Hydrological processes on the land surface: A survey of modelling approaches

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Contents

1 The hydrological cycle

The land surface, the underground and the atmosphere are the three principal elements involved in that continuous water exchange through vertical and horizontal mass fluxes which is termed hydrological cycle (Figure 1). Precipitation and runoff are only the visible components of this process. Other components such as evaporation, infiltration, transpiration, percolation, groundwater recharge and discharge etc. are other important mechanisms of this cycle. The water exchange between land surface and atmosphere is based on the continuous evaporation of water from the Earth, the temporary storage of water in the form of water vapour in the atmosphere and the return of water to the land surface through several forms of precipitation such as rain, snow, sleet or hail. Even if these mechanisms are relatively simple to schematize, the interaction between land surface and underground and in particular ground water movement is not so easy to visualize.

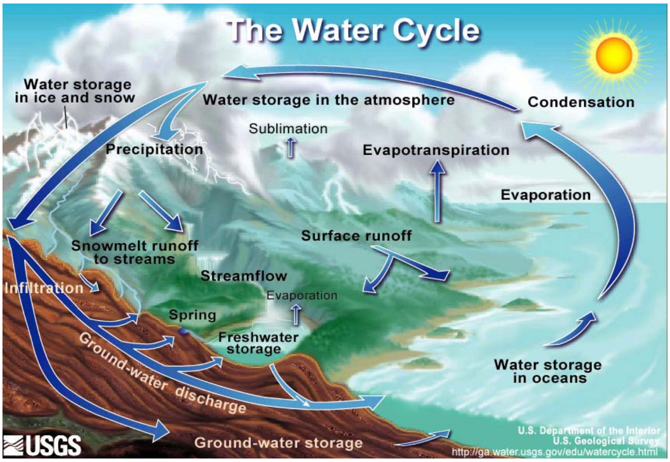

Figure 1. The water cycle (http://ga.water.usgs.gov/edu/watercycleprint.html).

1.1 *Physical elements of the hydrological cycle*

The top layer of land, crossed by roots and voids left by decayed roots and animals, is termed soil. It is generated by the alteration of the most superficial part of the rocks, determined by the physical, chemical and biological action of atmospheric factors and living organisms, which cause an extremely high variability of its features both in space and in time. Major soil physical properties, which influence hydrological characteristics, are the texture (the grain-size distribution of soil particles), the quantity of organic matter, the depth and the structure (the features of soil particles aggregation and of voids distribution, which depend on soil compaction).

Water below the land surface is substantially enclosed in two different type of zone: the unsaturated zone and the saturated one (Winter et al. 1998). In the unsaturated zone, which is also called the vadose zone and the zone of aeration, water and air are both contained into the void spaces between the grains of unconsolidated materials or into the rock fractures. In this zone water is held by capillarity forces and can't move by gravity or pumping. In the saturated zone the voids are completely filled with water, which is referred to as ground-water. The upper surface of the saturated zone, where pore water pressure is equal to atmospheric pressure, is termed water table. Immediately above the water table is the capillary fringe. The rise of water through the pores of the capillary fringe is strictly connected to the dimension and distribution of pores. Interconnected pores, which constitute the void spaces into the loose material, have the same behaviour of capillary tubes, so that the smaller the pore dimension, the greater is the rise. In loose material the thickness of the capillary fringe is variable with water alimentation: fringe thickness is greater when ground is drained by water table lowering, and is smaller when ground is made wet by water table rising.

According to the permeability of a porous medium we can distinguish four types of formations:

- the aquifer: as suggested by the Latin word etymology "Aqua" (water), and "ferre" (to carry, to bear), it is a hydro-geologic unit that can store, transmit, and yield groundwater to wells and springs;
- the aquitard: a semi-pervious hydro-geologic formation which can store water (so it may serve as a storage unit for ground water) but transmits water at a very slow rate compared to the aquifer;
- the aquiclude: a hydro-geologic unit which, although porous and capable of storing water, does not transmit it at rates sufficient to provide an appreciable supply for a well or spring;
- the aquifuge: a hydro-geologic unit which has no interconnected openings and, hence cannot store or transmit water.

Aquifers are generally classified into three main types: unconfined, confined and perched. The difference between them is related to the position of the impermeable layer and to the water pressure. A confined aquifer is a water-bearing zone bounded from above and from below (and so "confined") by a relatively impermeable layer (known as a confining bed) and exists under a pressure greater than atmospheric; as a consequence, the water level in a well (or a piezometer) that is drilled in such an aquifer is higher than the impervious surface that bounds the aquifer from above. When the piezometric surface is not only above the ceiling of the aquifer but also above the ground surface, the confined aquifer is referred to as an artesian aquifer.

An unconfined aquifer (or phreatic aquifer) has no confining layers between the zone of saturation and the land surface so that it receives water recharge directly from the infiltration of rainfall and surface water. Between a persistent water table of a phreatic aquifer and ground surface, can be located a saturated zone that overlies a low-permeability unit. This zone which is substantially smaller than the main unconfined aquifer, is the perched aquifer (Figure 2, American Ground Water Trust).

1.2 *Mechanisms of the hydrological cycle*

Precipitation that reaches the land surface is partially intercepted by vegetation. The amount of water intercepted by a plant, termed interception, largely depends on plant form. Water held on the leaf surface can trickle down reaching the ground or evaporate. Another part of water falls directly on the ground. Here water can evaporate with a rate that depends on wind speed, solar radiation, heat and humidity in the air, or enter into the soil, or flow down the land surface as runoff. The amount of water that penetrates into the surface of soil through the infiltration process is related to soil properties such as texture, structure, moisture content and in particular to soil permeability and porosity. Conditions at the soil‘s surface also influence infiltration. For example, a compacted soil surface or frozen soil conditions reduce the infiltration rate (Brown L.C.). Vegetation canopy protects the soil surface from compaction by heavy raindrops, slows down water that flows over the soil surface, and plant roots help to create openings in the soil. Infiltration rate is also related to the intensity and duration of precipitation.

To synthesize all these aspects we can briefly say that the rate at which water enters the soil from the surface is a function of water-input rate (snow melt and rain fall) and soil infiltration capacity (the maximum rate at which soil will accept incoming water). During a water input event, the infiltration rate declines exponentially and asymptotically to a near constant value (the equilibrium infiltration rate) limited by the saturation of the soil.

After entering the soil, water can take several paths. Some water becomes part of the soil storage. This water is not stationary; it is under the pull of gravity and moves downward at a rate that depends on various soil properties, such as permeability and porosity. Near the surface, in the root zone, some of this water is used by plants and eventually returned to the atmosphere as water vapour through a process called transpiration.

Water motion downward through the soil below the plant root zone and towards the underlying aquifer is called deep percolation. This water replenishes the groundwater supply and therefore this refilling process is called groundwater recharge.

As water percolates, some of it may reach a layer of soil or rock material that restricts downward movement, so that it moves laterally along this layer through the upper layers of soil above the water table and eventually discharge to a surface-water body, such as a stream or lake. This lateral movement of water is called interflow or subsurface flow.

Groundwater is in constant motion, although the rate at which it moves is generally slower than

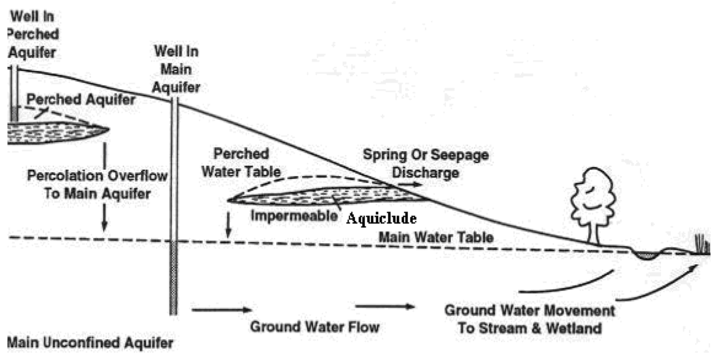

Figure 2. Perched aquifer in an unconfined aquifer.

that it would have in a stream because it must pass through the intricate passageways between free space in the rock (Commonwealth of Learning. Energy Research Group, 2003). Flow directions under the water table are determined by the potential energy distribution, and in particular water movement occurs from points with high hydraulic head (the sum of elevation and water pressure divided by the weight density of water) to points with low hydraulic head.

The rate at which groundwater moves through the saturated zone is controlled by ground permeability and by the hydraulic gradient that is defined as the difference in hydraulic head divided by the distance between two points on the water table. These terms are considered in the Darcy's Law that expresses, by means of an equation, the flow of a fluid through a porous medium.

In 1931 Lorenzo A. Richards applied to the Darcy's Law, developed for unsaturated flow in porous media, the requirements of mass conservation as suggested by Buckingham, and obtained a general partial differential equation governing unsaturated flow in a homogeneous, isotropic, non-deformable porous media under the conditions of no coupling between water and other pore fluids, and only viscous coupling between water and the solid matrix. Analytical solutions for Richards' equation are difficult to derive because of the highly nonlinear aspect of this partial differential equation (Tracy, 2007).

1.2.1 *Runoff generation: infiltration excess and saturation excess*

In 1930s the American hydrologist R. E. Horton enunciated the classical theory of runoff generation based on the infiltration excess (Figure 3) approach. The classical theory of Horton is based on the idea that it is possible to separate total stream flow into a quick-flow and a base-flow. Quick-flow is the portion of runoff that is "prompt" or rapid after the rain and it consists of overland flow. Base-flow is the contribution of groundwater. Horton first conceived "infiltration capacity" as a hyetograph separation rate that was generally applicable as a threshold for application to a rainfall intensity graph, the threshold intensity level being affected by soil conditions, seasons, and other phenomena (Horton, 1933). A few years later, Horton refined this "capacity" concept by referring it to an infiltration rate that declines exponentially during a storm, and published a conceptual derivation of the exponential decay infiltration equation (Horton, 1936a,b). He defined the maximum rate at which rain can be absorbed by the soil in a given condition as the infiltration capacity, hence overland flow is generated when the rainfall intensity

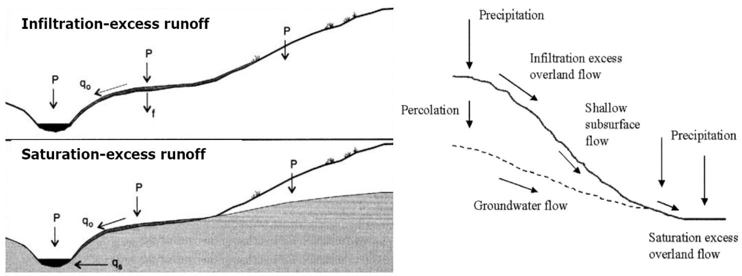

Figure 3. Infiltration and saturation excess runoff.

falling on the ground exceeds the infiltration capacity of the soil ("infiltration excess"), which in turn decreases with time during a rainfall event in proportion to the volume of infiltrated water. When the precipitation rate exceeds the infiltration rate of the soil, there is excess precipitation available for runoff and depression storage: depressions on the soil surface begin to fill until their storage is emptied and water flows out moving down-slope as overland flow or in defined channels (McIsaac).

The infiltration excess or Hortonian mechanism of runoff generation is controlled by two factors: rainfall intensity and soil properties like permeability. These factors can vary significantly in space and therefore it is possible that runoff generation does not occur uniformly on a catchment but only on partial areas. Furthermore the Hortonian scheme can't explain the overland flow generation for those soils which are so permeable that rainfall intensity almost never exceeds the infiltration capacity. For these cases hydrologists postulated in 1970s the saturation excess (Figure 3) mechanism, which is not controlled by rainfall intensity or infiltration capacity but by the available storage capacity of the underlying soils. During an event, rainfall that falls on the ground penetrates into the surface of soil and replenishes ground storage so that the water table rises. This process continues until water table intersects the land surface. It means that ground storage is locally filled or saturated, and in these points water begins to pond up at the surface, forming puddles and eventually starting to flow down-slope, or ex-filtrate, adding to the surface runoff. The mechanism, first identified in 1960s (Dunne, 1978), is particularly important where the infiltration capacity is relatively high and rain storms are of low to moderate intensity, and is controlled by the depth of bedrock and/or by the depth of water table. Both factors are strictly connected to the natural topography.

Saturation excess is the basis for the concept of dynamic or variable contributing areas that acknowledges that the spatial extent of saturation will vary seasonally depending on the relative rates of rainfall and evapo-transpiration. While in most watersheds both infiltration and saturation excess processes contribute to runoff generation, often one of them dominates.

Runoff generated by both of the above mechanisms tends to initially accumulate over the soil surface in depressions whose total volume is called surface storage capacity. This depends on geometric irregularities of the surface and on slope of the land. When surface storage is filled, water begins to overflow and actual runoff begins. Surface runoff runs down-slope increasing its depth and velocity, collecting in rills and gullies, before ending up in major streams and rivers

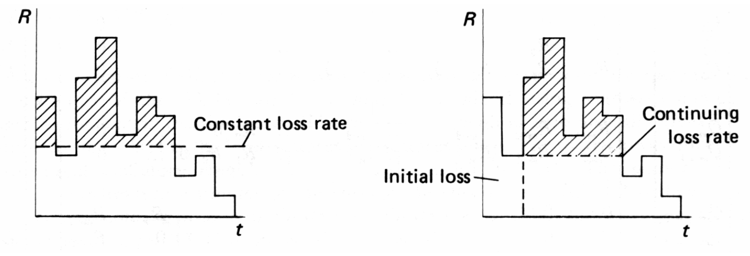

Figure 4. The Φ-index and W-Index methods. Left: Initial loss model. Right: Initial loss-constant loss model.

(McIsaac).

A number of models based on the infiltration excess theory are currently used in practical design applications or in mathematical models of hydrology. Some of them (for example the Φ-index or Constant Loss, CL method, which assumes a constant infiltration capacity throughout the event, and the W-Index or Initial Loss-Continuing Loss, IL-CL method, which assumes an initial loss of all storm rainfall up to a certain depth followed by a uniform or constant loss rate, the Φ-index), are based on infiltration indices (Figure 4). They ignore the fact that infiltration capacity usually decreases with time.

Other models, assuming a different variability law of infiltration capacity with time, use empirical equations, like the non-linear function proposed by Kostiakov (1932) and Horton (1933) or physically based equations like the Green-Ampt (1911) or Philip (1954) equations whose parameters are estimated by means of soil properties (porosity and hydraulic conductivity) measured in the field or in laboratory and tabulated for different type of soils. The physically based models, such as Green and Ampt, are based on simplified solutions to the Richards equation, which however still produce the exponentially decreasing relationship between infiltration capacity and cumulative infiltration (USACE, 1994: ch.6).

