infiltration of a pulse, as approximated by the Green Ampt equation, is a quickly changing process very difficult to measure. In any case, the solution depends on saturated conductivity values and their spatial distribution that in real cases are very heterogeneous and widely unknown, as they are specific hydrogeological aspects of detailed distribution of materials of different hydraulic conductivity in the river bed and below it. The assessment of recharge needs a detailed knowledge of hydrodynamic properties of the aquifer that it is almost never available. This has prompted an ample number of studies and research to first understand and latter quantify, minimizing time and cost of assessing transmission losses. Some of the methods used are regression, simplified differential equations, field experimentation, hydraulic streamflow routing, water balances or combination of several of the above mentioned methods, (Vivarelli and Perera 2003).

Abdulrazzak and Morel Seytoux (1983) analyse the infiltration process of flow in ephemeral streams. A saturation front descends according to the Green Ampt formula which they consider to reproduce well the real phenomenon. Infiltration flow is infinity for time 0 counted from the beginning of the flow and decreases as the saturation front deepens. When the flow ceases infiltration also ceases, and saturated mass continues its downward flow after eventually having reached the aquifer. It is at this moment when the aquifer recharge starts. A mound of higher groundwater head is produced below the river channel. This mound dissipates laterally according the groundwater flow equation laws. Once having reached the saturated zone, flow becomes horizontal, producing an expansion of the recharge front. The above mentioned authors suggest a method to calculate infiltration and the water table elevation from this moment.

Some solutions for the water table rise transmit immediately to the aquifer the infiltration in the stream bed without any retard. That can be inadequate for deep water tables. Initial aquifer recharge is null, and if we consider that behind the saturation front the water content is near saturation, when the front reaches the water table a rapid augment of the recharge occurs, which is equal to the water infiltrated in the stream bed until this moment. Abdulrazaaak and Morel Seytoux analysis continue after the saturation front reaches the water table, but they do not analyse what happens when bed infiltration ceases. Freiberg (1983) uses as well the Green Ampt formulation in a slightly different way. He considers suction head in the non saturated zone and the net change of water content above the saturation front, and takes in consideration the geometry of the channel bed. He argues that the application of Green Ampt matches well with numerical solutions of the Richards equation, whose solution is much more complex due to its nonlinearity and the need of a very detailed space and time discretization. He affirms as well that Green Ampt approach reproduces well laboratory experimental results. Freyberg (1983) does not consider either the process after the infiltration stops.

In any case Green Ampt formulation does not apply to heterogeneous or stratified media. Less pervious lenses or strata induce a horizontal component of the flow vector that enlarges the infiltration front. Existence of low pervious layers or high groundwater levels, in addition to allowing less recharge, can cause later loses by evaporation, outflow to drains and channels, or clogged areas. The consideration of the flow conceptual models as in figures 3, 4 and 8 simplifies reality because it ignores the differences in saturation, the existence of a capillary fringe and the ample variability of hydraulic conductivity with saturation and heterogeneity. However, they can help to understand most of the hydrologic phenomena related with aquifer river interaction. In any case, it seems necessary to distinguish between bed infiltration, deep percolation and aquifer recharge. Lerner et al (1990) referring to Wilson and Cook point out that the Santa Cruz River in Arizona USA near Tucson, only 33% or recharge appeared immediately as an increase of piezometric head, and the rest appeared several moths latter. Increase of water content of unsaturated zone can store some of the infiltrated water, not all of it ending as aquifer recharge. Initial water content, especially between relatively close recharge episodes, influences both the water recharged to the aquifer and the time of arrival of the wetting front to the aquifer. Aquifer recharge and water level increase depends on the initial water content in the unsaturated zone, from the tension saturation curve and from the hysteresis (Parissopoulos and Wheather 1982). For the same reason when water table rises by the effect of losses in the river bed, incorrect results can be produced if aquifer recharge is calculated considering an effective porosity multiplied by the increase in saturated volume (Sophocleus 1985). Extrapolation of head variations to zones of the aquifer without piezometric measurements can be difficult and without doubt is a possible source of error. Besbes et al (1978) also consider the difficulties presented by the determination of the effective porosity. Considering long term effects in total aquifer recharge, storage in the unsaturated zone can be estimated by the difference between river bed infiltration minus evaporation and evotranspiration. The problem is the errors in both determinations.

Besbes el al (1978) estimate the recharge trough a calibrated groundwater model. They choose a big flood after a prolonged dry period and adjust by trial and error the recharge needed to reproduce the observed piezometric oscillations observed in wells during and after the flood. Additionally they use the convolution to estimate the recharge. The net rise of groundwater levels in an observation point i, $s_i(t)$ is given by

$$s_i(t) = \int_o^t r(\tau)\, \Phi_i (t-\tau)\, d\tau$$

Φ_i being the linear response in piezometer i, r(t) is the recharge in the river bed, and $s_i(t) = h_i(t) - d_i(t)$ is the net rise between the measured water head and the one that would have had if flood would not have been produced. They suppose that measured heads are not influenced by other causes as pumping, or in such a case they correct them. Values $d_i(t)$ are estimated with the model or extrapolating precedent measures. The process requires two simplifications. The first is that the transmission of the recharge trough the unsaturated zone is a lineal process. To be lineal, in addition to multiply the unit response by the recharged volume, the arrival time to reach the aquifer must be the same. But this is not true as the arrival time is less when recharged volume augments. Probably the influence is not big and in any case the delay in time can be taken in consideration. The second is that the distribution of the recharge in the river bed is the same for all floods, which must be far from reality. To determine the impulse response function Φ_i they combine two methods. (1) The first uses the numerical model of the aquifer. A big flood can be calibrated reproducing the heads measured temporally in several piezometers. If we simulate a unit recharge with the same distribution in the channel bed and the same time lag that the recharge used in the simulation an impulse response function Φ_i (t) will be obtained for each well. (2) The second method is the observation of a flood. If an important flood with great head rise in piezometers and the recharge produced by the flood can be estimated, the impulse response function can be obtained directly by

$$\Phi_i = [h_i(t)–d_i(t)]/ \text{ volume recharged}$$

The method works until the next important flood modifies the system behaviour due to changes in the river bed. Recharge depends on the value of effective porosity. If it is overestimated recharge is also overestimated and if it is underestimated recharge is similarly underestimated. There were 23 years of piezometric data and an empirical relationship between runoff and recharge was established. This was made graphically after determining the recharge from eleven identified floods using deconvolution. The recharge-runoff relation was used to estimate the recharge induced by the different operation rules of a dam proposed to be built upstream.

