River Flow Simulation within Ungauged Catchments in Lebanon using a semi-distributed rainfall-runoff model

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Abstract: Development of methodologies to achieve a priori parameter estimation of hydrological models is fundamental in ungauged basins. This study shows that a semi-distributed physical model can be applied to ungauged watersheds in Lebanon and its parameters can be estimated using physical approach and remote sensing techniques. Most of the Lebanese coastal watersheds are affected by an important seasonal snow cover. The snowmelt contributes up to about two thirds of the total yearly discharge of the catchments. Most snowmelt water infiltrates the limestone and discharges at several karsts springs. A physical model based on the snowmelt mechanisms using the standard energy balance approach and a degree-day melting model allows simulating several spring discharges within the Nahr el Kelb and Nahr Ibrahim catchments. The snow cover area was calculated by combining TM5 images with a digital elevation model, and field observations made every three days, from 1400 to 2300 m altitude. Each subbasin is semi-distributed and divided into zones according to altitude. Therefore, this model was applied to Afqa spring, and the physical parameters were estimated. This model was validated on four other springs considered as ungauged basins.

Keywords: Hydrological modeling; Ungauged basins, remote sensing, snow melt

1. INTRODUCTION

Located on the eastern edge of the Mediterranean, Lebanon has a Mediterranean climate, marked by a long dry season (May to October) and a wet cold season, with precipitation varying from 800 to 2000 mm on the maritime facade. In the coastal zone of Mount-Lebanon, and at mid-to high elevations, snowfall is a substantial precipitation form which accumulates throughout the winter and melts during spring. Thus, most of the coastal watersheds are affected by an important seasonal snow cover. Snowfall and the resulting seasonal snowcover accumulated in altitude (between 1200 and 3000 meters) represent an important source of water especially in spring when the rainy season ends. This annual mechanism retains a crucial importance since it provides in a direct way, the support of the flows by spreading the period of low flow stage and by allowing a regulation of flow regimes. Moreover, snowmelt is an important contributor to streamflow and groundwater recharge. The Lebanese coastal rivers are fed mostly with more than 50% of their streamflow discharge by snowmelt runoff. Several attempts in applying traditional rainfall-runoff models on mountain catchments gave poor results evaluated by the Nash criterion, since accumulation and snow melt phenomena are not taken into consideration. Moreover, the presence of the snow-cover, which stores and releases at different time steps an important volume of water, the snowmelt, affects directly the mountain catchments hydrology (Jansson, 2003) and consequently makes the conventional rainfall-runoff models unusable under these conditions. (Rothlisberger and Lang, 1987). This directed the research efforts towards models specific to catchment affected by the snow-cover like CROCUS (Martin, Lejeune and Al 1996), VSAS2 (Barry and Prévost, 1990). These models, being of the physical type, integrate the whole catchment characteristics and proceed to a distributed approach of the snowpack thermodynamics; thus require a significant number of input data (complete radiative balance, topographic data....) incompatible with the requirements of operational hydrology (Beven, 1989). In this context, Ferguson (1999) formulated the most judicious approach by expressing that it was necessary to develop specific tools, called "snow modules" that can be integrated into the traditional rainfall runoff models, in order to extend their applicability to catchments affected by snow-cover. The majority of the models most largely used in hydrology were equipped with snow modules such TOPMODEL (Beven and Kirby; 1979). However, some models are completely dedicated to the treatment of the snow-cover such as the SRM (Rango and Martinec; 1984). The same types of modeling are found: Lumped, distributed, physical, conceptual, stochastic etc. The physical models are related directly to the study of the metamorphism of snow. The inefficiency of physical models led necessarily to a conceptual approach of modeling the snowcover. Nevertheless two approaches remain possible: purely empirical method known as the temperature index or degree-day and a method with more physical basis known as the energy budget (Anderson 1976).

