OVERVIEW

The goal of this handout is to summarize the key hydrological concepts related to water fluxes in a catchment. The handout is divided into the following sections:

1. Quantification of hydrologic fluxes: precipitation, evaporation, infiltration, and rainfall excess. This is followed by a short summary of routing methods.
2. Methods for estimating decision relevant metrics: streamflow observations, rainfall runoff models, and methods for prediction in ungaged basins.

ADDITIONAL READING

These notes present selected and condensed versions of topics in surface hydrology and readers are referred to Applied Hydrology by Chow et al. (Chow et al. 1962), for a full description. Similarly, streamflow simulation by rainfall runoff models and regionalization methods presented here are not exhaustive and additional readings are listed in the text.

1. Quantification of Major Hydrologic Fluxes

A. Atmospheric components of the hydrological cycle: an overview

The hydrological cycle is the movement of water in its three phases across the globe. The cycle is driven by the energy from the Sun that converts water from liquid or solid phase to vapor form. This process is termed as ***evaporation*** (liquid to vapor) or ***sublimation*** (solid to vapor). Water vapor is also released into the atmosphere by ***respiration*** of all living organisms. However, this flux is quite small as compared to that of ***transpiration*** which is the release of water vapor from plant stomata and occurs alongside photosynthesis by plants. As it is very difficult to distinguish the fluxes of water vapor contributed by transpiration of plants and direct evaporation, many times the term ***evapotranspiration*** is used to refer to the conversion of liquid water to water vapor in a catchment. The process of evaporation puts water into the atmosphere, from where under suitable conditions, it condenses out as ***precipitation*** and falls back to the land. Note that due to local, regional, and global circulation of atmosphere, water vapor is transported across the globe and does not necessarily precipitate in the region of origin. The amount of precipitation that does come from the water vapor generated within the same region is termed as ***recycling ratio***. Large forests such as Amazon have a very high recycling ratio, indicating that most of the precipitation comes from evapotranspiration within the same region.

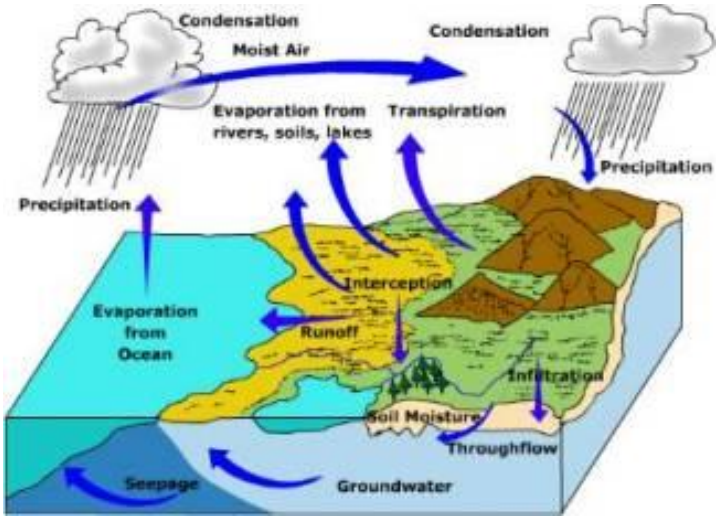

Figure 1. The global hydrological cycle. Image from: http://www.miwaterstewardship.org/Portals/0/images/hydrologic_cycle.323px.jpg

B. Quantification of Precipitation

Precipitation in the form of rain is generally measured by ***rain gages*** that collect the water falling into a marked circular cylinder. Readings can be taken manually at fixed time intervals (generally 24 hours), or more frequently in case of heavy rainfall. Alternatively, automatic recorders are available that provide a discrete to continuous measurement of rainfall as a function of time. Thus, units of measured rainfall are length per unit time, generally mm/day or inches/day. Other forms of precipitation such as snow can be measured using similar devices. Snow measurements are taken by allowing the snow to melt and are reported in length per unit time snow water equivalent (SWE). The graph of precipitation against time is termed the ***rainfall hyetograph***. The graph of cumulative precipitation against time is termed as the ***rainfall mass curve***. As rainfall is quite sporadic, rainfall hyetographs are ragged and generally composed of storms (a series of rainfall events) interspersed with zero rainfall events.

Generally, one is not interested in readings at a single rain gage site but in the average rainfall over an entire catchment. To obtain catchment average rainfall, several methods are available that employ rain gage observations from in and around the catchment. The simplest method is to take the average of all rain gage readings within the catchment. However, this is not recommended if gages have very different readings. The ***Theisson polygon*** method weighs the observation at each gage by the catchment area represented by that gage. It is based on the principle that rainfall at each point within the catchment is represented by the rain gage closest to it. The method also allows for inclusion of rain gages close to catchment boundary but outside it (See Figure 2). The ***Isohyetal method*** constructs isohyets or contours of equal rainfall in a catchment using point rain gage data. Then the catchment average rainfall is obtained by multiplying the area between each isohyet with average rainfall on the isohyets.