The Green and Ampt method (Figure 5) assumes the same simple soil model and initial conditions as that of the Richards' equation, a uniform soil profile of infinite extent, and constant initial water content. As the water content at the soil surface increases, the method models the movement of the infiltrated water by approximating the wetting front with a piston type displacement. Horton criticized the Green Ampt approach for not being physically realistic because it predicts an infinite infiltration capacity at t=0 and for neglecting the surface effects that he considered to be dominant (Beven, 2004).

Another widely used modelling approach is the Curve Number (CN: USDA-SCS, 1971; 1986) procedure (originally developed by the Soil Conservation Service of the US Department of

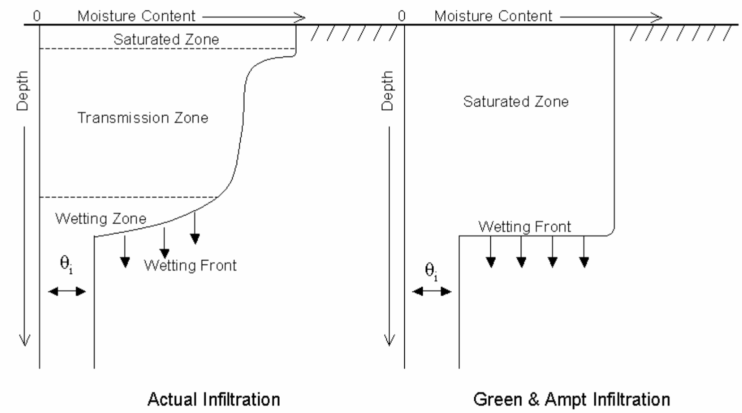

Figure 5. Comparison of moisture content distribution modelled by Green & Ampt and a typical observed distribution.

Agriculture – USDA-SCS, now Natural Resources Conservation Service or NRCS - in the 1950s). This is often considered among the methods based on infiltration excess overland flow, but this association is not completely correct (Garen and Moore, 2005). The CN method, originated as an empirical, event-based procedure for flood hydrology, and then adapted and used for both ordinary and large rainfall as well as with daily time series and events, is most correctly designed to predict "direct runoff". "Direct runoff" is a composite of channel runoff (direct precipitation on stream channels), surface flow (the overland flow) and subsurface flow. The two main parameters generally associated to the Curve Number model are: the Curve Number and the Initial abstraction. However initial abstraction is typically assumed as a function of the curve number. Curve numbers are determined by land cover type, hydrologic condition, antecedent moisture condition (AMC), and hydrologic soil group (HSG), which is related to the soil infiltration characteristics (Garen and Moore, 2005). This model assumes that runoff volumes, P, are proportional to the rainfall volumes, I, exceeding an Initial abstraction threshold, Ia, through a factor which is the ratio of the accumulated infiltration, F, to a storage capacity, S.

$$P = \begin{cases} \frac{F}{S}(I - I_a) & \text{for } I > I_a \\ 0 & \text{for } I \leq I_a \end{cases} \tag{1}$$

Together with the continuity equation that, for I>Ia states that:

$$I = F + I_a + P \tag{2}$$

the cumulative runoff volume becomes non-linearly related to the excess rainfall volume (I-Ia):

$$P = \frac{(I - I_a)^2}{I - I_a + S} \tag{3}$$

This non-linearity tends to disappear asymptotically for flood events with rainfall volumes exceeding the soil storage capacity S. Some authors like Beven (2000) consider the SCS-CN method similar to an infiltration excess procedure, if a spatially variable but time constant infiltration capacity is assumed in the catchment. However, because the threshold effects, peculiar to infiltration excess models, disappear in the Curve Number procedure as soon as the initial abstraction volumes are saturated, it is more obvious to consider it as a saturation excess model as do several other authors. The soil storage capacity, S, and watershed characteristics are related through an intermediate parameter, the curve number (commonly abbreviated CN) as shown in the below table where two expressions for different Unit Systems (English, with S expressed in inches, or Metric, with S expressed in mm) are reported (Ranzi et al., 2002).

In the following table (table 1), the expressions used by some of major methods for modelling infiltration are reported.

1.3 *The hydrograph: base-flow and quick-flow*

A hydrograph is the graphical representation of the time-series record of river levels or discharges, and shows the fluctuations in stream flow through time depending on the received water contributions. The hydro-graphic record essentially represents the net balance between gains to and losses from the stream (Australian Bureau of Rural Sciences). An observed hydrograph is commonly separated into two components: the quick-flow, which is the direct response to a

METHOD	FUNCTION	PARAMETERS	
		symbol	meaning
Horton	$i(t) = i_f + (i_o - i_f)e^{-\gamma t}$	i_0	initial intensity of infiltration
		i_f	final intensity of infiltration
		t	time from the start of the rainfall event
		γ	empirical constant value function of the type (nature) of the soil
Kostiakov	$i(t) = i_0^{-\alpha}$	i_0	initial infiltration capacity
		α	parameter function of soil conditions
Philip	$i(t) = 0.5st^{-0.5} + A$	s	sorptivity
		A	gravity component of the infiltration
Green/Ampt	$i(t) = k_s \left[1 + \dfrac{h_0 - h_f}{z_f(t)}\right]$	k_s	hydraulic conductivity at saturation
		h_0	surface pressure load
		h_f	pressure load at the humidity front
		z_f	humidity front depths
SCS	$P = \dfrac{(I - I_a)^2}{I - I_a + S}$	P	accumulated precipitation excess at time t
		I	accumulated rainfall at time t
		Ia	initial abstraction (initial loss)
	$I_a = 0.2S$	S	soil storage capacity
		CN	curve number
	$S = \left\{\dfrac{100 - 10CN}{CN}\right\}$ MKS system $S = \left\{\dfrac{25.4(100 - 10CN)}{CN}\right\}$ FPS system		SCS classifies soils into four hydrologic groups (A, B, C, D) based on their infiltration features, and defines three antecedent moisture conditions: I-dry, II-average moisture, and III-wet.

Table 1. Infiltration modelling methods.

rainfall event and includes the overland flow, the interflow and the direct contribution of precipitation onto the river surface; and the base-flow, which is mostly derived from groundwater.

There are many methods to separate base-flow component from a stream hydrograph. Graphical methods (Figure 6) are commonly used to plot the base-flow component of a flood hydrograph event. A relevant piece of information in graphical methods is the point where the base-flow intersects the falling limb, which is also the point where quick-flow has ceased. This point can be estimated by means of techniques based on the assumption of an empirical relation between the time of quick-flow ending and the catchment area, or on the hypothesis that the minimum stream-flow, immediately prior to the rising limb, is the constant discharge value which is the constant base-flow during a storm event (constant discharge method). The constant slope method connects the start of the rising limb with the inflection point on the receding limb. This method assumes an instant response in base-flow to the rainfall event. The concave method uses the declining hydrographic trend evident prior to the rainfall event until the peak time and then connects this point to the inflection point on the falling limb (Brodie and Hostetler, 2005).

These simple graphical methods are often focused on separating base-flow from a flood hydrograph and aimed at the estimation of the surface runoff component of a flood event, so that they are also named event-based separation methods. Beside them, there are also filtering separation techniques that separate the base-flow component of the stream-flow time series using data filtering procedures to derive a base-flow hydrograph from the processing of the entire stream hydrograph. They are designed to generate a base-flow hydrograph for a long-term period of observations so that they are also known as continuous separation techniques. These methods aim at generating an objective, repeatable and easily automated index that can be related to the base-flow response of a catchment (Nathan and McMahon, 1990), like the base-flow index (BFI) or reliability index which is the long-term ratio of base-flow to total stream-flow.

Other indices include the mean annual base-flow volume and the long-term average daily base-flow (Smakhtin, 2001). The Tennant (or Montana) method (1976) is the most common base-flow separation method applied worldwide, and has been used either in the original or in a modified form (Tharme, 2003), because of its simplicity and ease of use. The Tennant method uses a percentage of the mean annual flow (MAF) for two different six month periods to define the conditions of flow.

Other well-known techniques of this type are the smoothed minima method (FREND,1989) which uses the minima of 5-day non-overlapping periods derived from the hydrograph and generate the base-flow hydrograph by connecting a subset of points selected from this minima series, and recursive digital filters (Nathan and McMahon, 1990; Smakhtin and Watkins, 1997; Smakhtin

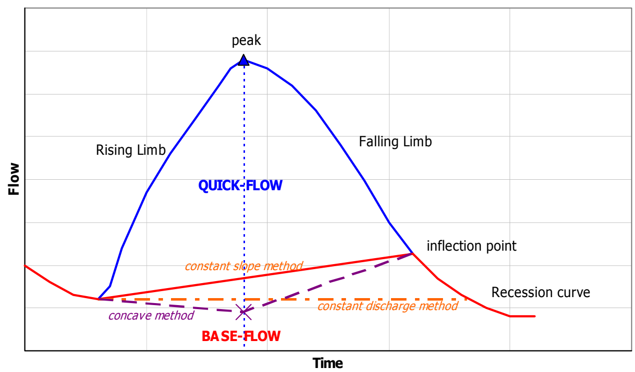

Figure 6. Examples of simple graphical base-flow separation methods.

2001) which are routine tools in signal analysis and processing used to remove the low-frequency and high-amplitude “flood flow” signal, and to derive the high-frequency and low-amplitude base-flow signal.

1.4 *Hydraulic and hydrologic approaches for channel routing*

A flood wave moving through a river reach is subject to several modifications, reflected by changes in hydrograph shape and timing, because of the effects of storage and flow resistance. Flood routing is the process used to model the temporal and spatial variations of a flood hydrograph through a river reach or reservoir. Routing techniques are generally classified into two categories, corresponding to two different approaches: hydraulic routing and hydrologic routing.

Hydraulic routing techniques are based on the solution of the partial differential equations of unsteady open channel flow, the Saint Venant equations, assuming the principles of mass and momentum conservation. Solving together the continuity equation and momentum equation with the proper boundary conditions, the complete or full dynamic wave equations are obtained (USACE, 1994: ch.9).

In case of 1-D unsteady flow in open channel, the Saint Venant system consists of the continuity (4) and the momentum (5) equations:

$$A\frac{\partial V}{\partial x} + VB\frac{\partial y}{\partial x} + B\frac{\partial V}{\partial t} = q \tag{4}$$

$$S_f = S_0 - \frac{\partial y}{\partial x} - \frac{V}{g}\frac{\partial V}{\partial x} - \frac{1}{g}\frac{\partial V}{\partial t} \tag{5}$$

where A = cross-sectional flow area; V = average velocity of water; x = distance along channel; B = water surface width; Y = depth of water; T = time; Q = lateral inflow; S_f = friction slope; S_0 = channel bed slope; g = gravitational acceleration. The second, third and fourth term on the RHS of Equation 5 represent pressure differential, convective acceleration and local acceleration respectively.

Depending on the relative importance of its various terms, the momentum equation can be simplified for various applications. These simplifications and the consequent approximations are justified if specific terms in the momentum equation are small in comparison to the bed slope. The most common approximations of the momentum equation are the following:

$$S_f = S_0 \tag{6}$$

$$S_f = S_0 - \frac{\partial y}{\partial x} \tag{7}$$

$$S_f = S_0 - \frac{\partial y}{\partial x} - \frac{V}{g}\frac{\partial V}{\partial x} \tag{8}$$

The Kinematic Wave approximation (Eq. 6): gravitational and frictional forces achieve a balance, and changes in velocity and depth with respect to time and distance are small in magnitude when compared to the bed slope of channel. It means that the momentum of flow can be approximated with a steady uniform flow assumption as described by Manning’s or Chezy’s equations.

The Diffusion Wave approximation (Eq. 7): it includes the pressure differential term that describe the attenuation or diffusion effect of the flood-wave and allows to specify a downstream boundary condition in the routing reach to account of backwater effects. The momentum of flow can be approximated with a steady non-uniform flow assumption.

The Quasi-Steady Dynamic Wave approximation (Eq. 8): it includes the convective acceleration term that accounts for changes in velocity with respect to distance. This approximation is usually applied to steady flow-water surface profile computations. The assumption that the local acceleration term is negligible introduces an error greater than the above approximations, that exclude both terms representing changes of velocity in time and distance.

Hydraulic routing techniques can incorporate backwater effects as well as internal boundary conditions, such as discontinuities associated with culverts, bridges, and weirs. When there are downstream controls that will have an effect on the routing process through an upstream reach, the channel configuration should be treated as one continuous system. This can only be accomplished with a hydraulic routing technique while the hydrologic approaches are not adequate. However hydraulic routing techniques require a large amount of data on channel cross section geometry, gradient, hydraulic roughness of channel and floodplain.

Hydrologic routing techniques employ the continuity equation and an analytical or an empirical relationship between storage, outflow and possibly inflow. In its simplest form, the continuity equation can be written as inflow minus outflow equals the rate of change of storage within the reach. One of the simplest routing applications is the analysis of a flood-wave that passes through an unregulated reservoir. The inflow hydrograph is known, and the outflow hydrograph from the reservoir has to be computed. In this case only a relationship between storage and outflow can be developed. The storage-outflow relationship provides the outflow for any storage level and it is used in combination with the equation defining storage routing, which is based on the principle of conservation of mass. The Modified Puls routing method, also known as storage routing or level-pool routing, applied to reservoirs consists of a repetitive solution of the continuity equation. It is based on solving the continuity equation by trial and error using the storage-outflow relationship to obtain the correspondent outflow or storage, having fixed one of them. It is assumed that the reservoir water surface remains horizontal, and therefore, outflow is a unique function of reservoir storage. To apply the modified Puls method to a channel routing problem, the storage within the river reach is approximated with a series of “cascading reservoirs”. Each reservoir is assumed to have a level pool and, therefore, a unique storage-discharge relationship. The cascading reservoir approach is capable of approximating the looped storage-outflow effect when evaluating the river reach as a whole. Indeed the storage and water surface slope within a natural river reach, for a given outflow, is greater during the rising stages of a flood-wave than during the falling ones. Therefore, the relationship between storage and discharge at the outlet of a channel is not a unique relationship, rather it is a looped relationship. So in the application of the Modified Puls method to rivers the rising and falling flood-wave are simulated with different storage levels in the cascade of reservoirs, thus producing a looped storage-outflow function for the total river reach (USACE, 1994: ch.9).