The method known as energy budget was developed at the beginning of the fifties a (US Army Corps of Engineers, 1956) and was formalized in exhaustive way by Anderson (1976). The efficiency of this approach at local scale was demonstrated. The application of the method on the catchment scale appears less convincing. The use of the equations of the energy budget requires an important number of input data: vapor pressure, speed of wind, value of different radiative fluxes, etc. It is a heavy method similar to physical methods using a large number of data which are not available. More numerous attempts of comparison were realized in order to establish the superiority in term of efficiency of the energy budget on more empirical approaches without significant results. (Rango and Martinec, 1995). The most delicate problem in the application of the energy budget theory on a catchment scale is the determination of the melted quantity. This evolution can be easily followed using remote sensing techniques by delineating snow-covered areas. But the frequency of these images is insufficient to constitute a continuous entry for modeling. Thus, we developed a physical based model having a snow-covered area determined from the water equivalent volume of snowpack. The relation is supposed to be linear. The coefficient was fixed using remote sensing techniques and measured discharge of an important spring fed mostly by snowmelt. The application of this relation to the whole of the snow-covered catchments allows transposing these results to a large number of high altitude springs in Mount Lebanon. The validity of this relation was tested on several gauged springs.

2. STUDY AREA

The coastal zone of Mount-Lebanon is split in a large number of catchments which sizes vary from few tens to a few hundreds of km^2. Most of these coastal rivers are fed with karstic springs fed by rainfall and snowmelt. The hydrological regime of these rivers is characterized by high flows in winter caused by intense precipitation as well as a spreading of these flows during the spring supported by snowmelt. In this work, we will be interested in two catchments: Nahr Ibrahim and Nahr el Kelb (Fig. 1)

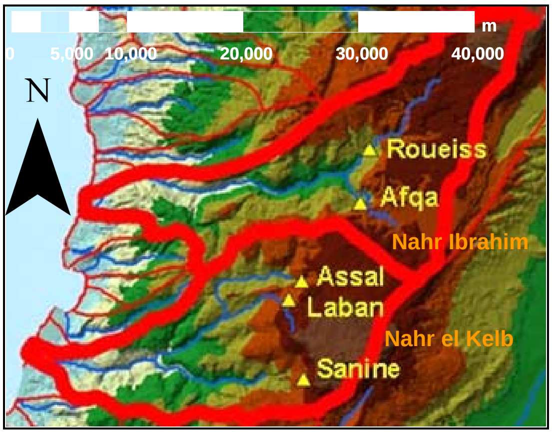

Figure 1. Study area

Most snowmelt water infiltrates the limestone plateau which groundwater discharges at two karst springs, the main springs of Nahr Ibrahim river: AFQA (1200m) and ROUEISS (1265m).

Nahr Ibrahim has a surface area of 341km^2. Nahr el Kelb, a 260 km^2 watershed, has a 2630 m elevation range and is fed by different springs. Those watersheds are strongly affected by snowcover for four months per year. The snowmelt contributes up to about two thirds of the total yearly discharge.

3. MODEL OF SNOWPACK EVOLUTION

The nature of precipitation (rain or snow) and the melting of the snow-cover are physical mechanisms governed by climatic variables (essentially precipitation and temperature). These variables are generally measured using climatic gauging stations. The great diversity of the climate implies considerable differences in the orders of magnitude of the causal variables. Transposition from a place to another is difficult. The first step is the analysis of the physical bases of melting phenomena, which are the energy mechanisms. Much research has been carried out on the energy balance snowcover (Anderson, 1976; Gray, 1987). Figure 2 represents the fluxes which cross the snow-atmosphere interface, with:

Q_n the net radiation, Q_h the sensible energy transferred by convection from the air, Q_l the latent energy provided by vapor condensation or necessary to evaporate liquid water, Q_G the fluxes which crosses the snow-atmosphere interface and which feeds the internal stock, Q_F the latent energy of snowmelt.

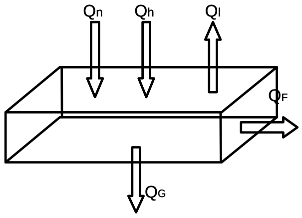

Figure 2. Energy Budget

The fluxes are counted algebraically. The energy balance shows that the sum of Qn, Q_h and Q_l is equal to the sum of Q_G and Q_F. Over a day, and under local conditions where the temperature of the ground is stable from one day to another, Q_G can be neglected in a daily balance. Each term of the balance were analyzed and estimated from the data collected at the weather station of Ouyoun Simane (1860 m) within the Nahr el Kelb catchment.