In Moench and Kiesel (1970) and Hall and Moench (1972) the convolution of the recharge is used with the analytical solution of heads in an one-dimensional aquifer homogeneous and infinite for an instantaneous unit recharge in a band of large *w*. If h(x, t) is the solution for h in a point distant x from the centre of the band considering the system as lineal, superposition can be used. Resulting piezometric head G(x, t) is:

$$G(x, t) = \int_o^t F(\tau)\, h(x, t-\tau)\, d\tau$$

The problem is to obtain F (t) from the function impulse response derived from certain diffusivity. Final results must be analyzed for every case considering observations and existing data.

Flugg et al (1980) use a model to estimate the recharge, based in the relation obtained between recharge and runoff determined by an ephemeral stream in Arizona. Recharge is determined by a lineal model that uses as independent variable the runoff duration and lumps all unknown factors in two parameters **a** and **b**. Parameter **b** is a threshold for the runoff duration below which there is not infiltration, the case being improbable that flows of very short duration might produce any recharge. Parameter **a** is a proportionality factor that transforms flow duration in recharge for a particular stream. Parameters **a** and **b** depend on stream physical and hydrogeological characteristics. They have been supposed to be independent from season and are obtained by calibration.

Volume v(j) produced by flood **j** is given by

$$v(j) = a\,[y(j) - b] \qquad y(j) \geq b$$
$$v(j) = 0 \qquad y(j) < b$$

Being:

j = 1,2,..J the floods and y (j) its duration. For each period recharge was estimated by the balance equation.

$$(V_F - V_I)\,S = R + Q + \Delta I$$

Being:

V_F	Saturated volume of the aquifer at the end of the period
V_I	Saturated volume of the aquifer at the beginning of the period
S	Effective porosity
R	Recharge
Q	Volume pumped in the period
ΔI	Net subterranean inflow estimated by Darcy´s law

Walters (1990) and Sorman and Abdulrazzak (1993) also developed regression relationships for ephemeral streams as some others refereed by Vivarelli and Perera (2003) that consider that the use of regression tools is questionable as they were derived at specific locations, from events subject to local influence parameters and they do not fully explain transmission losses. In any case they can be useful for the site they were developed.

Simplified differential equations have been developed by Peebles et al (1981) that conceptualize the ephemeral channel as a reservoir capable of represent streamflow recession. The model considers the continuity equation and relations between storage and discharge. They assumed a constant loss rate per unit area in the stream, infiltration loss as function of discharge and time and high loss at the onset followed by decrease to nearly constant. The model was calibrated using two parameters: initial storage and reservoir leakage rate. Jordan (1977) also uses differential equations assuming that loss rate between two gauging stations was proportional to the flow and depends on channel characteristics. He introduces the loss in the first mile of the stream to standardise data as flow decrease is nonlinear and compare results in different rivers. In the above cited work of Vivarelli and Perera (2003) more publications using both approaches of differential equations and regression are analysed including the works of Lane (1982) and Rao and Maurer (1996) and comment other methodologies such as stream routing and the use of hydrologic balance, (Smith 1972, Abdulrazzak and Sorman 1993, Gu and Deutschman 2001).

Mountain front recharge is the recharge in the mountain areas and headwaters of the basin. It constitutes only a minor fraction of the total amount of water in the system and cannot be estimated by more or less precise water balance equations. Data on the mountain and the mountain front region are ordinarily very limited to some spaced wells, springs, and flow in some stream reach. Nevertheless estimation of *mountain front recharge* is required for management purposes in groundwater basins where demand surpasses water availability. Different methods like chloride balance or other tracer as environmental isotopes (Simpson et al, 1970), and water yield regression (Scott and Anderholm 2000) provide wide discrepancies. To tackle the problem Chavez et al (1994) applied an analytical model of the seasonal streamflow where initial abstraction and the long term effective subsurface outflow, or mountain front recharge, are viewed as unknown model parameters. In addition they introduce a procedure that combines the water balance equation with a relation provided by the so called *"vegetal equilibrium hypothesis"* from Eagleson which enables the estimation of effective soil related parameters jointly with the mean seasonal evapotranspitation and surface runoff. The authors apply the method to a mountain basin in southern Arizona and claim it is useful when large uncertainties are associated with prior estimation of mountain front recharge.

Possibilities of evaluating transmission losses with scarce data.-

In few cases are there neither prolonged data of flow in ephemeral streams nor enough water heads data distributed on the aquifer during large time periods to allow to determine changes in storage. Usually there are few permanent gauging stations due to the flashy, erosive streamflows common in ephemeral channels which makes it impractical to deploy and maintain series of them in mobile bed alluvial channels. Erosion and sedimentation in such arid fluvial environments is the main cause of the lack of precision of most of the existent stations. When existing data are used to determine transmission losses by differences of flow between stations, possible errors and uncertainties must be considered. It is always possible to make hydrological analysis and develop models of rainfall-runoff-infiltration to determine both runoff and aquifer recharge. Thresholds determined for the start of runoff and transmission losses must be considered in the models and results must be compared to be coherent with existing data of flows and water levels in the aquifer. If a calibrated model of the aquifer is achieved the consequences of the determined aquifer recharge on aquifer head rise due to transmission loses must be checked out simulating the model, but unfortunately this is seldom the case.