The Muskingum method describes the storage within a reach as composed by two parts: the prism storage, which is the storage under the steady-flow water surface profile, and the wedge storage which is the storage under the actual water surface profile. The wedge storage is positive and is added to the prism storage during the rising stages of flood-wave but is negative and subtracted from the prism storage during the falling stages of flood-wave (Figure 7). The Muskingum routing equation defines the storage (S) in the reach as a linear function of weighted inflow (I) and outflow (O): S=KO + KX (I-O).

The weights of this equation are the coefficient K, corresponding to the travel time of the flood-wave from upstream to downstream end of the channel reach, and the parameter X, a dimensionless weighting factor, ranging from 0.0 to 0.5, expressing the relative effect of inflow and outflow on reach storage. The routing parameters K and X are related to flow and channel characteristics. K is indeed a function of channel length and flood wave velocity and X accounts for storage portion of routing. These parameters in the Muskingum method are determined by calibration using stream-flow records.

The Muskingum-Cunge method is an alternative method, closely related to the Muskingum procedure. It is a non linear coefficient method that accounts for hydrograph diffusion based on physical channel properties and the inflowing hydrograph. The main advantages of this procedure is that the parameters are more physically based than other hydrologic techniques. Furthermore the method has been shown to compare well against the full unsteady flow equations over a wide range of flow situations (Ponce, 1983 and Brunner, 1989) and the solution is independent of user-defined computation interval. However its main limitations are that it can't account for backwater effects, and that it begins to diverge from the full unsteady flow solution when very rapidly rising hydrographs are routed through flat channel sections.

In general hydrologic routing models, in the absence of significant backwater effects, offer the advantages of simplicity, ease of use and computational efficiency. Hydraulic models are more physically based since they only have one parameter (the roughness coefficient) to estimate or calibrate.

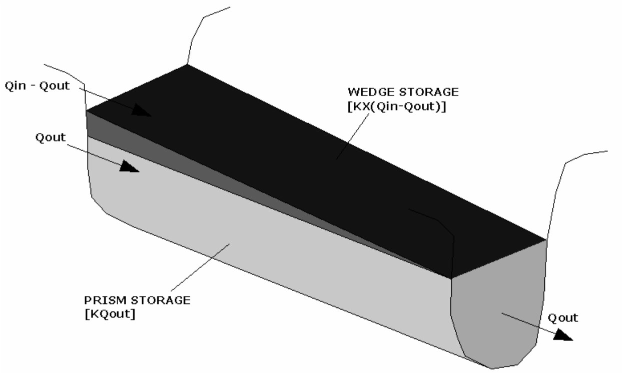

Figure 7. Prism and wedge storages in a reach segment.

2 Classification of hydrological models

The modelling of the physical phenomena which govern the response of a river basin to meteorological forcing can only be achieved with great difficulty, due to the complexity and variability in time and in space of the processes and elements involved in the transformation of rainfall into runoff.

Hydrological models are simplified systems to quantify the processes of the hydrological cycle in an entire river basin or parts of it. They are based on a set of interrelated equations that try to convert the physical laws, which govern extremely complex natural phenomena, to abstract mathematical forms. Any hydrological model emphasizes some aspects which are considered relevant instead of others considered of secondary importance, and should be sufficiently comprehensible and easy to be used and in the same way sufficiently complex to represent the physical studied problem. Moreover different varieties of models can be used, depending upon the conceived output, the existing database, input variables and required analysis.

According to Singh (1995) the rainfall-runoff models can be classified according to their degree of representation of the physical processes and to the spatial and temporal description (Melone et al., 2005).

2.1 *Representation of physical processes*

Representation of physical processes described by rainfall-runoff models can either be based on a simple mathematical link between input and output variables of the basin, or include the description of basic processes involved in the runoff generation and evolution. The two categories are often referred to as data-driven models and knowledge-driven models. These two philosophies correspond to different ways of using a priori knowledge, the former corresponding to the reduction of the model to a pure stochastic process without any cause-effect relationships, and the latter to a complete description of the water cycle dynamics based on the differential equations governing the system’s behaviour. The knowledge-driven models implies a good knowledge about the physical phenomena to be modelled. They are generally structured in two basic processes: the mathematical modelling which concerns the description of the natural phenomena by mathematical equations, and the numerical solution which involves computing an efficient and accurate solution of the mathematical equations. The so-called data-driven models are used in situations when the knowledge of the system under study is limited but there is enough data concerning some of variables characterizing a particular process. These data can be used to represent the relationship between the input and the output associated to the system considered as a “black box”: the physical principles that govern the system are not made explicit but mathematical and statistical concept are used to link a certain input with a certain output. The data-driven models include various statistical approaches, neural network, machine learning, fuzzy systems and chaos theory and can be used in combination with physically based models (hybrid models) especially when there are not enough data to train data-driven models or when physically based models do not exhibit the needed accuracy or are inadequate to model particular situations.

Summarizing, the different approaches used to treat a-priori knowledge lead to the distinction between:

– knowledge-driven, theoretical or physically-based (white-box) models;

– conceptual (grey-box) models;
– data-driven, empirical or stochastic (black-box) models.

All these three types of mathematical models represent different levels of approximation of reality: physically-based models are the closest representation of the real system and try to incorporate as many components of actual physical processes as possible; empirical models do not aid in physical understanding so that their parameters may have some little physical significance and can be estimated only by using concurrent measurements of input and output variables; conceptual models may be considered intermediate between the two previous type of models, because they consider physical laws but in highly simplified form.

2.1.1 *Spatial description*

On the basis of spatial description, the hydrological models can be classified into three main categories: lumped models, semi-distributed models and distributed models. In lumped models the whole catchment is considered as a single entity, spatial variations are averaged or ignored and basin response is evaluated only at the outlet. In semi-distributed models spatial variation is partially allowed dividing the basin into a number of smaller entity or sub-basins. In distributed models spatial variability of processes, input, boundary conditions, watershed characteristics and output is defined by the user and is strictly dependent by the resolution of available data.

Lumped conceptual models are characterized by simple structure, minimum data requirements, fast set up and calibration and by being easy to use (Cunderlik, 2003). They can be less responsive to very intense but short-lived rainfalls. In addiction, these models are not usually applicable to event-scale process, while in the discharge prediction, they can provide just as good simulation as complex physically based models.

Semi-distributed models are more physically-based than lumped models and less demanding on input data than fully distributed models. They are frequently used for flood control but their disadvantage is calibration (valid only for the actual conditions) so this type of model is not indicated for real-time forecasting. Semi-distributed continuous models are often used to evaluate monthly and annual water balance.

Distributed physically-based models can provide the highest accuracy in the modelling of precipitation-runoff processes (Cunderlik, 2003). Parameters of these models are fully spatially-varied at a given resolution and therefore require considerably more input data, often unavailable, than semi-distributed models and they have a direct physical interpretation. Distributed physically-based models are typically applied in case of un-gauged catchments, to analyze catchment changes, spatial variability in catchments inputs and outputs, movement of pollutants and sediments.

Physically-based distributed-parameter models were developed starting from the 1980s at the same time with the development of Remote Sensing Techniques and Geographical Information System Tools (Xu, 2002). While traditionally hydrological data are point measurements, remote sensing data (aerial photography, radar, and satellite imaging) incorporate spatial information, which is typically the information obtained from maps or aerial photography, and gives the possibility to observe an entire areas or river basins rather than single points. Particularly for large regional scales and in areas where the availability and quality of hydrological data are limited, remote sensing information appears to offer useful data that can not only fill some of the gaps in data availability, but also supply data at the appropriate scale for distributed hydrological models. Furthermore GIS has provided a new environment to develop distributed hydrological models and represents a powerful tool to organize, process and visualize spatial data. The pixel format of digital remote sensing data makes it ideal for merging with Geographical Information Systems (GIS) and for being used in hydrologic and runoff modelling (U.S. National Report to IUGG, 1991-1994). Two are the main applications for remote sensing coupled with GIS tools: to determine watershed geometry, drainage network, and other map-type information for distributed hydrologic models and for empirical flood peak, annual runoff or low flow equations; and to provide input data such as soil moisture or delineated land use classes that are used to define runoff coefficients.

2.1.2 *Temporal description*

The time factor in rainfall-runoff models can be considered both referring to changes of input-output relationships with time (time-variant models), and to the temporal resolution and duration of hydrologic simulations. More specifically, models can be considered either as event-driven models (capable of simulating short-term events), or as continuous-processes models (capable of simulating continuous events). An event model simulates a single storm. The duration of the storm may range from a few hours to a few days. Event models are used particularly in cases where direct runoff is a major contributor to total runoff, hence these models have no provisions for moisture recovery between storm events and, therefore, are not suited for the simulation of dry weather flows. A continuous model simulates instead a longer period, predicting watershed response both during and between precipitation events. Continuous models take explicit account of all runoff components, including direct runoff and indirect runoff and are suited for simulation of daily, monthly, or seasonal stream flows, usually for long term runoff volume forecasting and estimates of water yield (Cunderlik, 2003).

Another distinction among hydrological models is sometimes made according to their application for different purposes. Hydrological models, in fact, may be used in design problems, to take decisions and for planning, as an evaluation tool for water resources management, but also for research purposes.

2.2 ***Guidelines for model choice***

The selection of a particular model is a key issue to get satisfactory answers to a given problem. Today available models are really numerous and generally it’s more frequent to choose a model from a list of existing ones than to develop a new model. Although there are no clear rules for making a choice between models, some simple guidelines can be stated. Starting from the studied physical system, the first step is to define the problem and determine what information is needed and what questions need to be answered. This means that it is necessary to evaluate the required output, the hydrologic processes that need to be modelled, the availability of input data. Subsequently the simplest method that can provide the answer to the questions has to be chosen. In particular it’s necessary to identify the simplest model that will yield adequate accuracy, bearing in mind that model complexity is not synonymous with the accuracy of the results, that the model has to be characterized by flexibility, by the possibility of making it applicable under various spatial and temporal conditions and that increased accuracy has to be worth the increased effort. A possible methodology for selecting a rainfall-runoff model is proposed by Anderson and Burt (1985) and showed in the following figure (Perrin et al., 2002).

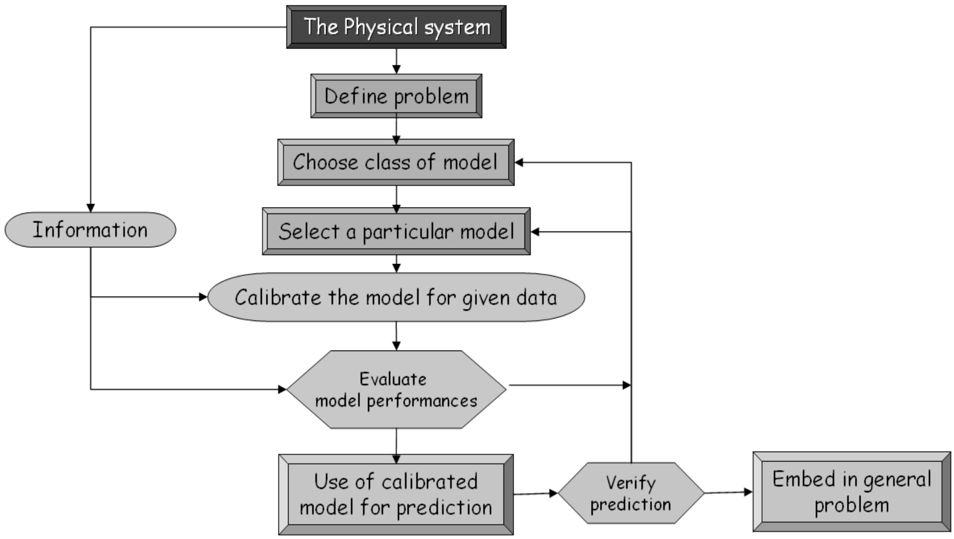

Figure 8. Anderson and Burt (1985) suggested scheme for selecting rainfall-runoff model.

In the scheme in figure 8 the two steps of calibration (calibrate the model for given data) and validation (evaluate model performances) are present. The parameter values that characterize a model have to be determined on the basis of measurements, or available data. Usually, this is possible only for some of the parameters and the others have to be estimated by calibration. In this case, their values are changed (manually or by some automatic calibration algorithm) until a satisfactory agreement between observed and simulated runoff is achieved. Before the model with its calibrated or measured parameter values is used to solve the problem, its capability to simulate an independent period should be tested. This test usually is called validation.

3 Commonly used hydrological models

In this section a brief and general description of some most frequently applied hydrological models is presented. This list of 8 models derives from a survey conducted within the FORALPS project, to assess the state-of-the-art of rainfall-runoff models and to identify and suggest the most useful for applications in the Alpine Space. A description of each model is provided, considering their capability to describe flow conditions and to represent spatial and temporal variability. The typical sector of application, the processes of hydrological cycle that are modelled, and tools for calibration and pre- or post-processing of input and output data are described. Additional information regarding program features, references and availability of the sofware are also given.

Some of these water models are not only aimed at flood forecasting and flood impact analysis, but more in general at environmental impact studies, integrating quality and quantity concerns. ON one hand, hydrologic processes are influenced by various factors, e.g. the spatial variability of soils, topography, land use and cover, climate, and by changes produced by human presence on territory. On the other hand, contaminants are transported by runoff to surface waters and by infiltration and deep percolation to groundwater. Therefore, models of hydrologic processes are often at the core of studies about water quality and quantity, two aspects of water resource management that are strictly interrelated.

3.1 *HEC-HMS (Rel. 3.1.0)*

3.1.1 *General information*

The Hydrologic Modelling System HEC-HMS is a program developed by the Hydrologic Engineering Centre (HEC) of the US Army Corps of Engineers. It is the successor to and the replacement for HEC's HEC-1 program (US-ACE, 1998), whose characteristics are improved with additional capabilities for distributed modelling and continuous simulation. The program is designed to simulate the precipitation-runoff processes and routing processes, both natural and controlled.

3.1.2 *Flow Conditions/Representation in Space and Time*

HEC-HMS is designed to simulate both single events and continuous long periods. HMS models (US-ACE, HEC, 2000; 2006) are typically lumped, but also grid cell methods are included in the program. The physical representation of a watershed is accomplished with a basin model, where hydrologic elements (sub-basin, reach, junction, reservoir, diversion, source, and sink) are connected in a dendritic network to simulate runoff processes. However, spatially distributed runoff can be computed with the quasi-distributed linear transform (ModClark) of cell-based precipitation and infiltration.