Estimation of Q_n: Q_n is the sum of net short wavelength radiation (incidental solar energy Q_{ns} less the reflected part), and long-wave radiation Q_{nl} (energy emitted by the ground less the reflected part by the atmosphere). Incidental solar energy Qs is measured by a pyranometer. Its coefficient of reflection (albedo) is estimated according to the age of snow:

Table 1. Monthly Distribution of Albedo (a)

Nov	Dec	Jan	Fev	Mar	Avr	Mai
0.8	0.7	0.6	0.5	0.4	0.4	0.4

$$Q_{ns} = Q_s.(1-a) \quad (1)$$

For the long wavelength energy, we assume that the absolute temperature of snow is equal to 273°; Thus:27216 kJ/Day. In the absence of data regarding the state of the sky, we can retain a reflection mean value of 0,85. That is to say approximately 4100 kJ/day. (Or a little less if the temperature of snow is lower than 0°C).

Estimation of turbulent transfers: The balance equations are different according to whether snow is melting (T(0) = 0) or not (T (0) < 0).

$$Q_h = K_h U(z)(T(z) - T(o)) \quad (2)$$
$$Q_l = K_l U(z)(e(z) - e(o)) \quad (3)$$

where $K_h = \rho_a C_p f$ and $K_l = 0{,}622 \rho_a Lf / P$, U(z)= speed of the wind at altitude Z, ρ_a = air density=1kg/m3 à 800hPa, C_p = Specific heat of the air =103 J/kg.°C, the heat of vaporization L is equal to 2,5.10^6 J/kg, P the atmospheric pressure and f a friction coefficient (Karman).

Table 2. Friction coefficient versus roughness

k(m)	0,001	0,01	0,1
f(10^{-3})	2,77	5,7	17,8

The saturated vapour pressure at a temperature θ can be estimated in a very exact way between -5°C and 20°C using a linear expression:

$$e_{sat}(\theta) = 6{,}11(1 + \theta/10) \text{ in hPa} \quad (4)$$

h being the relative humidity measured at 2 meters high and ranging between 0 and 1: H = e/esat

Let: $e(z) = 6{,}11(Th/10 + h)$ On the ground if there is water or melting snow h= 1

$$e(z) - e(0) = 6{,}11\left[\left(\frac{Th}{10} + h - 1\right) - \frac{T(0)}{10}\right] \quad (5)$$

At the top of the catchment, pressure is 800 hPa

$$Q_l = 10^5 fU[Th + 10[h-1] - T(0)] \text{ kJ/day} \quad (6)$$

In spite of close expressions, behaviors of the two radiant energy fluxes are very different. Measurements of the values show that velocity is the most fluctuating; it is not correlated with the two others (Fig. 3), and it is independent of time during the winter period. Under these conditions, with a rather long step (10 days), we can replace velocity by its mean value U_0 and $k = 10^5 fU_0$

$$B = Q_n + k[0{,}86T(z) + hT(z) + 10(h-1)] - 1{,}86T(0)h \quad (7)$$

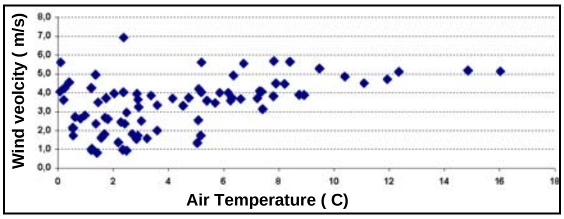

Figure 3. Mean velocity versus temperature

The balance will be written in two ways according to whether snow is frozen(T(0)<0), or melting(T(0)=0). If the snow is melting, B>0, available energy is transformed into melting heat: B = Fhρ where F is the melting heat = 335kJ/kg , ρ the density of water, and h the height in meter of the snowmelt. If the snow is melting T (0) = 0; If not F = 0. The transition between the period of frost and melt is done for F = 0 and T = 0.

$$B_o = Q_n + [0{,}86T(z) + hT(z) - 10(1-h)]k = 0 \quad (8)$$

The representation of B_0 is given in Figure 4. It shows that in 2003, the beginning of snow melt was in first April (J = 213) whereas h was 50% and T of about 7°. With h=0, B, the Bo equation = 0 give T(z)= 2°C.