Lerner et al (1990) transcribe, from a research made by the Water Resources Research Centre of the University of Arizona in 1980, the most important factors that influence in the recharge of ephemeral streams in arid zones:

- Characteristics of flood and of the channel
- Infiltration augments with the velocity of the flow and with temperature.
- Infiltration decreases when suspended solids augment.
- Infiltration is lower downstream, as channel sediments permeability lowers and suspension solids augment downstream.
- Infiltration decreases when water table is near to surface.
- Lithology of the source area of sediments to the stream influences their permeability.
- Volume infiltrated in the stream increases linearly with flow, until some point above which losses do not increase.
- Winter floods seem to produce more recharge that summer ones.
- In heterogeneous stratified materials only a part of the losses reach the saturated zone immediately. The rest drains slowly from the vadose zone during months to the aquifer.
- Recharge is mostly produced in main channels having higher volumes of pervious alluvium and more flow.
- The importance of silt and clay layer is crucial and can drastically reduce the recharge.

To make more direct and reliable experimental determinations of the proportion of runoff transmission losses that escape from near channel evapotranspiration and wetted channel evaporation to become deep groundwater recharge, a series of alternative measurements of temperature, changes in microgravity, as well as isotopic and chemical determination have been used recently, Goodrich et al (2003). Temperature plays a key role in the health of streams, including the benthic habitat of streambed sediments. Stream temperatures are influenced by changes between streams and nearby groundwater. Heat provides a natural tracer of groundwater movement readily obtained by measuring temperature. Water that moves between a stream and adjacent sediments carries heat with it. Equally fluctuations of land surface temperature provide boundary conditions for tracking heat exchanges between surface water and groundwater. Daily and seasonal temperature fluctuations are modulated by heat carried upward and downward flowing water. This advective heat transport imparts a distinct pattern to gaining versus losing streams. Figures 10 to 13, taken from USGS (2004) figures 3, 4 and 5, show graphically this concept.

Streambed temperature data were collected at the surface or at shallow depth along the stream channel, using thermistors encased in waterproof data loggers. They allow to detect the presence and duration of sporadic floods and to identify gaining and losing reaches. The thermal properties of sediments vary within narrow ranges and can be obtained from literature values. In contrast, the hydraulic properties of sediments vary orders of magnitude and are typically highly uncertain. Knowledge gained from thermal studies is useful for evaluating the sustainability of biological and water resources. A simulation model was developed to compare results with a series of hypothetical situations. The models clearly showed the importance of lateral as well as vertical components of subchannel flow (Constantz and others 2002, Goodrich and others 2003, USGS 2004).

The temporal- gravity method was used to estimate water storage changes and specific yield values in the Rillito Creek, Tucson, Arizona USA. The method applies the Newton's law of gravity to measure local changes on gravity caused by changes in the mass and volume of groundwater. Water levels in wells rise as 90 feet , and gravity increased as much as 90 microgals. Water level declined and gravity decreased near the stream after the mayor winter flow but continued to rise respectively downstream. Good correlation between water levels and gravity resulted at five nearest wells to the stream, but poor correlation was obtained from groundwater storage change in perched aquifers and in the unsaturated zone near the stream. Microgravity determinations would probably be in the near future an interesting research tool for a better understanding of flow interchanges in ephemeral streams, (Goodrich et al 2003).

EPHEMERAL STREAMS IN KARSTIC AREAS.

Hydraulic conductivity of carbonate formations varies orders of magnitude not only between different areas but also within few meters of the same formation. Abrupt changes in permeability are due to the networks of solutionally enlarged fissures and the presence of caves and solution cavities. Nevertheless in most cases they can be analysed with classical hydrogeological methods, which is the current practice. Their main hydrological particularity is that they present wide areas of losing rivers even in wet areas due to the high hydraulic conductivity of some calcareous or dolomite formations. In the Spanish Mediterranean areas there are ample areas without permanent rivers (Camarasa and Segura 2001). Many of them are coastal basins with ephemeral rivers denominated rambla, riera, arroyo or simply río seco. In many detailed cartographic maps of Spain it is possible to find several Río Seco, (Dry River in Spanish) as the name of an ephemeral river. Those ephemeral rivers feed freshwater wetlands all along the coast North and South of Valencia, although in few cases detailed hydrogeological studies have been made. They are dry all the time except when sporadically the carry very high destructive floods. In other cases karstic aquifers feed coastal or submarine springs or discharge to the sea as coastal aquifers. All along the Mediterranean karstic aquifers feed coastal or submarine springs or discharge to the sea as coastal aquifers. Attempt of capturing submarine or coastal flow from karstic aquifer have failed due to "up-coning", the small thickness of the fresh water lens, the amplitude of the mixing zone, or the easy connexion with salt water. Fissuration and solution channels make karstic aquifers highly vulnerable to contamination, so provisions must be taken to prevent dumping of solid and liquid wastes, particularly in dry channels.

The Rambla de la Viuda 80 km North of Valencia with a basin of 1.300 km^2 produced in 1962 a flood with an estimated peak of near 2.000 m^3/sec. Some dams have been built to capture flood water. Most of them, as should have been expected, have had big water losses more or less attenuated by grouting works. Maria Cristina Dam in the lower reaches of the Rambla de la Viuda was built in the earliest 1920s. Figures 12 and 13 show a general view of the dam a cross section and the evolution of storage and piezometers. In the cross section piezometric levels are up to 50m below the river bed when the reservoir is empty. Higher levels show the situation for full reservoir. In this situation reservoir loses are about 1 m^3/sec after big grouting works. Figure 13 shows a typical annual evolution of the reservoir and piezometres. Reservoir fills up in fall and reservoir empties progressively due to water irrigation intakes and leakage. At the early autumn the reservoir is filled very quickly by a flood and in few days or hours water level goes up more than 50 m. The process can be compared to the very rapid descent of saturated front in figure 8 that in this case is very quick probably due more to the low storativity of limestone, on the order of 0.01 to 0.02, than to the rock permeability. It seems possible that reservoir losses in the first hours after reservoir filling reached more than 20 m^3/sec. After very few days or hours groundwater levels near the dam are high and the reservoir leakage stabilises around 1 m^3/sec. After the latte 1960s no more attempts were made to stop leakage in this and other reservoir whose free aquifer recharge is jointly used with surface water to made a very efficient conjunctive use all over the Valencia Community. Even more several dams have been projected not to store water but to recharge flood water trough the losses in the storage. Such is the case of the Algar dam built in an ephemeral reach of the Palancia River upstream Sagunto, North of Valencia.