3.1.3 *Type of application*

HMS is applicable in a wide range of geographic areas for solving the widest possible range of problems. These include water supply in large river basins, flood hydrology, and small urban or natural watershed runoff.

3.1.4 *Representation of hydrological processes*

HMS uses a separate model to represent each component of the runoff process, as illustrated in Figure 9. Thus it includes separate models to compute runoff volume, direct runoff (overland flow and interflow), base-flow and channel flow.

– precipitation. The program provides precipitation-specification options which can describe an observed (historical) precipitation event, a frequency-based hypothetical precipitation event, or an event that represents the upper limit of precipitation possible at a given location. HMS also provides a gridded precipitation method that uses radar rainfall data and retrieves gridded records from a database.

– snow melt and accumulation. Currently, only one snowmelt method is available. It is a temperature index method that dynamically computes the melt rate based on current atmospheric conditions and past conditions in the snow-pack. A gridded Temperature Index method is designed to work with the gridded ModClark transform method.

– reservoirs. The reservoir model that is included in the program is appropriate for simulating any configuration of outlets and ponds. However, the model assumes that outflow is inlet-controlled, so the reservoir model should not be used if the configuration of the reservoir and outlet works is such that the outflow is controlled by a backwater effect.

– evaporation and transpiration are combined, and collectively referred to as evapo-transpiration (ET), both in the Soil-Moisture Accounting Model (SMA) model and in the meteorological input to the program. Monthly-varying ET values are specified, along with an ET coefficient. The potential ET rate for all time periods within the month is computed as the product of the monthly value and the coefficient. An implementation of the Priestley-Taylor method is also available.

– interception and infiltration. In the HEC-HMS computer model, the land surface interception, depression storage, and infiltration are referred to as loss rates. Both event and continuous methods are available to account for cumulative losses. The event models included are: IC, SCS curve number, gridded SCS curve number, exponential, Green-Ampt, and Smith-Parlange (Smith and Parlange, 1978). The one-layer deficit constant method can be used for simple continuous modelling. The five-layer Soil Moisture Accounting method can be used for continuous modelling of complex infiltration and evapo-transpiration environments.

– base-flow. Five alternative models of base-flow are included: the recession method, which gives an exponentially decreasing base-flow from a single event or from multiple sequential events; the constant monthly method, which works well for continuous simulation; the linear reservoir method, which conserves mass by routing infiltrated precipitation to the channel; the nonlinear Boussinesq method, that provides a response

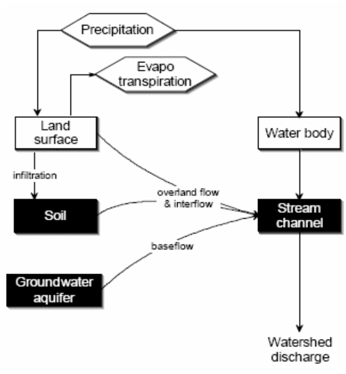

Figure 9. Typical HEC-HMS representation of watershed runoff

similar to the recession method but allows estimating parameters from measurable qualities of the watershed.

– channel routing. Six routing models are available to model channel flows: the Lag method; the traditional Muskingum method, included along with the straddle stagger method for simple approximations of attenuation; the Modified Puls method that can be used to model a reach as a series of cascading level pools with a user-specified storage-discharge relationship; the Kinematic-wave method; the Muskingum Cunge method, either in a standard configuration (channels with trapezoidal, rectangular, triangular, or circular cross sections) or in a 8-point cross section configuration.

– surface runoff. The program provides two options for direct runoff, including overland flow and interflow: empirical models (the traditional unit hydrograph UH models) and a conceptual model that consists of a kinematic-wave model of overland flow.

3.1.5 *Additional information: Program Features and Related Tools.*

HMS is a public domain software. Free download of software and documentation can be performed from the website: http://www.hec.usace.army.mil. The model is a completely integrated working environment including a database, data entry utilities, computation engine, and results reporting tools. A graphical user interface allows the user seamless movement between the different parts of the program. Time-series, paired, and gridded data are stored in the Data Storage System HEC-DSS.

HEC-HMS has automatic calibration capabilities and includes the ability to chose one of six different objective functions and one of two different search algorithms (USACE, 2006) (the univariate-gradient search algorithm and the derivative-free minimization algorithm of Nelder and Mead (1965). HEC-HMS supports constrained optimization for both search methods. Moreover it can take advantage of a hydrologic GIS pre-processor developed by the same Hydrologic Engineering Centre: HEC-GeoHMS. The HEC-Geospatial Hydrologic Modelling Extension (GeoHMS) is a public-domain extension to ArcView GIS and Spatial Analyst extension (USACE, 2000a) developed to use available digital geospatial information to construct hydrologic models more expediently than using manual methods.

3.2 ***WATFLOOD***

3.2.1 *General information*

WATFLOOD was developed for the Surveys and Information Branch of the Ecosystem Science and Evaluation Directorate of Environment Canada by Nicholas Kouwen of the Department of Civil Engineering at the University of Waterloo (Canada). The WATFLOOD programs are mostly a set of FORTRAN programs for DOS, compiled in Visual Fortran Ver. 6. It is a Visual Basic program that is used for data input and output and calls the DOS programs in a shell. All computations can be run in DOS, as well as on various Unix platforms (SUN Solaris, SGI and Linux systems). All programs have been converted to the Fortran 95 standard with dynamic memory allocation. The system is completely modular but has a consistent data structure throughout and has been under continuous development since 1972.

3.2.2 *Flow Conditions/Representation in Space and Time*

WATFLOOD is a distributed hydrologic model. It has been used with grid sizes from 1 to 25 km and for watershed areas from 15 to 1700000 km^2. This integrated set of computer programs has been applied to forecast flood flows for watershed having response times ranging from one hour to several weeks (Narula et al., 2002). Continuous simulation can be carried out by chaining up to 36 events. The SIMPLE (SPL) hydrological modelling component uses the Grouped Response Unit (GRU) method to model large watersheds. The basic premise of the GRU method is that vegetation and/or land use is the predominant hydrological indicator of hydrological response.

3.2.3 *Type of application*

The model is aimed at flood forecasting, climate change and environmental impact studies.

3.2.4 *Representation of hydrological processes*

The model SPL9 is a physically-based simulation model of the hydrologic budget of a watershed. It was designed for distributed modelling using remotely sensed data, particularly from remotely sensed land cover maps and weather radar. The SPL is a three layer model composed by the UZ-Upper Zone Storage (saturated), the IZ-Intermediate Zone Storage (unsaturated) whose moisture content the initial moisture refers to, and the LZ-Lower Zone Storage (saturated).

- <u>precipitation</u>. WATFLOOD uses rain gauge point data, radar or numerical weather models data. Rain gauge data are also used to adjust radar data, especially to fill in missing radar rainfall measurements. All data are converted in a grid geo-referenced format. As with rainfall, temperatures are required for each grid; therefore these data, which are generally collected or predicted at specific point locations, are converted into a grid format. In older versions, only daily maximum and minimum temperature values are required and the program calculates hourly data using a simple cosine function between highs and lows. In the current SPL9 version, only hourly temperatures are used.
- <u>snow melt and accumulation</u>. In WATFLOOD, snow-free and snow covered areas are modelled separately. Energy to melt snow is applied only to the snow covered area and as the snow covered area is reduced, surface storage and upper zone storage for the previously snow covered area is transferred to the snow free area. WATFLOOD uses a temperature index algorithm based on the National Weather Service River Flow Forecast system by Anderson (1973). A radiation-temperature index model, was recently incorporated (but not available to users) into the WATFLOOD model (Hamlin, 1996). It is a combination of the temperature index and the surface radiation budget, as proposed by Martinec and de Quervain (1975), Ambach (1988), and Martinec (1989).
- <u>reservoir</u>. Three reservoir routing options are available in SPL9. In the first, the user simply enters the reservoir releases which are then routed downstream. The second method is a storage-discharge lake routing method and the third is an external method, where reach numbers are specified and contributions to these reaches are printed to a file in the format used by the NWS Dynamic Wave Operational Model (DWOPER).
- <u>evaporation and transpiration</u>. Three methods for estimating potential evapo-transpiration (PET) are provided: the Priestley-Taylor equation (when radiation data are available), the Hargreaves equation (when only temperature data are available), and the original method

of estimating evapo-transpiration from published values (when neither temperature nor radiation data are available).

– interception. The Linsley equation is applied to model interception of precipitation by vegetation. This method calculates the total possible interception as the sum of the maximum canopy storage and the amount of interception storage, IET, during the storm event.

– infiltration. The infiltration process is modelled by the Philip (1954) formula.

– base-flow. Infiltrated water is stored in a soil reservoir referred to as the Upper Zone Storage (UZS). Water within this layer percolates downward or is ex-filtrated to nearby water courses, and is called interflow. Interflow is represented by a simple storage-discharge relation. Ground water, or Lower Zone Storage (LZS), is replenished by recharge from the UZS according to a power function. A ground-water depletion function is used to gradually diminish the base flow.

– channel routing. The routing of water through the channel system is accomplished using a storage routing technique. The method involves an application of the continuity equation and of Manning's formula to relate flow to the storage. The channel inflow is the sum of the discharge entering the channel at the upstream boundary, and any lateral flow added or removed by hydrologic processes during the current time step. The lateral flow is the sum of interflow, overland flow, base flow and precipitation falling on the stream, minus evaporation. The original cross section shape for WATFLOOD was triangular. The program was then modified to make the channel section more realistic, considering that most river cross sections are rectangular with flat bottoms and near vertical sides. The over-bank cross section is assumed to be triangular with a constant width to depth ratio of 100:1, while the main channel is assumed rectangular (Figure 10). A relation to give the bank-full channel cross sectional area at any point in the basin is required. Different Manning's n parameters can be given for the main channel and the flood plain.

– surface runoff. When the infiltration capacity is exceeded by the water supply, and the depression storage has been satisfied, water is discharged to the channel drainage system. The relationship employed is based on the Manning's formula where a combined roughness and channel length parameter is used to express the relative effects of surface

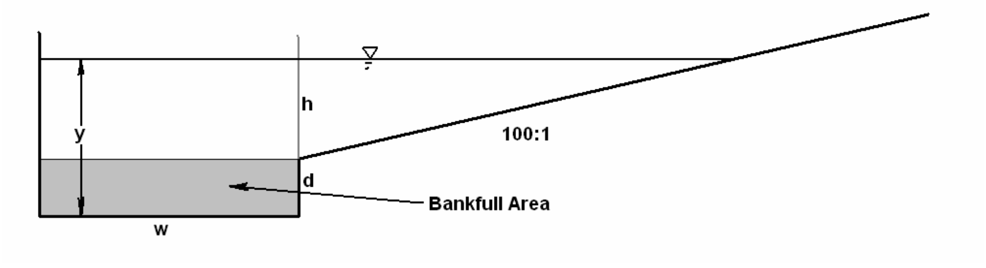

Figure 10. WATFLOOD representative river cross section.

roughness and drainage density.

3.2.5 *Additional information: Program Features and Related Tools*

WATFLOOD is a commercial software package. However a simplified and limited version of WATFLOOD, WATFLOOD LITE is available for student use with no key requirement. The documentation and the oftware can be downloaded from the website:

http://www.civil.uwaterloo.ca/Watflood/index.htm.

In addition to SPL9, there are a number of support programs to provide for data preparation and output presentation. Example of WATFLOOD tools are: RADMET, that converts the radar data file to a SPL9 compatible format; CALMET, that fills in missing radar data with rain gauge data if available and can also be used to adjust the radar data using the Brandes method if the parameters are set to do so; BSN, that may be used to assemble and create a “basin file” for SPL9; and PLOTHYD, which is a program to plot hydrographs on a colour monitor. The Hooke and Jeeves (1961) automatic pattern search optimization algorithm taken from Monro (1971) is available. The program can be run to automatically determine which combination of parameters fits the measured conditions best. The parameters considered for optimization are soil permeability, overland flow roughness, channel roughness, depression storage, and an upper zone depletion factor.

3.3 ***TOPMODEL***

3.3.1 *General information*

The development of TOPMODEL was initiated by Kirkby at the School of Geography, University of Leeds. The model was further developed by Keith Beven at the Lancaster University (Beven and Kirkby, 1979). Since 1974 there have been many variants of TOPMODEL, but never a definitive version (Beven, 1997).

3.3.2 *Flow Conditions/Representation in Space and Time*

TOPMODEL is a semi-distributed and partly physically based model. It is a topography based hydrological program and allows single or multiple sub-catchment calculations with average rainfall and potential evapo-transpiration inputs to the whole catchment (Narula et al., 2002). This model accounts for variability in the hydrological response of different areas of a catchment by using an index of catchment wetness based on topography. This means that areas possessing the same value of the topographic index are assumed to have the same hydrological behaviour. The local topography index is defined as ln (a/tan b) where (a) is the draining area per unit contour length through a location in the catchment, and (tan b) is the hydraulic gradient, i.e. the slope of the catchment surface at the same location. TOPMODEL is run on a daily time step for days with no precipitation and on an hourly time step for days with precipitation. For days with precipitation, the total daily precipitation was distributed over a randomly selected number of consecutive hours within the day (Wolock, 2003). The topographic index of TOPMODEL is scale dependent so that parameter values and consequent results are strictly dependent and sensitive to grid size or DEM resolution. For this reason the model requires a high quality DEM, without sinks. The recommended resolution of the grid size is not greater than 50m.

3.3.3 *Type of application*

TOPMODEL is mainly used to simulate humid or dry catchment responses, predict flood frequency, analyze land surface to atmospheric interactions and predict geochemical characteristics.

3.3.4 *Representation of hydrological processes*

TOPMODEL is defined as a variable contributing area conceptual model in which the dynamics of surface and subsurface saturated areas is estimated on the basis of storage discharge relationships established from a simplified steady state theory for down-slope saturated zone flows. The theory assumes that the local hydraulic gradient is equal to the local surface slope and implies that all points with the same value of the topographic index, will respond in a hydrologically similar way. The soil profile is defined by a set of stores. In the root zone storage (SRZ) rainfall infiltrates until the field capacity is reached. An additional interception and surface storage may be necessary, to take in account the canopy interception of precipitation. Once the moisture content of the root zone storage has exceeded the field capacity, the gravity drainage store (SUZ) starts to fill and continues until the water content reaches the saturation. This gravity drainage storage links the unsaturated and saturated zones according to a linear function that includes a time delay parameter for vertical routing through the unsaturated zone.