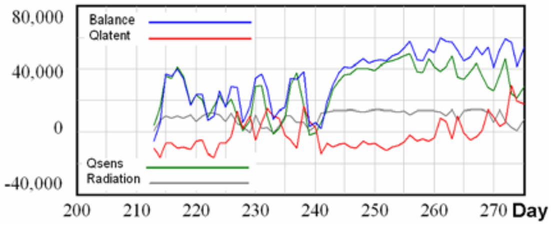

Figure 4. Balance of Energy (KJ/day)

The cumulative balance from the first day of melting is presented in Figure 5. The dominating term is the sensible heat (approximately 75%), followed by the net radiation (about 25%). Latent heat is weak, with alternate sign marking days of evaporation followed by frost. It is globally negative, which indicates that evaporation is stronger on average than the deposit of frost. Within 70 days, 23 days were frost and 47 days were evaporation. Under these conditions, the latent heat remains weak in the cumulated balance. Figure 6 shows that there is a linear relation between $B/F\rho$ and T, which justifies the application of the “degree day “under the conditions of the Lebanese mountain.

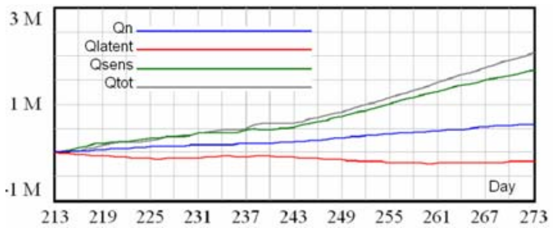

Figure 5. Cumulated energy balance (KJ)

That is to say: h = δ.T, H being the height daily melting (in water equivalent). The numerical value of δ depends on f, which is unknown. Thus, this value must be considered as a calibration parameter. This relation was firstly used in the Alpine glaciers by Finsterwalder and Schunk (1887) and was since very widely used and refined (Braithwate, 1995; Rango and Martinec, 1995). The measurements made during campaigns in 2000-2001, allow to estimate the degree day at 0.008 mm/day.degree (Fig.7). The cumulated balances, presented in Figure 4 show that the balance of latent heat, on average over a rather long period, can be estimated at 15% of the total balance. The evolution of the snowpack was followed during several years in Oyoun e Simane's station to determine the characteristic thresholds and relations (Aouad-Rizk, 2005).

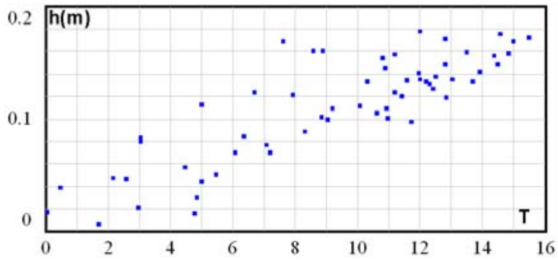

Figure 6. Mean temperature – snowmelt relation

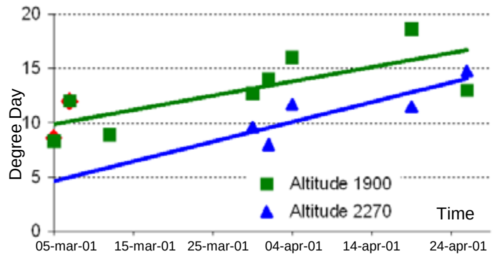

Figure 7. Measured Degree day

4. THE HYDROLOGICAL MODEL

The model is made up of three parts: management of snowcover, the production function and the transfer function. The model works in meters of water per m^2 of snowcover area.