After 1960 the reservoir has spilled three times: in 1962, 1969 and 2000 and probably it did not spill many more times. Consequently no flow has been left downstream the dam. The solid volume trapped by the reservoir is in the order of one or two million m^3. That represents a very small volume as compared with the usual solid flow in detritic basins. When the dam once dried completely the author overheard someone say that big quantities of eels could be seen in the mud. No other environmental consequences or effects have been described.

A heavily exploited karstic aquifer in La Mancha area in Central Spain feeds one of the emblematic wetland in Spain, Las Tablas de Daimiel with a total wet surface of 20.000 hectares that was reduced to around 200 due to the heavy exploitation of the aquifer. Groundwater pumping was scaled down as consequence of an EU program to compensate farmers for giving up irrigating lands. That and some humid years produced a slight increase in wetland acreage.

ECOLOGICAL FUNCTIONS OF EPHEMERAL STREAMS

At first glance there exist two main environmental concerns related with ephemeral rivers. One is the influence of aquifers exploitation on wetlands. The other is related with groundwater contamination caused by pouring contaminants to ephemeral channels.

Groundwater pumping in ephemeral or perched rivers can lower groundwater levels. But it we look at figure 4 it does not produce an increase of river infiltration, nor augments flow velocity in the unsaturated areas above the water table that can lower the arrival time of eventual contamination plumes. Making the same analysis with the figure 8 we can conclude with similar results. Figure 5 suggest different conclusions. Groundwater pumping in the aquifer will induce drawdown in the aquifer water table, displacing downward the point where saturation line cuts the river channel, augmenting the total length of losing reaches.

It makes no sense to discuss the effect of pumping in an intermittent river weakly connected to low pervious rocks, because no wells are drilled in such sites. But of course it seems appropriate to consider the important hydrologic and environmental changes produced by intensive groundwater abstraction that transform a gaining river into a losing or even ephemeral river. To analyse those question as well as the influence of ground water pumping on possible wetlands downstream, routine but tough hydrogeological studies should be performed.

In dense populated areas, ephemeral stream channels have been used traditionally as uncontrolled waste sites for wastes and wastewaters. Many of them have resulted in the creation of effluent dependent aquatic systems. Over time, these aquatic systems developed riparian communities that provide diverse habitat and new created ecosystems. Because no opportunity exists of in-stream dilution exist discharges to dry ephemeral or effluent dependent streams have to meet stringent water quality criteria. In many cases reused water is the most economic alternative to solve the increasing water demands. Then the loss of discharge results in the loss of a potentially critical aquatic resource. Decisions about the chemical water quality criteria to be apply to waste water are an important issue, but probably more important than the chemical criteria, is the need to determine what should be protected and what level of protection is necessary to achieve this goal, (Meyerhoff 2003). In the other hand, wastes and particularly toxic wastes can contaminate aquifers feed by losing streams. Before taking decisions it is necessary to understand the paths as well as hydrologic, chemical and biological changes of groundwater from headwaters and ephemeral channels to downstream. Problems on some karstic aquifers related with ephemeral rivers have specific particularities that must be discussed.

Riparian areas have been described as the transitional zones between aquatic and terrestrial environments. Riparian areas occurring along the banks of moving water (i.e., streams or rivers) are often called *lotic* systems, whereas those occurring along the banks of stationary water (i.e., lakes or ponds) are called *lentic* systems. As a transitional zone between aquatic and upland environments, riparian systems have characteristics of both; but they are not as dry as upland environment and they are not as wet as wetland ecosystems. What kind of organisms lives in the zone near and below the ephemeral channels? This zone is unsaturated during dry periods and saturated during sporadic short periods. In addition to this, does the hyporheic zone (defined as saturated sediment beneath or besides river and streams) in ephemeral streams function as a refuge for organism during dry periods? Does it exist as such kind of ecosystems? What is the importance of hyporheic zone?

A hydrologic function of runoff is the transport of sediments, salinity, nutrients and dissolved constituents. Floods transport with water vegetal debris, sheds and chemicals used in agriculture as fertilizers and biocides. They sediment in ephemeral streams as they lose water and at the end of the pipe at wetlands, estuaries or artificial dams built by man. What is the ecological function of fine sediments, organic compounds of disintegrated vegetal debris and others? What is known and what is essentially unknown? What is known from the chemical characteristics of water in ephemeral stream floods? Or which is the influence of dam construction or water derivation on the ecosystems in ephemeral streams?

WATER RESOURCES MANAGEMENT ASPECTS.

Groundwater in aquifers associated with ephemeral river hold a valuable resource that has to be handled adequately. It can be used conjunctively with surface water to obtain the maximum benefit and avoid groundwater contamination and nuisance to the environment. Ephemeral channels can be used to be a tool of artificial recharge in the channel with or without adequate preparation for increase its infiltration capacity avoiding the entry of contaminants. Flood water is an important resource of water that has to be settled and adequately treated to avoid clogging. Above the use of leaking dams to increase aquifer recharge has been mentioned, as well as the use of treated waste water poured into ephemeral channels to use aquifer as storage and distribution system and the unsaturated zone in detritic aquifers as a treatment and filtration device to improve water quality.