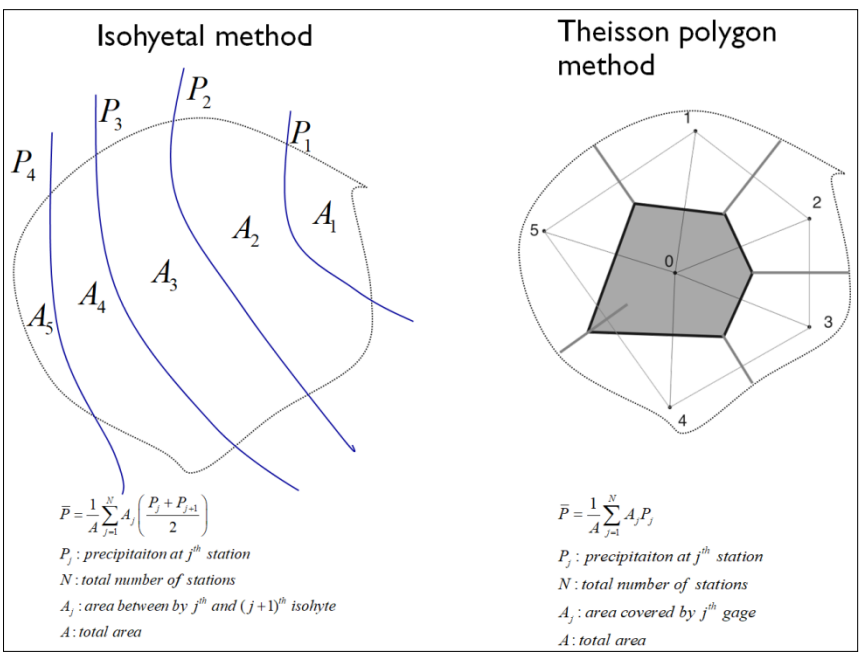

Figure 2. Pictorial representation of two ways to obtain catchment average precipitation from point rain gage data.

C. Quantification of Evaporation/Evapotranspiration

Evaporation is expressed as depth of water evaporated per unit time. Evaporation from an open water body is governed by available energy and vapor transport. Evaporation in such cases is generally quantified either by using the energy balance or the aerodynamic method. The ***energy balance method*** assumes that energy is the limiting factor in evaporation while the ***aerodynamic method*** assumes that vapor transport is limiting. The energy balance method estimates evaporation rate expressed as depth of water evaporated per unit time, E_r,as:

$$E_r = \frac{R_n}{l_v \rho_w} \tag{1}$$

Where, R_n is the net radiation flux at the surface (after accounting for albedo, sensible heat flux, and ground heat flux), $l_v$is thelatent heat of vaporization and ρ_w is density of water. The latent heat of vaporization is the energy needed to convert 1 kg of water from liquid phase to vapor phase at a given temperature. Note that latent heat of vaporization is itself a function of temperature. The units of radiation flux, latent heat of vaporization, and density of water are J/s/m^2, J/kg, and kg/m^3, respectively. The aerodynamic method estimates E_a as:

$$E_a = B\left(e_s - e_a\right) \tag{2}$$

Where, e_s(Pa) is the saturation vapor pressure corresponding to ambient air temperature, e_a(Pa) is the ambient vapor pressure, and B (m/Pa.s) is the vapor transfer coefficient that varies from one place to another and embeds in itself wind velocity, roughness heights, ambient air pressure, air density, etc. Many times, both energy and aerodynamic estimates of evaporation are combined together using the combination equation:

$$E = \frac{\Delta}{\Delta + \gamma} E_r + \frac{\gamma}{\Delta + \gamma} E_a \tag{3}$$

Where Δ is the gradient of the saturated vapor pressure curve at ambient air temperature and γ is the psychrometric constant. The Priestly-Taylor equation assumes that contribution of the aerodynamic term in the combined method is 30% of the energy term, simplifying the estimation of evaporation.