- The snowcover management module: The intensity of precipitation is deduced from measurements of rain extrapolated in altitude. The intensity of snowmelt is determined by the equation of the degree-day. In the model, the type of precipitation (rainfall – snow) is commanded by a threshold of temperature: when temperature at 2 meters off the ground is lower than T_0, precipitation consists of snow. When this temperature is higher than T_1, snow starts to melt. T_0 and T_1 are estimated at T_0 = 3°C and T_1=0°C. Between 0°C and 3°C, the two mechanisms are simultaneous and the stock evolves differently according to the intensity of precipitation and snowmelt. Local temperature is estimated using climatic gauging stations with a constant gradient of temperature. The volume of snow is stored in a lumped reservoir. With each time step, the theory of the degree-day applied to the snow-covered surface, allows to calculate the loss of snow volume. Snow-covered surface area will evolve with the snowpack volume. When the stock of snow is important, the entire surface is snow-covered. When the stock is nil, snow-covered surface is nil. We will consider that there is a linear relation between volume and surface: V=H.S where H is a parameter which will be calibrated using remote sensing techniques.

- The function of production transforms rainfall into net rainfall by applying a constant loss. It is represented, according to the structure of the model MEDOR (Hreiche, 2002) by a reservoir having a capacity h0 and a variable rate of filling H/h_0 ($0<h/h_0<1$). The input consists of rainfall and snow melt. Two exits are provided: - A proportional exit to the entry with a coefficient related to the rate of filling. Output= Input × (Rate of filling)^2 and a loss function connected to temperature and proportional to the rate of filling. Loss = F(temp)×(Rate of filling). This structure allows simulating evaporation in winter and draining the tank in summer. Thus adjusting the annual balance using the parameter h_0.

- The transfer module: The module of production feeds the module of transfer which contains two reservoirs: one produces the fast drainage and the other the slow drainage. Net rainfall will be divided between the two reservoirs, with the proportion being controlled by a calibrated coefficient. The two reservoirs have linear exits. For the two reservoirs the streamflow discharge is the sum of the two exits.

This model is physically based and parameters can be classified in 3 types: parameters resulting from the physical analysis of phenomena: (Threshold rainfall-snow T_0, Threshold snowmelt T_1); measured parameters (Degree-day); parameter determined by remote sensing: This parameter "H" expresses the relation between snowpack and snowcover surface.

5. CALIBRATION OF THE VOLUME-SURFACE RELATION USING REMOTE SENSING

The snow covered surface evolves quickly in time, since temperatures are sufficiently high in February and March. Thus, reduction in snowcover is important, especially within altitudes lower than 2000 meters. TM5 and TM7 satellite images (European Space Agency), in low definition (1x1km) which are available above the Lebanon every eight days, are used.

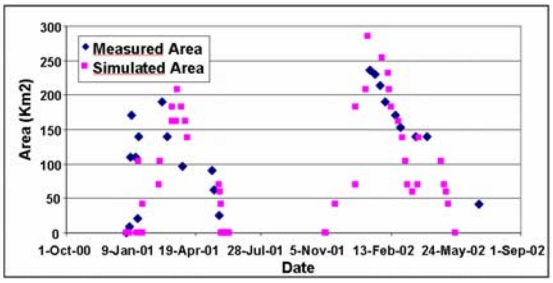

Figure 8. Adjustment of the simulated snowcover surface to the measured one (2001-2002)

Twenty two images were treated and the snow covered surface was calculated as well as the snow line using a digital elevation model. The coefficient "h" was adjusted to 0.6m, by simulations, so that measured snowcover surfaces and simulated are close (Fig.8).

6. RESULTS FOR AFQA SPRING

The comparison of the measured and simulated flows of Afqa spring shows Nash values of 0.85 and 0.81 respectively in calibration (1965-1968) and in validation (1968-1971). This model represents correctly the curves of recession as well as the peaks of flows (Fig. 9).

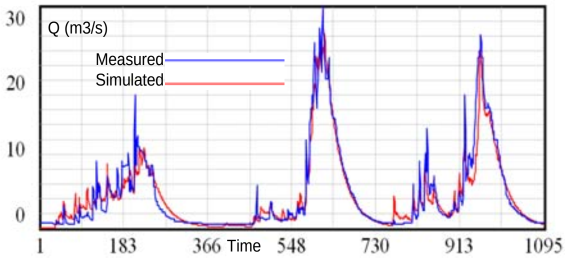

Figure 9. Measured and simulated streamflow of Afqa spring (1965-1968)

7. TRANSPOSITON OF THE RESULTS TO UNGAUGED SPRINGS FED BY SNOWMELT

This physical based model was transposed to 4 springs, considered as ungauged, for which some data exist. The model parameters calibrated on Afqa spring remain unchanged. Rainfall and temperature were distributed on each catchment. The model show successful results on the 4 springs (Nash values between 0.75-0.85).