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LIST OF FIGURES

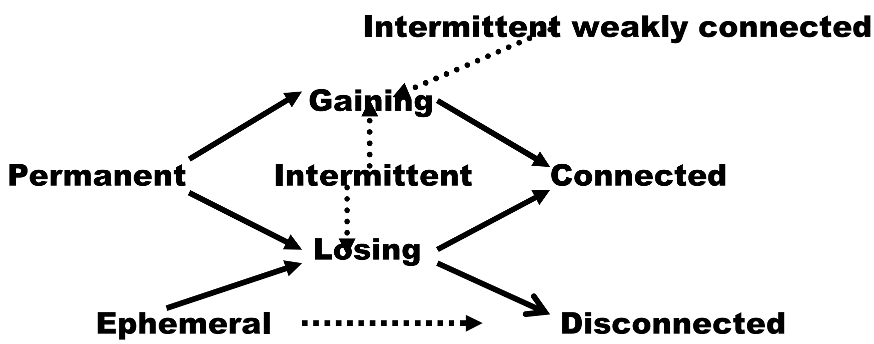

Figure 1. Different aquifer river situations

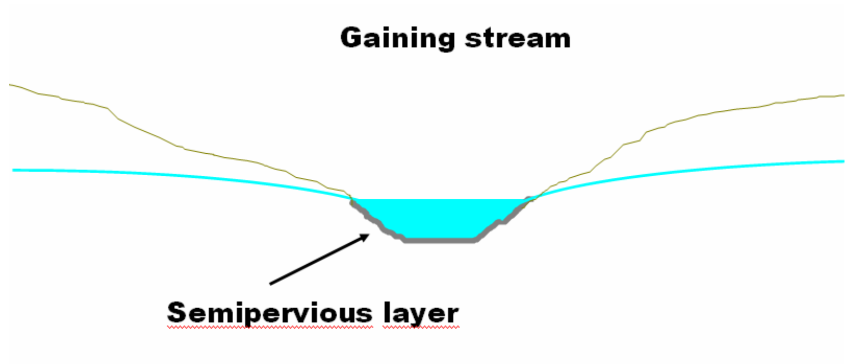

Figure 2. Gaining river

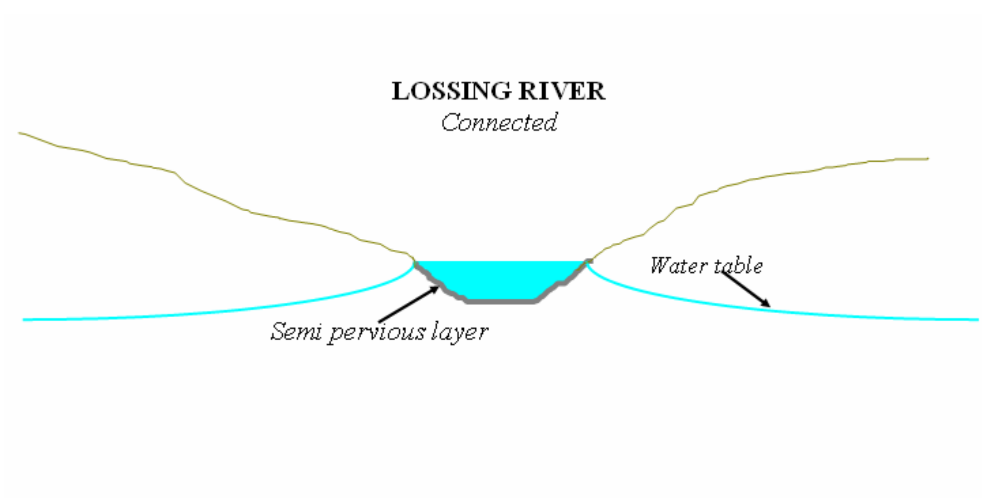

Figure 3. Losing stream

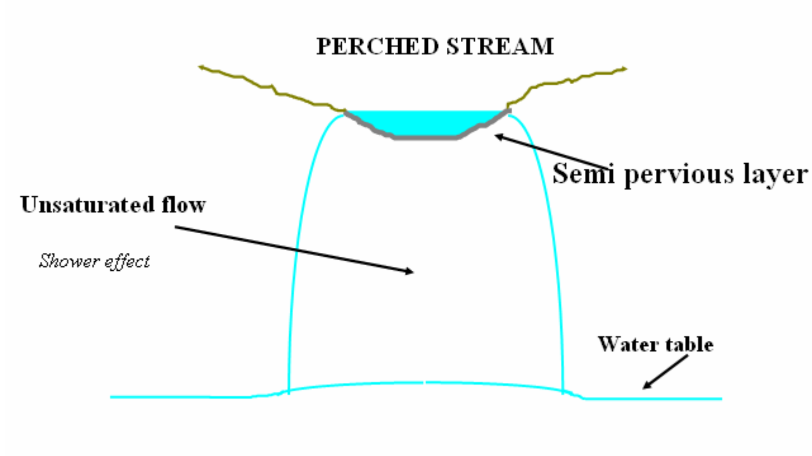

Figure 4. Perched stream

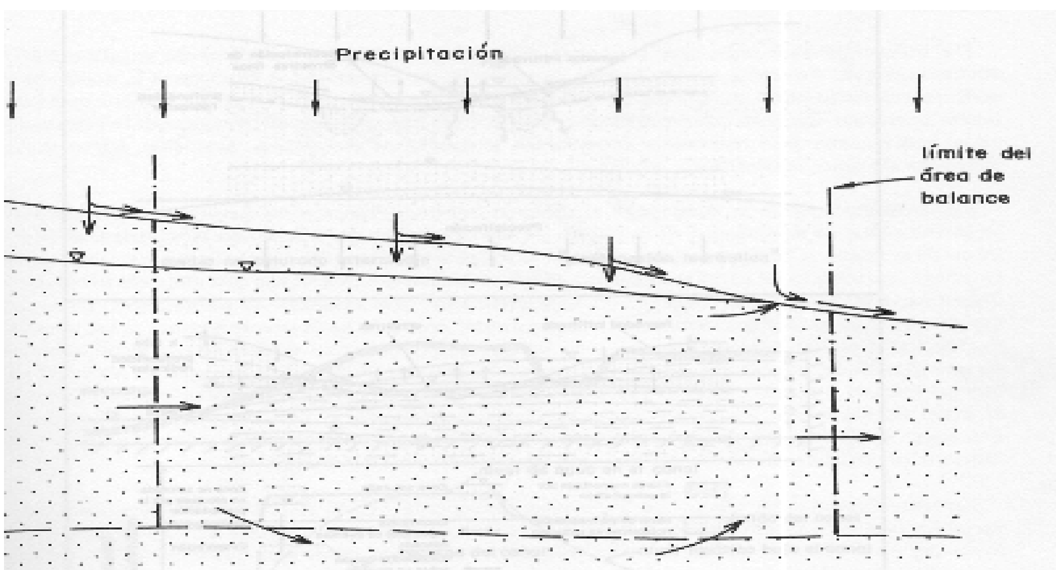

Figure 5. Longitudinal profile trough a river channel.

As groundwater heads augment in wet periods the saturation line cuts the channel more upstream and gaining reaches of the river augment. In dry periods losing reaches augment downstream