- precipitation. The input file contains the number of time steps and time step in hours on first line and then rainfall, evapo-transpiration and observed discharge at the basin outlet (one line per time step). All these data should be in units of m depth/time step/unit area (Beven, Quinn et al ., 1995). The rainfall and potential evapo-transpiration data are averaged for the whole catchment (Montesinos-Barrios and Beven, 2004).
- snow melt and accumulation. In the different variants of TOPMODEL generally a snowmelt component is not included but there is a version incorporating the ETH snowmelt component (Ambroise et al., 1996 a,b)
- reservoir. This component is not included.
- evaporation and transpiration. In TOPMODEL, hourly potential evapo-transpiration is input by the user. It can be an actual data or can be generated using a program called EVAP which can be downloaded with TOPMODEL. The method used is a combination of sine curves: an annual curve for mean daily evapo-transpiration as used in the soil moisture deficit modelling of Calder et al. (1983), and a daily curve in which the effective day length for evapo-transpiration also changes seasonally.
- interception. Considered when forest canopy appear, it is modelled through a maximum storage capacity related to the vegetation type.
- infiltration. The model includes a routine to simulate infiltration excess overland flow. This uses a Green-Ampt type solution allowing for an exponentially declining conductivity profile.
- base-flow. The model equations are based on the continuity equation, Darcy's Law, and on several simplifications: the hydraulic gradient of the saturated zone is assumed to be equal to the local topographic slope; the groundwater table is assumed to be in a quasi steady state condition; the un-saturation zone will constantly recharge water to the groundwater at a recharge rate assumed as spatially uniform over the catchment; the

distribution of down-slope transmissivity with depth is an exponential function of storage deficit or depth to the water table.

– channel routing. Sub-catchment discharges are routed to the catchment outlet using a linear routing algorithm with constant main channel velocity and internal sub-catchment routing velocity.

– surface runoff. The TOPMODEL 95.02 version includes two mechanisms to estimate the surface runoff production: the infiltration excess and the saturation excess. When the deficit in the gravity drainage store or the water table depth equals 0 the saturation condition is reached and the rainfall produces direct surface runoff (Montesinos-Barrios and Beven, 2004).

3.3.5 *Additional information: Program Features and Related Tools.*

A demonstration version of TOPMODEL for Windows developed from versions used for teaching purposes over a number of years in the Environmental Science degree course at Lancaster University can be downloaded from the web site:

http://www.es.lancs.ac.uk/hfdg/freeware/hfdg_freeware_top.htm.

A package called TOPSIMPL is a lumped version of TOPMODEL that allows calibration of a simplified four-parameter (homogeneous soil) version of the model. TOPSIMPL has the capability of a graphical display of results, including predicted dynamic contributing areas. The GRIDATB and DTM9501 programs are available separately and may be used to derive a distribution of ln(a/tanb) values from a regular raster grid of elevations for any catchment or sub-catchment using the multiple direction flow algorithm of Quinn et al. (1991).

3.4 ***ARNO***

3.4.1 *General information*

The ARNO rainfall-runoff model (Todini, 1996) was originally developed as part of a real time flood forecasting system for River Arno (Italy). Input data required by ARNO model are: topographic data (hill-slope and channel slopes), precipitation, temperature and river levels at several stations within the catchment, soil type and land use data and rating curves for hydrometric stations where the discharge is to be simulated. Some of the parameters can be determined from the above data, but most of them have to be determined by trial and error. However, this is a relatively straightforward process since the model parameters have clearly defined effects on catchment response.

3.4.2 *Flow Conditions/Representation in Space and Time*

The ARNO model is a variable contributing area semi-distributed conceptual model with two main components: the first one describes the soil moisture balance, while the second one describes the transfer of runoff to the outlet of the basin. The catchment is divided in a series of sub-basins, so that the sub-basin closing section coincides with the cross-sections of interest (because of the presence of hydrometric measuring stations or because they are interesting for flood forecasting). Data have to be sampled at the appropriate time (1h for sub-catchment sizes of 200-300 km^2 or 3h for sub-catchment of the order of 1000-2000 km^2 as an upper limit).

3.4.3 *Type of application*

The ARNO modelling system is widely used for real-time flood forecasting but is also employed in land-surface-atmosphere process research and as a tool for investigating land use changes.

3.4.4 *Representation of hydrological processes*

The main physical phenomena represented in the model are: the water balance in the soil, the water losses through evapo-transpiration, snow melt, overland flow, groundwater flow and channel flow routing. These processes have been developed into the model as inter-linked modules. The soil moisture balance is determined by dividing the catchment into elements classified as pervious or impervious, and then assuming that a distribution function can describe the proportion of pervious area that is saturated.

– precipitation. Rainfall and temperature inputs are provided to the model as area averages by means of weights generally based upon Thiessen polygons for rainfall, and taking into account of thermal gradient with elevation for temperature.
– snow melt and accumulation. The snowmelt module is driven by a radiation factor estimated from air temperature measurements. The sub-catchment area is subdivided into equi-elevation zones to take account of the role that altitude may play in combination with the thermal gradient. Snow accumulation and/or melt are based on the energy balance of the snow cover (when present) as a function of air temperature and precipitation.
– reservoirs. Not included.
– evaporation and transpiration. Evapo-transpiration is assumed to depend on: crop, wind speed, temperature, altitude, extraterrestrial radiation and the proportion of sunshine. The calculation of evapo-transpiration is based on a simplified form of the Penman-Monteith equation.
– interception. An interception component is not explicitly included.
– infiltration. Infiltration is expressed by an empirical relationship as a non-linear function of the soil moisture content.
– base-flow. It is generated by the presence of a groundwater table fed by the percolation, and computed by the groundwater module.
– channel routing. The channel routing is also performed by means of a distributed input linear parabolic model, and the contribution from upstream sub-catchments is routed downstream by means of a linear concentrated-input parabolic model.
– surface runoff. Once the runoff is obtained from the soil moisture balance, the routing of runoff on the hill-slopes of the watershed is simulated by applying a distributed inflow linear parabolic model to an “open book” representation of the hill-slope elements.

3.4.5 *Additional information: Program Features and Related Tools.*

ARNO is a commercial software package.

3.5 *SHE*

3.5.1 *General information*

The Système Hydrologique Européen (SHE) was produced jointly by the Danish Hydraulic Institute (DHI), the British Institute of Hydrology, and the French consulting company SOGREAH, with the financial support of the Commission of European Communities. It has been developed as a fully modular system for the mathematical description of the land phase of the hydrological cycle.

3.5.2 *Flow Conditions/Representation in Space and Time*

SHE (Abbot, 1986 – SHE1,2) is a physically-based, distributed model. The spatial distribution of catchment parameters, precipitation input and hydrological response is achieved in the horizontal through representation of the catchment by an orthogonal grid network and in the vertical by a column of horizontal layers at each grid square.

3.5.3 *Type of application*

Fields of application suggested by Abbott (Abbott, 1986 – SHE1) are: irrigation, land-use change, water developments, groundwater contamination, erosion/sediment transfer, flood prediction.

3.5.4 *Representation of hydrological processes*

The physical processes considered in the SHE are schematized in figure 11. Each of the major hydrological processes of water movement is considered and is modelled either by finite difference representations of partial differential equations of mass and energy conservation or by empirical equations derived from independent experimental research. The model is based on a modular scheme where each component representing different hydrological processes can be modified or omitted depending on the hydrological conditions and availability of data, and it is relatively simple to add further components. A "frame-central-control" component coordinates the parallel running of the other components by selecting their different time scales and organizing their data interchanges (Abbott, 1986 – SHE2). In the system are comprised the following models: a one-dimensional interception and evapo-transpiration model (the ET component), a two-dimensional overland flow model and one dimensional river/channel flow model (the OC component), a one-dimensional unsaturated zone flow model (the UZ component), a two-dimensional saturated flow (ground water) model (the SZ component; a three-dimensional SZ component has recently been developed), a one dimensional snow melt model (the SM component) and a two dimensional irrigation model (IR component).

– <u>precipitation</u>. Meteorological data, even if measured at a point scale, are spatially distributed.
– <u>snow melt and accumulation</u>. Depending on data availability or on general requirements two different calculation methods can be used to determine the total heat flux. The simplest is an adaptation of degree-day method. It is an empirical method that can be used when available data are limited to air temperatures. A more sophisticated method is based on the calculation of the heat flux from a budget of the energy inputs and outputs. The

snowmelt resulting from total heat flux is derived from an energy balance equation, in which also the latent heat gained by movement of water into the snow-pack is considered.

– reservoirs. Not included.

– evaporation and transpiration. The evapo-transpiration component can use three different approaches according to data availability: in two approaches actual evapo-transpiration at sub-potential rates is assumed to be partially limited by vegetation factors; in the third approach the limitation is assumed to be due only to the resistance of the unsaturated soil to water movement. However the most complex and physically realistic method used in SHE model is the Penman-Monteith equation for actual evapo-transpiration.

– interception. It is modelled by a modified Rutter model (Rutter et al., 1971-1972), an accounting procedure for canopy storage, in which the canopy is considered to have a surface storage of capacity, that is filled by rainfall and emptied by evaporation and drainage. Interception is modelled for only one vegetation in each grid square, so secondary vegetation like grass below a tree cover, is ignored. However a space-averaged vegetation type to represent mixed vegetation is possible to use.

– infiltration. Assuming that in the unsaturated zone, the zone extending from the ground surface to the phreatic surface, flow is in the vertical only, the one-dimensional Richards equation is used to determine infiltration.

– base-flow. The variation in time of phreatic surface level at each square is modelled by the non linear Boussinesq equation, that combines Darcy's law and the mass conservation of two-dimensional laminar flow in an anisotropic, heterogeneous aquifer.

– channel routing. A one dimensional solution based on the diffusion wave approximation of St. Venant equations together with the Strickler/Manning resistance law are employed

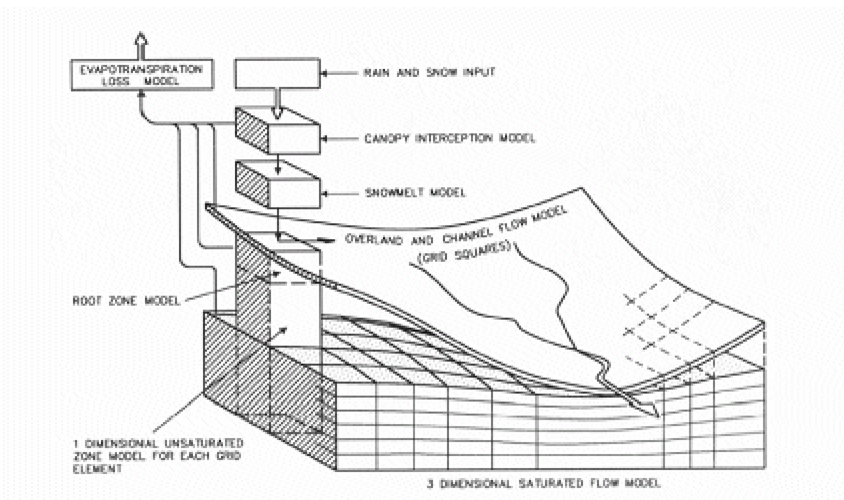

Figure 11. Schematic representation of the SHE system structure.

for modelling flow along the channel.

– surface runoff. A two dimensional solution based on the diffusion wave approximation of St. Venant equations together with the Strickler/Manning resistance law are employed for modelling overland flow.

3.5.5 *Additional information: Program Features and Related Tools.*

SHE is a commercial software package.

3.6 ***MIKE 11/MIKE SHE***

3.6.1 *General information*

The MIKE SHE model is the subsequent enhancement of SHE model. The most recent model enhancement was the development of an integrated surface water and ground water model by linking the water movement module of MIKE SHE with the channel simulation component of MIKE 11 (DHI, 1998). The MIKE SHE modelling system consists of a water movement module and several water quality modules (Yan and Zhang, 2004). The water movement module simulates the hydrological components including evapo-transpiration, soil water movement, overland flow, channel flow, and ground water flow. The related water quality modules are: 1) advection-dispersion, 2) particle tracking, 3) sorption and degradation, 4) geochemistry, 5) biodegradation, and 6) crop yield and nitrogen consumption.

3.6.2 *Flow Conditions/Representation in Space and Time*

MIKE SHE is a modular, physically based model, fully distributed in space and time. Within each module, MIKE SHE offers several different approaches ranging from simple, lumped and conceptual approaches to advanced, distributed and physically-based approaches.

3.6.3 *Type of application*

MIKE SHE can be used for the analysis, planning, and management of a wide range of water resources and environmental problems related to surface water and groundwater, such as: surface water impact from groundwater withdrawal, conjunctive use of groundwater and surface water, wetland management and restoration, river basin management and planning, environmental impact assessments, aquifer vulnerability mapping with dynamic recharge and surface water boundaries, groundwater management, floodplain studies, impact studies for changes in land use and climate, impact studies of agricultural practices including irrigation, drainage, and nutrient and pesticide management with DAISY (Hansen et al., 1990) model, developed by the Royal Danish Veterinarian and Agricultural University.

3.6.4 *Representation of hydrological processes*

The MIKE SHE modelling system simulates most major hydrological processes of water movement, including canopy and land surface interception after precipitation, snowmelt, evapo-transpiration, overland flow, channel flow, unsaturated subsurface flow, and saturated ground water flow. A grid network represents spatial distributions of the model parameters, inputs, and results with vertical layers for each grid.