Table 3. Springs used for validation

Spring	Area(km^2)	Altitude(m)	Data	Nash
Assal	20	1835	1968-1971	0.82
Laban	36	2012	1968-1971	0.78
Sanine	8	2090	1970-1971	0.75
Roueiss	90	1265	1965-1971	0.77

8. CONCLUSION

Conceptual rainfall-runoff modeling is well adapted to small catchments (100 to 300 km^2) since distributed modeling is very difficult within complex basins, like in the littoral catchments of Lebanon. But problems appear when these basins are subject to a significant snowcover, producing a significant part of streamflow. The imput of the model cannot be reduced to rainfall. It is necessary to add the temperature, whose role is complex.

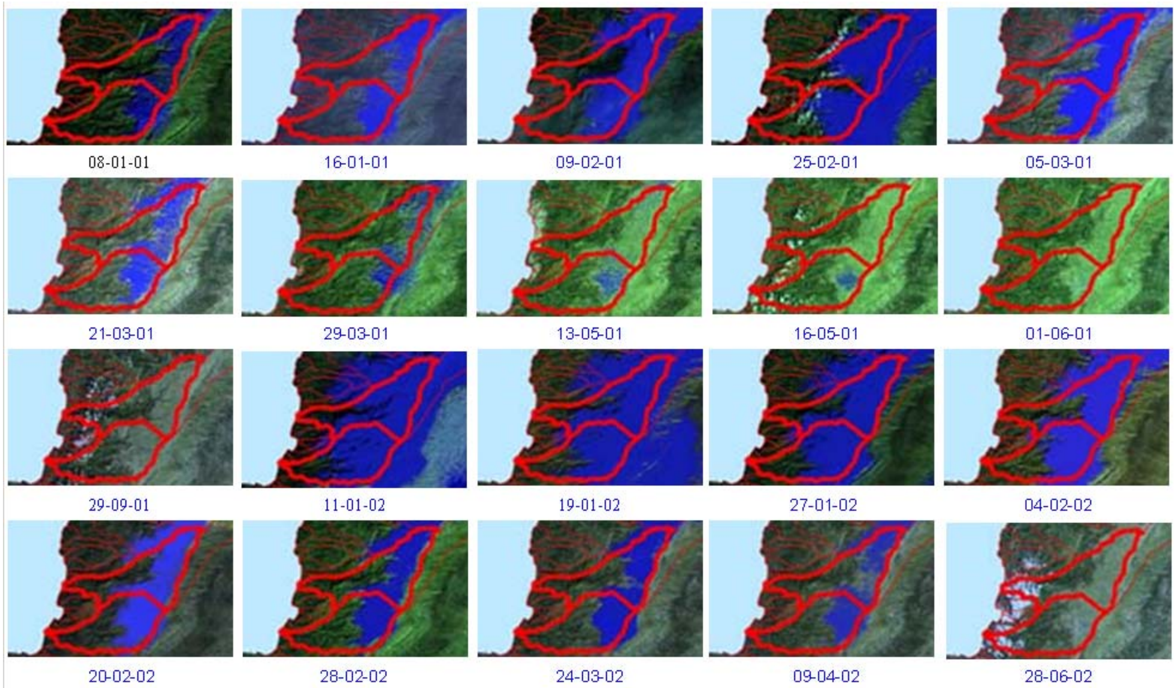

Figure 10. Evolution of the snowcover and snowline (2001-2002)

The temperature has a double effect: it ensures the transformation of rainfall into snow, if the temperature is rather low, and of snow into rain, if it is rather high. Because of the variations of these characteristics with altitude, a lumped modeling appears unsuitable. The snow line, with an altimetric cutting, defines a variable surface versus time. Thus, it would be necessary to distribute the snow covered area. The determination of a relation between the snowpack and the snowcover using remote sensing techniques allowed the development of a lumped conceptual model, which was applied to a number of altitude spring in Mount Lebanon.

9. ACKNOWLEDGEMENTS

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