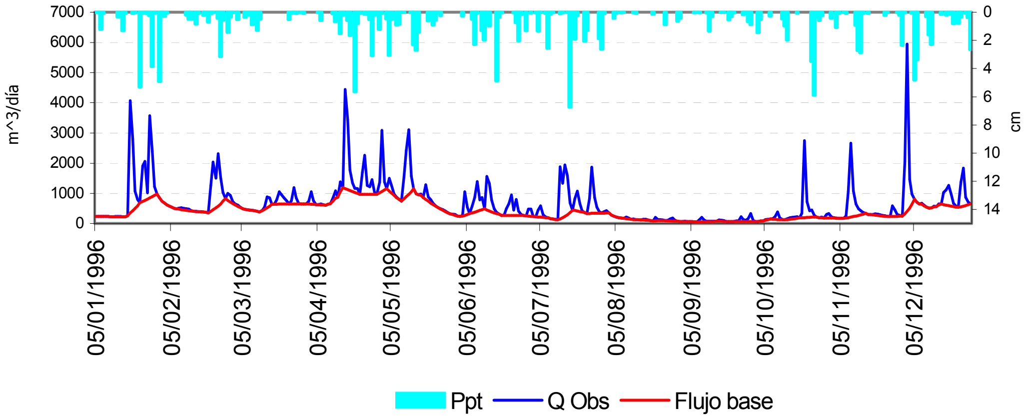

Figure 6. Base flow. Granite in a humid basin

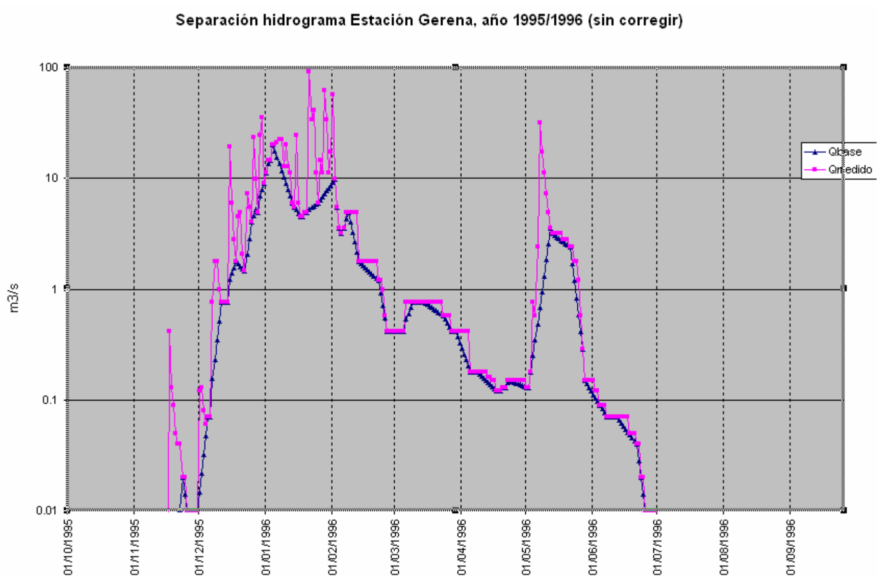

Figure 7. - Base flow in an intermittent river

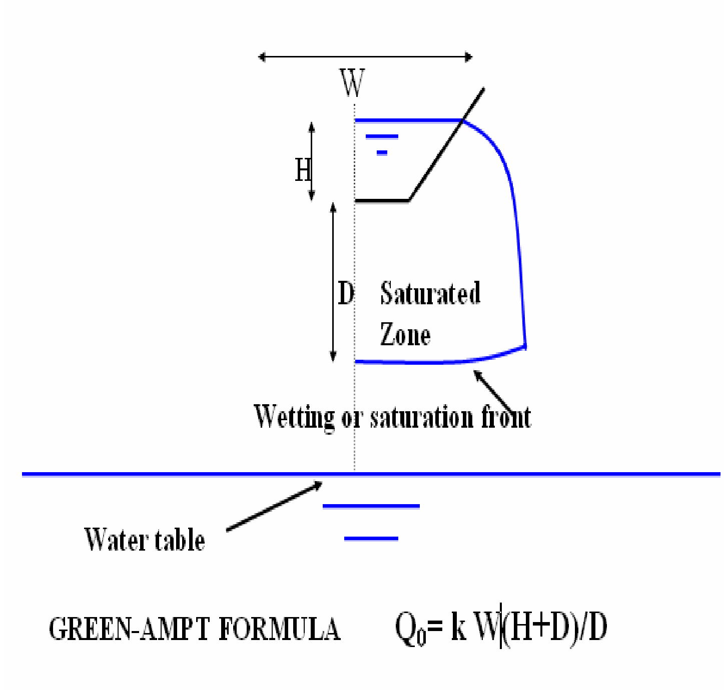

Figure 8. Infiltration pulse in an ephemeral river

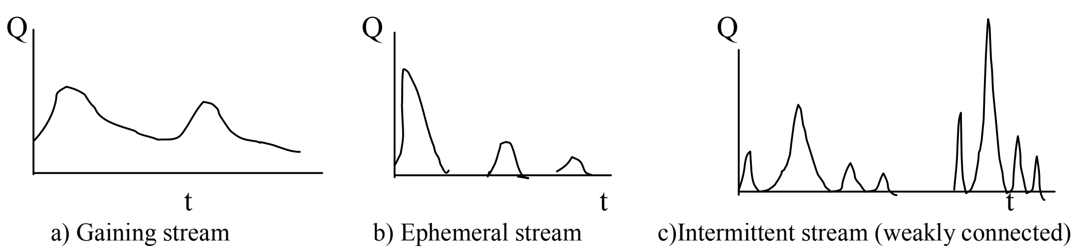

Figure 9.- Streamflow hydrographs for gaining, ephemeral and intermittent weakly connected streams

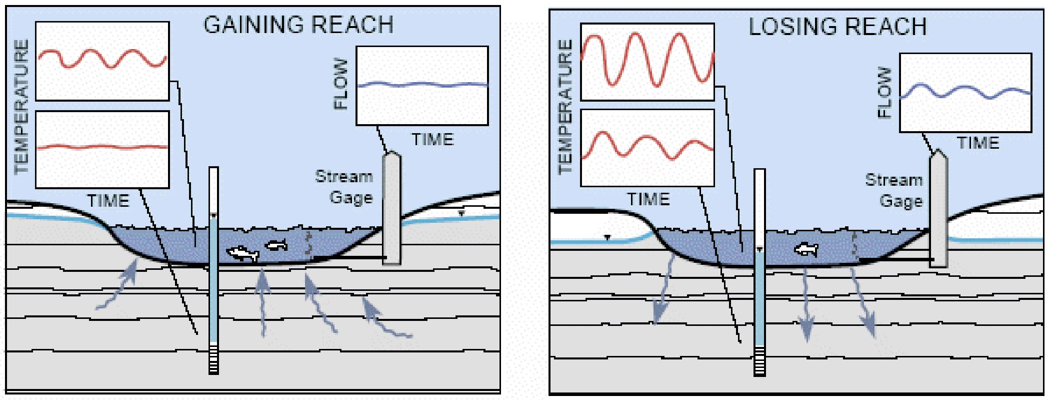

Figure 3. Stream flow and temperature histories for gaining and losing reaches of a stream coupled to the local ground-water system. Ground water is buffered from temperature fluctuations at