- precipitation. Input can be specified as constant values or time series and can be distributed in space using stations (for instance, Thiessen polygons) or as cell-by-cell values. The time series for each station can contain different time periods with different, non-equidistant, time steps.
- snow melt and accumulation. If the air temperature is below the freezing point, precipitation will accumulate as snow until the temperature increases to the melting point. In MIKE SHE, the amount of accumulated snow that melts is calculated on the basis of the degree-day using a simple method that only requires the air temperature, a degree-day factor (the amount of snow that melts per day for every degree the temperature is above the threshold melting temperature) and a threshold melting temperature (the temperature at which melting begins, usually 0°C).
- reservoir. The MIKE SHE/MIKE 11 coupling enables the simulation of large water bodies such as lakes, reservoirs and flood plains. The Flood Area or Inundation Area option allows that a number of model grids are flooded (being part of a river, lake, reservoir etc.). The flood area may be defined as no flooding, auto(matic) or manual (DHI, 2000).
- evaporation and transpiration. Actual evapo-transpiration is calculated from potential evaporation data. Two methods are included. The first is based on the Rutter model/Penman-Monteith equation. This calculates the evaporation, the actual storage on the canopy, and the net rainfall reaching the ground surface as canopy drainage and through fall. The second method is based on the Kristensen-Jensen model (Kristensen and Jensen, 1975), where the interception storage is calculated based on the actual Leaf-Area-Index and an interception capacity coefficient.
- interception. The interception is calculated as a function of the vegetation coverage described in terms of Leaf-Area-Index.
- infiltration. Infiltration is calculated using an unsaturated zone flow model. MIKE SHE offers three different approaches including a simple 2-layer root-zone mass balance approach, a gravity flow model and a full Richards equation model. All three approaches require specification of certain soil-properties. The unsaturated zone model interacts with MIKE SHE's evapo-transpiration model which calculates actual evapo-transpiration as a function of reference ET, soil moisture and crop characteristics.
- base-flow. MIKE SHE includes a traditional 2D or 3D finite-difference groundwater model. The geology is described in terms of layers or lenses with attached hydraulic properties, that can be specified either on a cell-by-cell basis or by property zones defined by ".shp" files (polygons) or grid-code files.
- channel routing. MIKE SHE's river modelling component is the MIKE 11 modelling system for river hydraulics. MIKE 11 is a dynamic, 1-D modelling tool for the design, management and operation of river and channel systems. It consists of a hydrodynamic core module and a number of add-on modules, each simulating certain phenomena in a river system. The hydrodynamic module (HD), which is the core of MIKE 11 contains an implicit, finite difference computation of unsteady flows in rivers and estuaries. The complete non-linear equations of open channel flow (Saint-Venant) can be solved numerically between all grid points at specified time intervals for given boundary conditions. However other flow descriptions are available, such as high-order, fully dynamic, diffusive wave, kinematic wave, quasi-steady state, kinematic routing

(Muskingum, Muskingum-Cunge). Within the standard HD module advanced computational formulations enable flow over a variety of structures (broad-crested weirs, culverts, bridges, pumps, regulating, control, dam-break structures).

– surface runoff. MIKE SHE's overland-flow component includes a 2D finite difference diffusive wave approach using the same 2D mesh as the groundwater component.

3.6.5 *Additional information: Program Features and Related Tools*

MIKE SHE and MIKE 11 are proprietary software, owned and distributed by DHI. DHI's web site is at http://www.dhigroup.com/Software/WaterResources/MIKESHE.aspx where are also proposed several papers and articles in the section "Reference". AUTOCAL is a DHI generic tool for performing automatic calibration, sensitivity analysis, and scenario management with MIKE SHE, MIKE 11, MIKE 21 and MIKE 3. It is an extremely powerful tool for modellers when setting up and calibrating a model, assessing the sensitivity of model results with respect to different input variables, and for simulation of different modelling scenarios. It includes the Shuffled Complex Evolution (SCE) algorithm, for automatic calibration and parameter optimization.

MIKE SHE can be linked to ESRI's ArcView for advanced Geographic Information System (GIS) applications, and however most data preparation and model set-up can be completed using GIS software, ArcView, or MIKE SHE's built-in graphic pre-processor.

3.7 ***LISFLOOD***

3.7.1 *General information*

The LISFLOOD (De Roo, 2000; 2003) is a physically-based model specifically developed at the EC Joint Research Centre to simulate floods in large European drainage basins. LISFLOOD name derives from a soil erosion model made by De Roo for the province of Limburg and then modified for flood scenario modelling. The model was built using the PCRaster Dynamic Modelling Language a GIS capable of dynamic modelling. This model takes into account the influence of topography, precipitation amounts and intensities, antecedent soil moisture content, land use type and soil type.

3.7.2 *Flow Conditions/Representation in Space and Time*

LISFLOOD is a physically based grid-based catchment model. The user can define the spatial and temporal resolution of the model. Both should be adapted to the simulated catchment and processes. The model can run in either a "water balance mode" (using a time step of typically 1 day) or a "flood mode" (using a smaller time step; typically 1 hour) (van der Knijff, De Roo, 2004). The typical model grid-size is 1 by 1 km, although any resolution is possible.

3.7.3 *Type of application*

The LISFLOOD model has been developed to investigate the causes of the flooding and the influence of land use, soil characteristics and antecedent catchment moisture conditions in large river catchments. It is used in the European Flood Alert System (EFAS) for flood forecasting

3.7.4 *Representation of hydrological processes*

The model consists of three modules that differ mostly in time step and spatial resolution: the Water Balance (WB) model, the Flood simulation model (FS) and the Flood-Plain simulation model (FP). The WB model is run with a daily time step and usually runs for at least one hydrological year to provide the initial conditions for the flood model. The FS model is run with an hourly time step and used to simulate the actual flood event. It uses as initial conditions output from the water balance model and runs with the same spatial resolution. The FP model uses the discharge simulation of the flood model as input, but otherwise runs on a much higher spatial resolution and higher temporal resolution. It also requires information about a very high resolution DEM to simulate the inundation of a flood plain. Processes simulated are: precipitation, snowmelt, soil freezing, interception of rainfall by vegetation, infiltration into the soil, evapo-transpiration, percolation, groundwater flow, lateral flow, surface runoff and channel routing.

– <u>precipitation</u>. Precipitation data from individual stations can be used in LISFLOOD, which are then interpolated using an inverse distance method of the 5 closest stations. Precipitation is corrected for altitude effects, based on precipitation-altitude relations found in the catchment to be simulated (De Roo, Odijk et al., 2000)
– <u>snow melt and accumulation</u>. Snowfall is simulated when the average daily temperature is lower than 1.0°C. Minimum and maximum daily temperature values from stations are interpolated using an inverse distance method of the 5 closest stations, and on each pixel are corrected for altitude. Snowmelt is simulated using a degree-day method (Baumgartner et al., 1994), when the average daily temperature is above 0°C. Soil freezing is simulated using a degree-day method. If the soil is frozen to a certain degree, infiltration is reduced to zero.
– <u>reservoir</u>. Special structures such as water reservoirs and retention areas or polders can be simulated by giving their location, size and in- and outflow boundary conditions (maximum storage volume, minimum and maximum outflow, reservoir management parameters). Furthermore, specific algorithms are incorporated to simulate the buffering effect of large lakes.
– <u>evaporation and transpiration</u>. Evapo-transpiration is simulated using the Penman-Monteith method, as applied in the Wofost model (Supit et al., 1994, Van Der Goot, 1997). For forests, the Priestley-Taylor equation is used, as modified by Shuttleworth and Calder (1979). Meteorological variables used are temperature, wind speed, sunshine duration, cloud cover and actual vapour pressure, which are all interpolated from station data using an inverse distance method and where appropriate corrected for altitude. The Leaf-Area-Index of each simulated pixel is used to calculate actual evapo-transpiration from potential evapo-transpiration.
– <u>interception</u>. Interception of rainfall by the vegetation is simulated using the method of Von Hoyningen-Huene (1981) for all land use except forests, for which the approach of Shuttleworth and Calder (1979) is used. The equations are based on the Leaf-Area-Index of the vegetation. Seasonal changes of LAI are taken into account.
– <u>infiltration</u>. Infiltration is simulated using the Smith-Parlange equation (Smith and Parlange, 1978). The capillary drive value is based on topsoil texture. Saturated hydraulic conductivity values are based on topsoil texture and land use. In city areas and on water bodies no infiltration takes place.

– base-flow. Vertical transport of water in the two soil layers is simulated using a one-dimensional form of the Richard's equation. Soil water retention and conductivity curves are described by Van Genuchten's (1980) relationships. Pedotransfer-functions from the HYPRES project (Wösten et al, 1998) are used to calculate the water retention and conductivity curves from soil texture. Both soil texture and soil depth are derived from the European Soils Database (Finke et al., 1998) or local soil maps. Percolation to the groundwater store is calculated using the Darcy equation. Groundwater storage and transport to the channel system are simulated with an upper and a lower groundwater zone, and groundwater is then routed using a response function similar to the one adopted in the HBV model (Lindström et al., 1997).
– channel routing. Channel flow is simulated using a four-point finite-difference solution of the kinematic wave together with Manning's equation. The channel and floodplain dimensions (width and depth) are used to calculate the wetted perimeter. A correction of the Manning roughness value is applied to simulate the momentum exchange, which occurs across the shear layer between main channel and floodplain flows.
– surface runoff. Overland flow and transport to the channel system is simulated using a four-point finite-difference solution of the kinematic wave together with Manning's equation.

3.7.5 *Additional information: Program Features and Related Tools*

The outputs of LISFLOOD consist of hydrographs at user-defined locations in the catchment, usually the locations where also measured discharge is known. Furthermore, time series of for example evapo-transpiration, soil moisture content or snow depth can be created at selected locations, if validation data are available. The model produces a number of GIS maps, such as water source areas, discharge coefficient, total precipitation, total evapo-transpiration, total groundwater recharge and soil moisture maps.

3.8 ***TOPKAPI***

3.8.1 *General information*

The TOPKAPI (TOPographic Kinematic APproximation and Integration) hydrological model was developed by Prof. E. Todini and the hydrology research group at the University of Bologna. It is based on the idea of combining the kinematic approach and the topography of the basin. It has been developed on the basis of a critical analysis of two popular hydrologic models, the ARNO model and the TOPMODEL model, with the purpose of realizing hydrologic model with a strong physical base and a parsimonious number of physically meaningful parameters, allowing for the application of the model at increasing spatial scale without loosing the physical meaning of the parameters, and overcoming traditional limits of distributed modelling such as small catchments, long computation times, long calibration times, ect. It was proposed by Todini (1995) to exploit the potentiality of distributed models based upon physically meaningful parameters, to overcome the scale dependency of parameters as in TOPMODEL and to allow for the application of the model at increasing spatial scale, from hill slope to catchment scale (PROGEA, 2007).

3.8.2 *Flow Conditions/Representation in Space and Time*

The TOPKAPI model is a fully-distributed physically-based hydrologic model. It can be applied for both event and continuous simulations. The topography of the basin is described by a Digital Elevation Model (DEM), which subdivides the application domain by means of squared cells, whose size generally increases with the overall dimensions of the basin. Each cell of the DEM is assigned a value for each of the physical parameters represented in the model. The flow paths and slopes are evaluated starting from the DEM, according to a neighbourhood relationship based on the principle of the minimum energy cost, namely the maximum elevation difference. Due to the finite difference approach underpinning the model, TOPKAPI works with a four direction scheme (N-S and W-E): the model takes into account the links between the active cell and the four surrounding cells connected along the edges; the active cell is assumed to be connected downstream with a sole cell, while it can receive upstream contributions up to three cells. The integration of the fundamental equations is performed on the individual cells of the DEM.

3.8.3 *Type of application*

TOPKAPI, widely used for real-time flood forecasting, is suitable for land-use and climate change impact assessment, for extreme flood analysis, given the possibility of its extension to un-gauged catchments, and is a promising tool for use with General Circulation Models (GCMs).

3.8.4 *Representation of hydrological processes*

The present TOPKAPI model is structured around five modules (Figure 12) that represent the evapo-transpiration, snowmelt, soil water, surface water and channel water components respectively. For the deep aquifer flow, the response time caused by the vertical transport of water through the thick soil above this aquifer is so large that horizontal flow in the aquifer can be

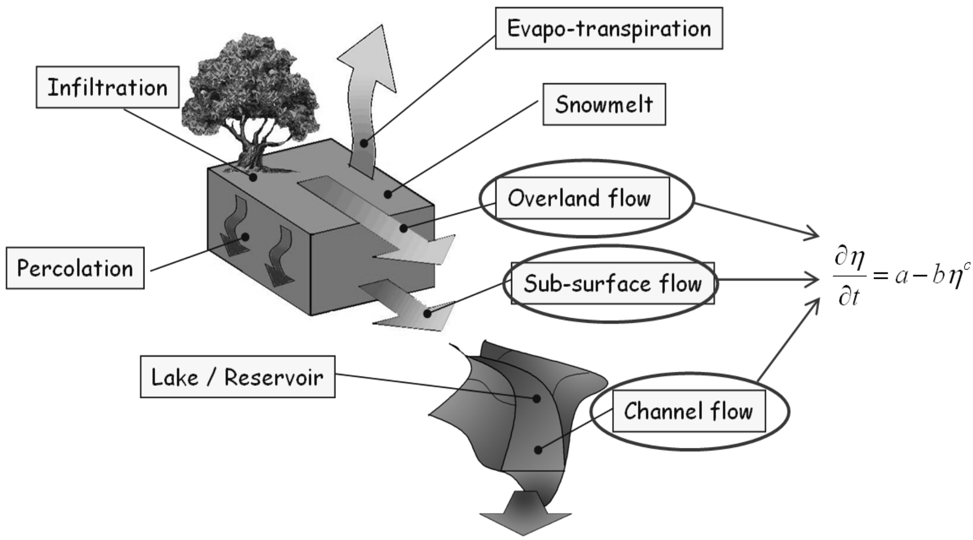

Figure 12. The basic components of the TOPKAPI model.

assumed to be almost constant with no significant response on one specific storm event in a catchment (Todini, 1995). The model accounts for water percolation towards the deeper subsoil layers even though it does not contribute to the discharge. The integration in space of the non-linear kinematic wave equations representing subsurface flow, overland flow and channel flow results in three “structurally-similar” zero-dimensional non-linear reservoir equations. The parameters of the model are shown to be scale independent and obtainable from digital elevation maps, soil maps and vegetation or land use maps in terms of slopes, soil permeability, channel and surface roughness.