the land surface. Temperature fluctuations in and beneath the gaining reach are therefore muted (left panel) compared to temperatures in and beneath the losing reach (right panel).

Figure 10. Stream flow and temperature in gaining and losing reaches (From USGS 2004)

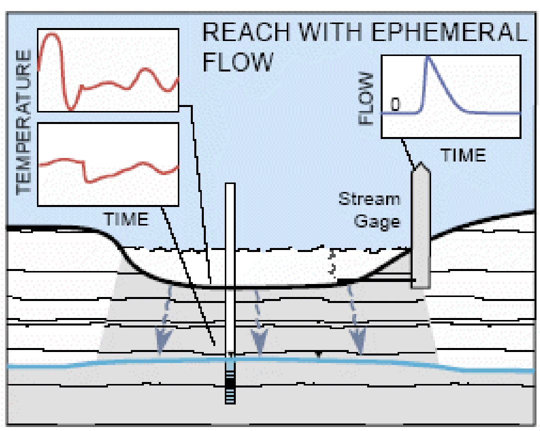

Figure 4. Ephemeral flow increases heat transport, as evidenced by thermographs.

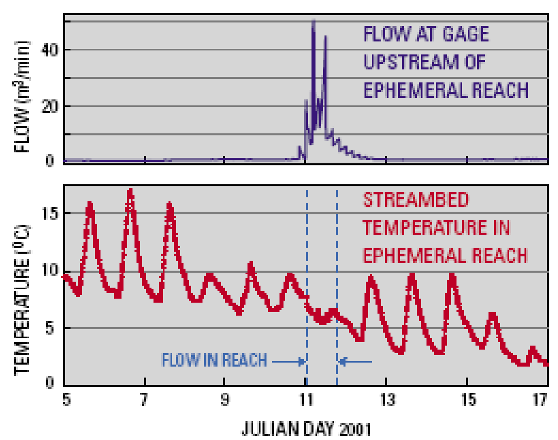

Figure 5. Streambed temperatures before, during, and after ephemeral flow event (example from the Amargosa River, Nev.; sensor is ~0.1 m deep).

Fig 11 Temperature transport in ephemeral flow, (From USGS 2004)