– precipitation. Precipitation is assumed to be constant over the integration domain (namely the single cell), by means of suitable averaging operations on the local rainfall data, such as Thiessen polygons techniques, Block Kriging or others.
– snow melt and accumulation. For reasons of limited data availability, the snowmelt module of the TOPKAPI model is driven by a radiation estimate based upon the air temperature measurements; in practice, the inputs to the module are the precipitation, the temperature, and the same radiation approximation used in the evapo-transpiration module.
– reservoir. In the TOPKAPI model a lake or a reservoir is conceived as a “multiple-input and single-output” system. A mechanism has been developed for identifying the receiving cells and the draining cells of the lake or reservoir in a distributed model. The elevation-storage curve and the elevation-discharge curve are the necessary input data for reservoir routing. Lakes and reservoirs can be represented by defining the cells within the lake or the reservoir with the land-use type of “water body”. The water storage of a lake or a reservoir is the summation of the single water storage of each cell within the lake or the reservoir. The simple lake/reservoir model included in the TOPKAPI model for an uncontrolled lake or reservoir is based on the mass balance law.
– evaporation and transpiration. Evapo-transpiration is taken into account as water loss, subtracted from the soil’s water balance. TOPKAPI computes hourly evapo-transpiration using a radiation method (Doorembos et al., 1984) starting from air temperature and from other topographic, geographic and climatic information. In particular the empirical equation used, which is structurally similar to the radiation method formula, relates the reference potential evapo-transpiration, computed on a monthly basis using one of the available simplified expressions such as for instance the one due to Thornthwaite and Mather (1955), to the compensation factor, to the mean recorded temperature of the month and the maximum number of hours of sunshine of the month.
– interception. TOPKAPI includes rainfall interception and snow-fall interception. For snowfall interception, TOPKAPI uses equivalent water depth to represent the snow-pack on the vegetation instead of actual snow depth. A simple empirical equation can be used to determine the interception capacity of particular vegetation from the information about the vegetation’s crop density and Leaf-Area-Index (LAI) for each month of the year. Actual vegetation interception during a simulation is determined by the amount of precipitation and the deficit in the vegetation water storage.
– infiltration. All of the precipitation falling on the soil infiltrates into it, unless the soil is already saturated in a particular zone; this is equivalent to adopting as the sole mechanism for the formation of overland flow, the saturation mechanism from below (Dunne, 1978). In the TOPKAPI model only the effect of land cover on infiltration is accounted for.

- base-flow. The soil water component is affected by subsurface flow (or interflow) in a horizontal direction defined as drainage; drainage occurs in a surface soil layer, with limited thickness and with high hydraulic conductivity due to its macro porosity. The drainage mechanism plays a fundamental role in the model both as a direct contribution to the flow in the drainage network and most of all as a factor regulating the soil water balance, particularly in activating the production of overland flow. The subsurface flow at a point in the soil is approximated by means of a kinematic wave model. The point assumption is integrated up to a finite dimension in a generic cell, thus converting the original differential equation into a non-linear reservoir differential equation. The averaged interflow from the cell during the computation time interval can be calculated from respective water balance calculation in the cell. Subsequently, at each time step the saturation excess volume can be obtained by calculating the soil water balance (Liu and Todini, 2002). The TOPKAPI model accounts for water percolation towards the deeper subsoil layers even though it does not contribute to the discharge. It is assumed that percolation starts if the soil moisture content of the upper soil layer exceeds its field capacity. The percolation rate from the upper soil layer is assumed to increase as a function of the soil water content according to an experimentally determined power law (Clapp and Hornberger, 1978; Liu et al., 2005).
- channel routing. The channel network is assumed to be tree shaped with reaches having rectangular or triangular cross sections. Channel routing is performed using a kinematic approach for steep cells. Where channel slope is too small to use a kinematic approach TOPKAPI uses a parabolic routing component based on Muskingum-Cunge method.
- surface runoff. In the cells having an upstream drained area greater than the threshold value (the threshold area value is fixed by the user) the channel is present. In these cells the calculated total subsurface flow must be divided into two parts, one flowing toward the channel itself, and the remaining flowing toward the downstream cell via subsurface flow. This separation is made according to the local topography, particularly according to the average slope of the cells surrounding the active cell. Overland flow routing is described similarly to the soil component, according to the kinematic approach, in which the momentum equation is approximated by means of the Manning's formula.

3.8.5 *Additional information: Program Features and Related Tools.*

TOPKAPI is a proprietary software, owned and distributed by PROGEA (PROtezione e GEstione Ambientale) SRL.

An advantage of the TOPKAPI model is its physical basis. It doesn't need a "real" calibration in the common sense of the expression. The calibration work (which is carried out by fitting the observed stream-flow hydrograph with the simulated one) is due to the unavoidable averaging and approximation in the input data representing various phenomena. This reduces the calibration work as well as the length of data required.

TOPKAPI produces maps of all the major hydrological state variables (soil moisture conditions, runoff generation, snow accumulation, evapo-transpiration etc.) at a specified temporal resolution. In addition to instantaneous basin states, the model generates the catchment water balance. TOPKAPI include the possibility to be coupled with HEC-RAS hydraulic model (directly) or with Delft-SOBEK hydraulic model (via FEWS).

4 Experience by FORALPS partners - APAT: TOPKAPI coupled with QBOLAM

A first attempt to couple the QBOLAM (the Quadrics Bologna Limited Area Model) meteorological model, operational at APAT as a part of the Sistema Idro-Meteo-Mare (SIMM) forecasting chain, in cascade with a rainfall-runoff model , was performed by APAT in the frame of the EU NETWET II project (INTERREG IIIB CADSES programme). The physically based, fully distributed, rainfall-runoff TOPKAPI model (Todini and Ciarapica, 2001; Liu, 2002; Liu and Todini, 2002) was chosen, and the coupling was performed over the Reno river basin (central Italy). Subsequently, as an extension, the calibration of the hydro-meteorological coupled system over the Adige river basin has been performed during 2007 in the framework of the FORALPS project.

Actually these attempts have been concretized in an operative system composed by a calculation module and a display module called PROSIM. These two modules are connected to a database that can be used both for a forecasting simulation, and to verify the system behaviour in past events. PROSIM (Figure 13) is a visual interface that can be used to organize simulations and to visualize the simulation results in terms of river flow time series, soil moisture time series, meteorological data and water balance in every point of the drainage network (1D output). It can also be used to analyze and compare different simulations, to compute indexes to evaluate simulations, and to produce printable reports. The calculation module is composed by three modules:

– The HUBR (Hydrological Bayesian Uncertainty – Rainfall) module, to correct the Quantitative Precipitation Forecast (QPF) performed by the QBOLAM meteorological model by removing local bias;
– The NN module, where the Nearest Neighbour non-parametric regression technique is used to generate a stationary local model for a prediction of rainfall data sequences, choosing from a set of past situations analogous to the present one;

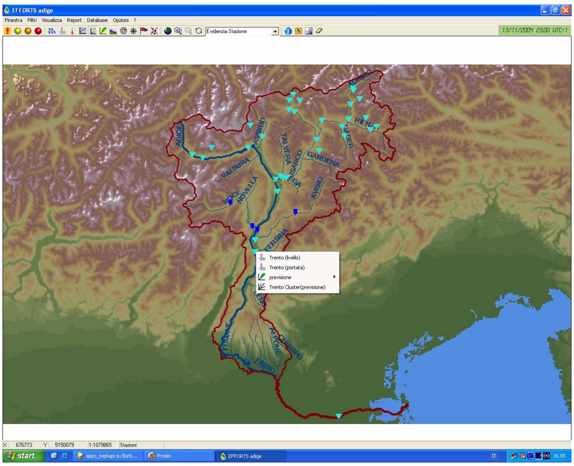

Figure 13. The display of PROSIM.

- The TOPKAPI hydrological model, to transform observed and forecasted precipitation into discharge.

4.1 *Input data*

The TOPKAPI model is a physically-based distributed model; therefore all the parameters used in the model have a physical meaning and their values can be deduced from environmental maps. For this reason before start working with the TOPKAPI model it is necessary to collect all useful information on the basin.

The following information are necessary for setting up the model:

- Contour of the basin. An ARCGis Shapefile containing the contour of the basin which is the model application area.
- River network. An ARCGis Shapefile containing the main rivers of the basin (Figure 14).
- Lakes and reservoirs. An ARCGis Shapefile containing the contour of lakes and reservoirs. If the basin has no lakes this map is not necessary. The data source of basin contour, river network and lakes contours is the APAT "Hydrologic GIS" (hydrographic and hydro-geological layers - scale 1:250.000) available on APAT Sinanet website. http://www.mais.sinanet.apat.it/cartanetms/
- Digital Elevation Model (DEM). An ESRI ASCII GRID map containing the DEM of the basin area at an appropriate spatial resolution. The initial resolution of DEM (20x20m) (source: APAT) was interpolated to a resolution of 1000×1000m. This spatial resolution was chosen in order to reduce computation time while preserving a good representation of the river network. The DEM of the basin is corrected through a depicting process to remove false outlets and sinks. The raster map of the real river and a parameter called enforcement ratio are used to carve the path of the real river on the DEM (Figure 14).
- Pedology. An ESRI ASCII GRID map (Figure 15) containing information on superficial soil type, in particular on pedology or soil texture classes. Information on superficial soil depth is also necessary. The European Soil Database (distribution version v2.0) is used (http://eusoils.jrc.it/ESDB_Archive/ESDBv2/index.htm).
- Land use: An ESRI ASCII GRID map (Figure 15) containing information on land use or vegetation. The Corine Land Cover 2000 map available on APAT Sinanet website http://www.mais.sinanet.apat.it/cartanetms/ is used.
- Rainfall and temperature data. Two files are necessary for each type of measurement: a file containing the coordinates of the instruments and a file containing point data. A raster map containing monthly mean temperature areas at the model resolution was also created. This map is necessary for snowmelt and evapo-transpiration components. The precipitation point data (available for 129 stations – Figure 16) are interpolated through the RAINMUSIC software, which performs a spatialization by weighting data by the inverse square of distance. Stations used for each cell are those falling inside a 10 km buffer. The temperature point data (available for 171 stations – Figure 16) are also distributed through the RAINMUSIC software. For each model cell, RAINMUSIC searches for measurements from weather stations within a 25 km buffer. Then temperature and precipitation point data are transformed into hourly maps with a resolution of 2000 meters.

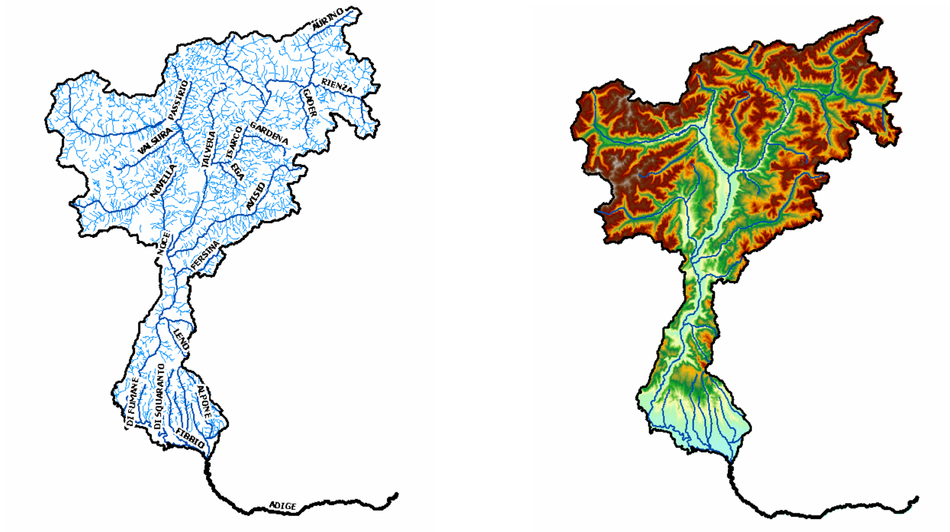

Figure 14. Input data for TOPKAPI model. Left: Adige river network. Right: Adige basin Digital Elevation Model.

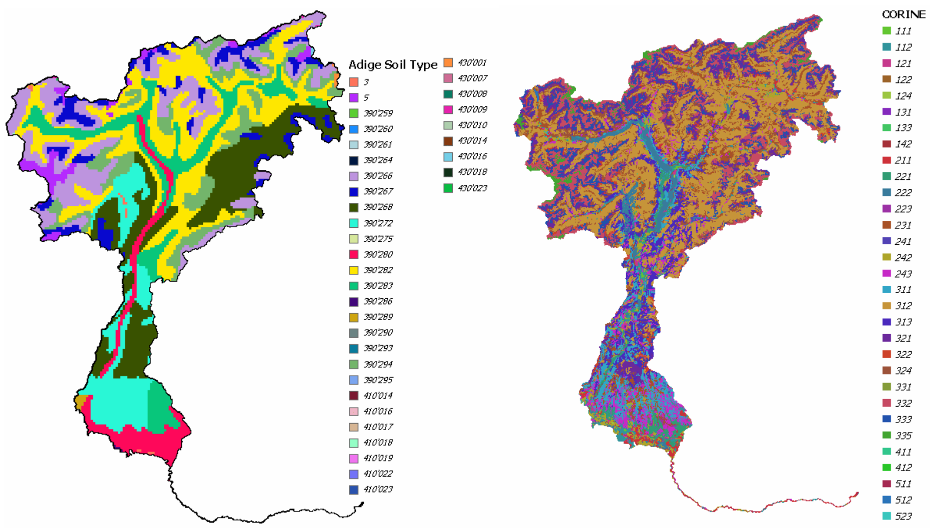

Figure 15. Input data for TOPKAPI model. Left: Adige basin pedology map. Right: Adige basin landuse map.

– Observed discharge data. Assigned at the output sections of the model, in order to compare the simulated discharges with the observed ones. Hydrometric data are available for 38 stations along the river network (Figure 16). Stage-discharge curves are used to obtain discharge data from hydrometric levels, in some river sections (the hydrometric stations used as reference for model calibration).

The hydro-meteorological data used for TOPKAPI model, which cover the time period from 1999 to 2007, are provided by the APAT Inland and Marine Waters Protection Department, the Hydrographic Office of the Province of Bolzano and the Office for Forecasts and Organization of the Province of Trento.

4.2 *Application of the TOPKAPI system to the Adige river basin*

The system is implemented on the Adige river basin situated in the regions of Trentino-Alto Adige and Veneto, in north-eastern Italy. The Adige River flows 409 km from its source, near the Resia Pass in the Tyrolean Alps, to its outlet, in the Adriatic Sea at Porto Fossone, draining an area of about 12.000 km^2. It flows along Val Venosta to Merano, touching the city of Bolzano (where the Isarco feeds it with the waters of the Rienza); the Noce flows into Adige from the right (Sole and Non valleys) and the Avisio River (Fassa, Fiemme and Cembra valleys) from the left. Then Adige flows through the city of Trento and into Lagarina valley.

The hydrometric stations chosen and used as reference (Figure 17) during the calibration are ADIGE at Ponte Adige (ETSCH bei Sigmundskron), ISARCO at Bressanone (EISACK bei Brixen), ADIGE at Bronzolo (ETSCH bei Branzoll), RIENZA at Vandoies (Rienz bei Vintl), GADERA at Mantana (Gader bei Montal). The hydrometric stations of ADIGE at Trento and ADIGE at Boara Pisani present discontinuous records, data missing for long time periods, and during flood events their hydrograph is interrupted at the peak flow. The discontinuity and scarcity of available data implies that the basin portion downstream from TRENTO was not calibrated. Model parameters in this area derived from calibration in the upstream sections. The comparison of simulated discharges with the few available observed discharges confirm, however, the effective correctness of this choice.

The calibration process results generally good in all seasons, except for winter and for spring during the snow melting period, when the flood hydrograph is strongly underestimated (Figure 18). The major events are correctly reproduced and the model is able to provide reliable results in the main hydrometric stations running in continuous along the time period 1999-2007 (Figure 19).

It has to be mentioned that the data used in calibration suffered from several anomalies. For instance, an analysis of the spatial distribution of total precipitation calculated over the time period 1999-2007 highlighted that some rain gauges happened to record zero data during a precipitation event which was confirmed by the nearest gauges. In other cases, a zero rainfall measurement during a wet period is contemporary to a thermometer failure, raising the suspect that the precipitation datum is not valid as well. Others recorded values much lower than the nearest gauges, especially at temperatures below 0°C. In some cases these anomalies can justify the underestimation of total precipitation fall on the ground and as a consequence the underestimation of simulated discharges.

Further problems for calibration came from the presence of many hydroelectric dams that were not modeled (e.g. the two dams of San Valentino on the Adige river and of Santa Giustina on the Noce river drastically affect the basin response, but they are not modelled). Nearly every measurement

Figure 16. Input data for TOPKAPI model. Upper left: the 129 rain gauges in the Adige Basin. Upper right: the 171 thermometers on Adige Basin. Bottom: hydrometric stations.

section shows the classical fluctuations in the water discharge on a daily basis as a result of the modulation system used by hydroelectric dams, especially during low flow periods.

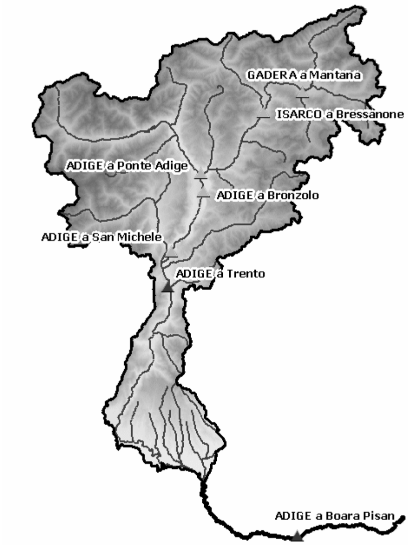

Figure 17. Reference hydrometric stations.

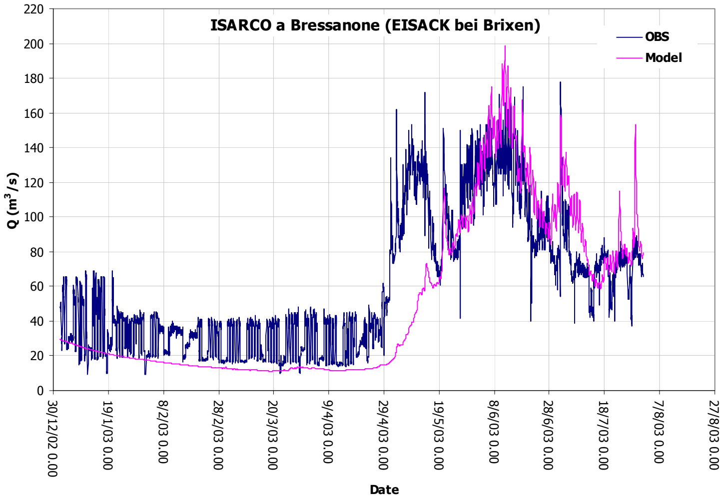

Figure 18. Results of model calibration during the spring season.

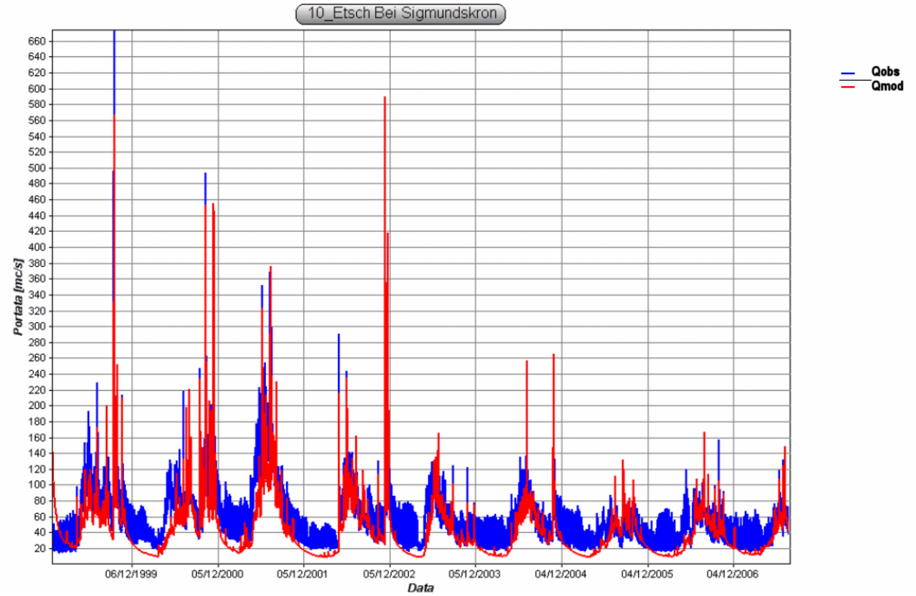

Figure 19. Discharge hydrographs from observed data (blue) and from the TOPKAPI model (red), at Adige/Ponte Adige.

At the moment the Quantitative Precipitation Forecast (QPF) supplied by the QBOLAM meteorological model covers the period between 2002 and 2006. During this time interval, which is included into the model calibration period (1999-2007), some significant flood events are identified using a 500 m^3/s peak discharge as threshold value in the ADIGE at Bronzolo station. This means assuming that only events with a peak discharge (in the ADIGE at Bronzolo station) higher than 500 m^3/s are considered hydrologically significant.

During these events, TOPKAPI, starting from an initial state provided by observed data, is run initializing the model with the QPF provided by QBOLAM at 6, 9, 12, 18 and 24 hours, and then using the same QPF corrected through the HUBR method. As an alternative, QPF at 24 hours with a 1 hour time step is determined through the application of the Nearest Neighbours method and used to initialize the hydrological model. Finally, a zero QPF has also been used as input to the model.

The following figures (20-21) show the comparison of simulations vs. observations at different forecast times for the event of 27 Oct. 2004 – 13 Nov. 2004 and the hydrometric station of ADIGE at Bronzolo. A general underestimation (both in terms of peak flow and volume) and delay of the discharge peak is apparent in forecasts beyond 12 hours. The decreasing discharge forecast quality with increasing forecasting time depends on the fact that hydrologic simulations at longer ranges are much more influenced by the quality of rainfall forecasts. If the rainfall forecast quality is poor, the hydrological forecast quality is poor as well.

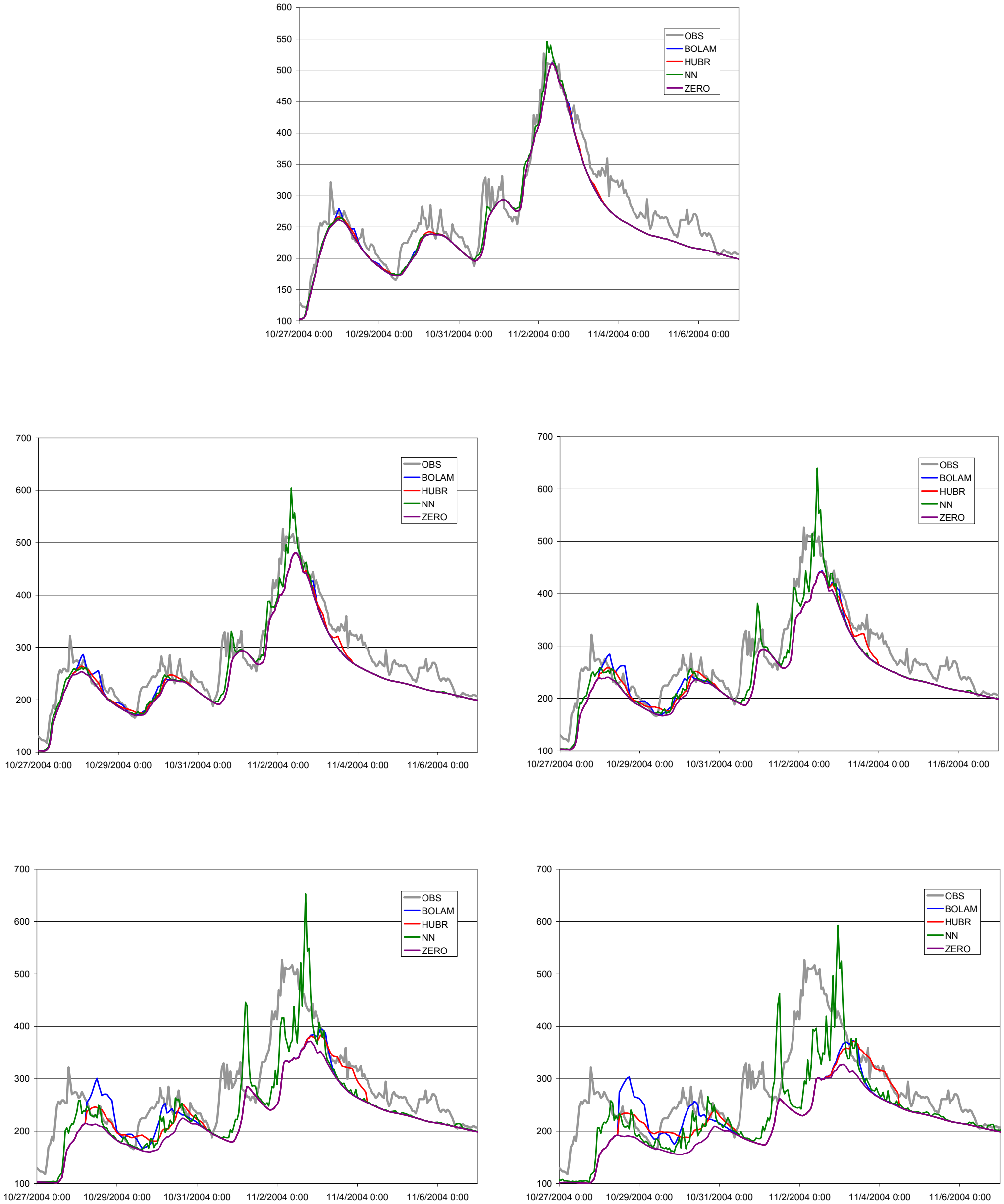

Figure 20. Discharge forecasts with different models and with a lead time of 6, 9, 12, 18, 24 hours (from top to bottom right), referred to the ADIGE river in Bronzolo (27 Oct. 2004 - 13 Nov.2004).

The meteorological forecasts used in this TOPKAPI application were produced using a QBOLAM version which suffered from strong limitations on the model code due to computer architecture constraints; a simplified convective parameterization scheme had to be implemented, which degraded the overall forecast quality. A massive re-forecasting program with a new model formulation, including the state-of-the-art Kain-Fritsch convection scheme, is currently undergoing at APAT with the aim of reconstruct the 2000-2007 forecast series. Once produced, reforecasting data should be used for a review of this hydrological modeling study.

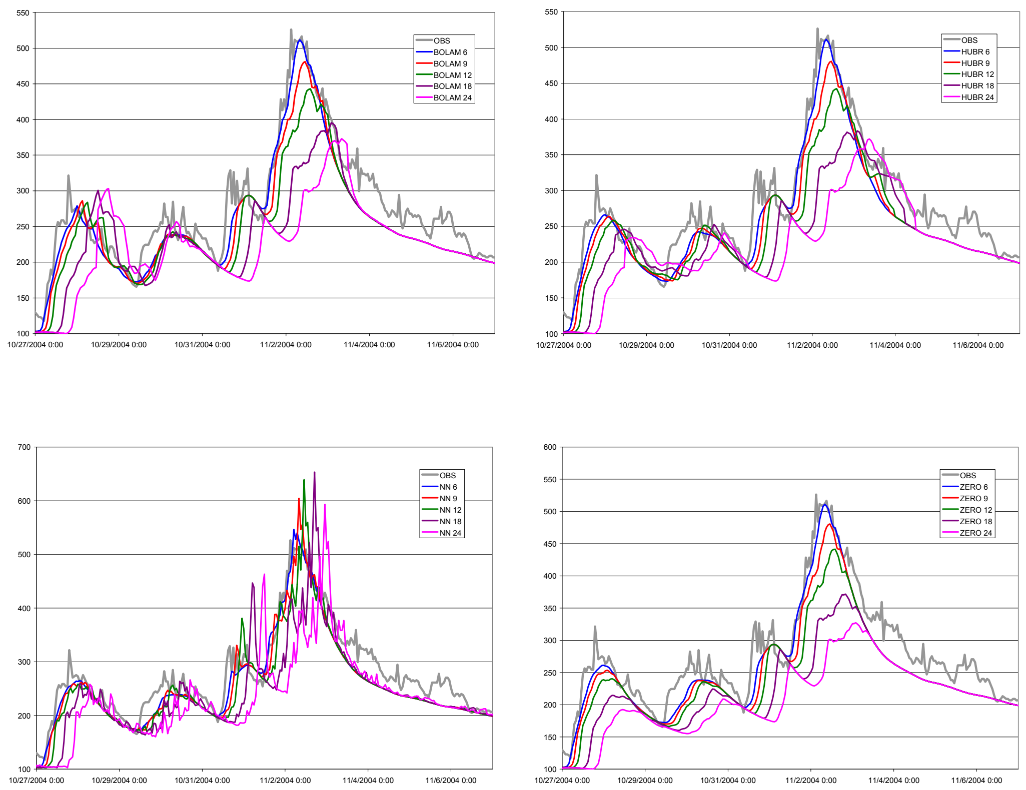

Figure 21. Discharge forecasts with different models (top left: QBOLAM; top right: HUBR-QBOLAM; bottom left: NN; bottom right: no precipitation) and with 6, 9, 12, 18, 24 hours advance, referred to the ADIGE river in Bronzolo (27 Oct. 2004 - 13 Nov.2004).

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