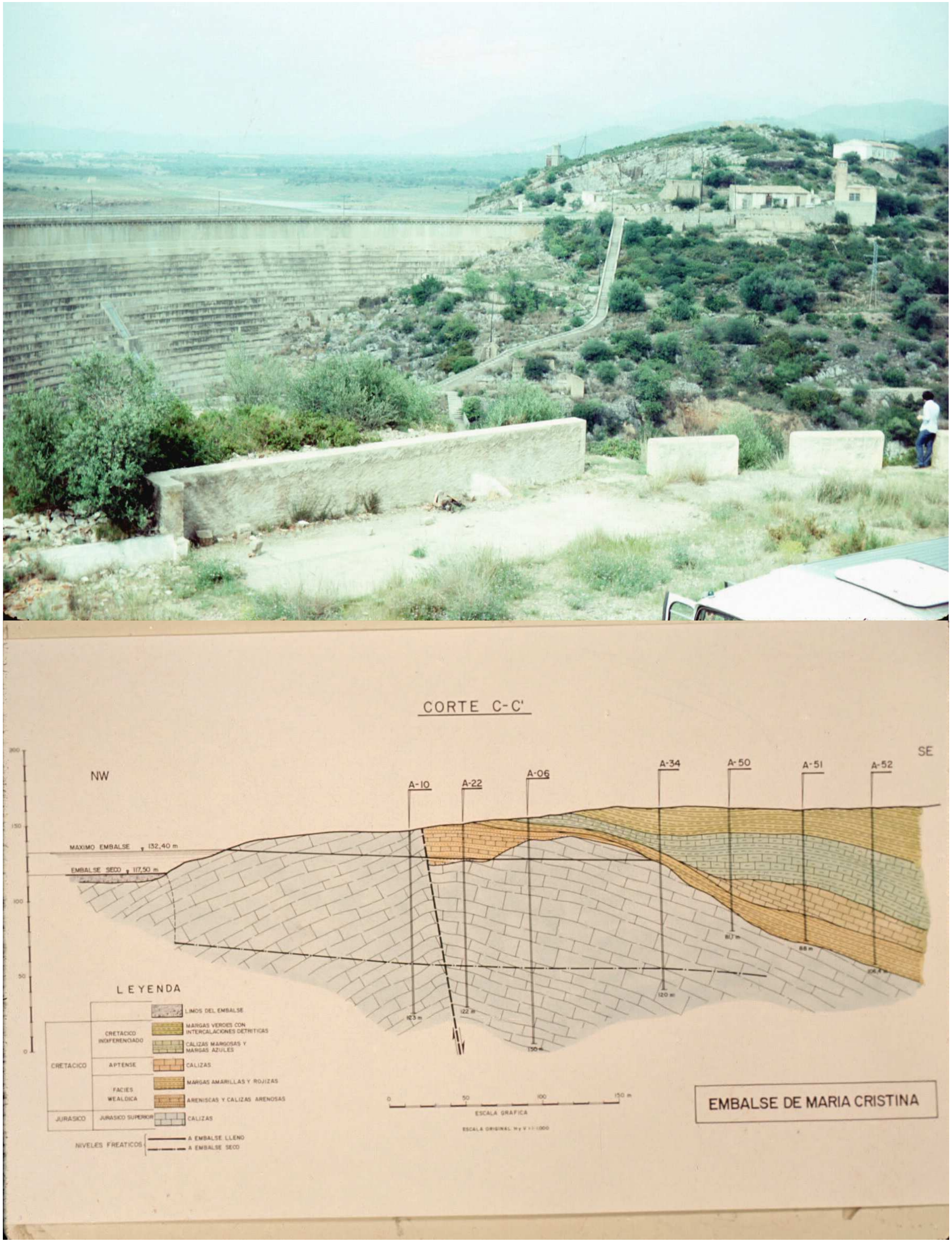

Figure 12 Maria Cristina Dam and Hydrogeologic Cross Section

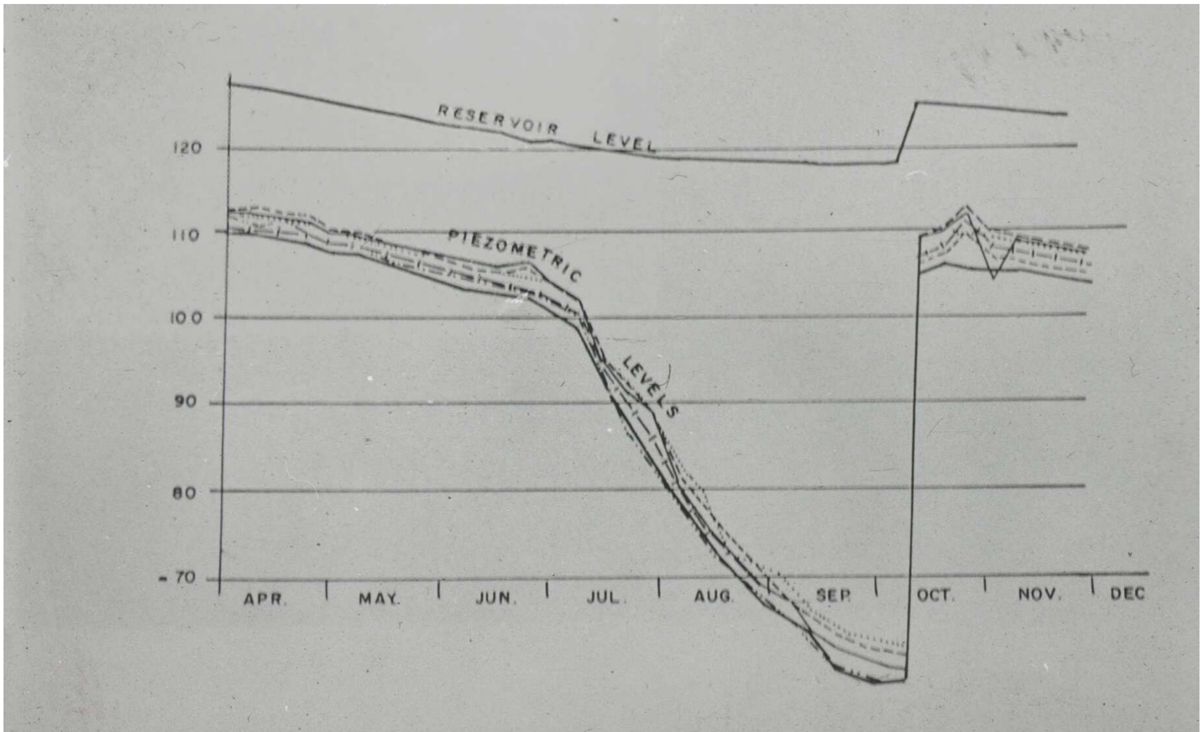

FIG. 2 - PIEZOMETRIC OSCILATION MARIA CRISTINA DAM.

Figure 13. Maria Cristina Dam. Reservoir levels and resulting piezometric levels