It is worth pointing out here that the equations above estimate evaporation from an open water body, while evaporation over land is more complicated. As mentioned earlier, evapotranspiration over land is a combination of evaporation of water from land surface and transpiration from living organisms. The maximum possible evapotranspiration when water is not limited in the system is often termed as potential evapotranspiration. Potential evapotranspiration for a land surface depends upon the degree of saturation of soils, the nature of plants grown over the land surface, and the growth stage of plants. In order to standardize the value of potential evapotranspiration, it is often reported for a reference grass crop allowed to cover the entire land surface and grown with no

water shortage. This value generally depends upon climatic conditions such as radiation, wind speed, relative humidity, and air temperature. This value is multiplied by a crop factor to account for different crop types, and then by another correction factor to account for non-standard growing conditions. Food and Agriculture Organization maintains a great reference for estimation of potential evapotranspiration and accounting for crop type and farm conditions. Please look for more information at: http://www.fao.org/docrep/x0490e/x0490e04.htm

D. The Hydrological Cycle: Land Surface Components

On its way to the land surface,precipitation in the form of rain, snow, hail, sleet, etc., may be ***intercepted*** by leaves of trees. This intercepted water can evaporate back to the atmosphere or fall towards the ground as ***throughfall***. Alternatively, water may move along the trunks of trees or stems of smaller plants and reach the ground as ***stemflow***. Once it reaches the land surface, it may be retained in small depressions on the surface as ***detention storage***. After this, the water will start to ***infiltrate*** the soil strata and make its way through the pores in the soil matrix. These pores, when partially filled with water constitute the ***vadose zone*** where ***unsaturated flow*** occurs, and when fully filled with water will constitute the ***groundwater*** where ***saturated flow*** occurs. Once water enters the soil, it becomes ***soil moisture***.

E. Infiltration and its quantification

In unsaturated soils, three forces are present: gravity, friction, and suction. The suction force is a result of electrostatic forces between water molecules' polar bonds and soil particles. This is similar to the forces that lead to raising of water in a small glass tube. Suction forces cause the soil medium to hold on tightly to water molecules. However, as more and more water molecules fill up the voids between soil particles, these forces become weaker and eventually vanish for saturated soil media. Thus, in saturated soil only gravity and friction forces are present. Laws of mass and momentum conservation can be applied either for representative elementary volumes (a small enough volume so that soil can still be considered as a matrix of soil particles interspersed with voids), or at the scale of soil columns. Darcy's law is the momentum equation for subsurface flow:

$$q = -K \frac{\partial h}{\partial z} \tag{4}$$

Where, q (mm/day) is the Darcy flux or the rate at which water moves through the soil medium, K is the hydraulic conductivity (a function of soil type and soil moisture content, θ), h (mm) is the total head of flow, and z (mm) is the distance in vertical direction. The negative sign indicates that water moves in direction of decreasing head. The head, h, includes suction forces in case of unsaturated flows while only depends upon the depth in case of saturated flows.

When Darcy's law is combined with mass conservation law for representative elementary volumes in unsaturated soils, the ***Richard's equation***for unsteady unsaturated flow though porous medium is obtained. The equation relates the partial derivatives of soil moisture content with respect to time and depth as:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z}\left(D\frac{\partial \theta}{\partial z} + K \right), D = K\frac{\partial \psi}{\partial \theta} \tag{5}$$

Where, θ (unit less) is the soil moisture content, Ψ(mm) is the suction head, D (mm^2/day) is the soil water diffusivity, K (mm/day) is the hydraulic conductivity, z (mm) is the depth, and t (day) is time. Given boundary conditions and soil properties, this equation can be solved to obtain the variation of soil moisture content, θ, as a function of space and time. Basically, the Richard's equation can be solved to understand how water infiltrates through soil. However, solving these partial differential equations requires the knowledge of soil properties throughout the domain as well as proper determination of boundary conditions. As it is very hard to exactly measure soil properties due to heterogeneity (variation of soil properties in a given direction in space) and anisotropy (variation of soil properties in different directions at a given point in space), alternative conceptualizations of infiltration are used. These conceptualizations are at a scale much larger than that employed to derive Richard's equation, typically for entire soil columns. An example is the Horton's equation for infiltration:

$$f(t) = f_c + \left(f_o - f_c \right) e^{-kt} \tag{6}$$

Where, f(t) (mm/day) is the rate of infiltration of water into the soil matrix, f_o(mm/day) is the rate of infiltration at the beginning of infiltration, and f_c (mm/day) is the rate after a long time of infiltration. This model assumes that initially, there is a fast rate of infiltration of water into the soil (f_o), which reduces with time but becomes constant at f_c. The transition from the initial high rate to the final lower constant rate is governed by the decay constant, k (day^{-1}). f_o, f_c, and k are parameters of the infiltration equation that should be determined by field experiments. Alternative conceptualizations exist such as the Philip's equation or the Green-Ampt method. Readers are referred to texts on hydrology for details.

F. Rainfall Excess Generation

As water infiltrates a soil surface, there may come a time when infiltration rate is lower than the incident rainfall rate. In such a situation, the excess water will start collecting over the land surface. This water is termed as ***rainfall excess*** and gives rise to ***overland flow***. As catchments are heterogeneous systems, a rainfall event will generate varying amounts of rainfall excess across a catchment. When infiltration rates are used to determine generation of rainfall excess, the overland flow is termed as ***Hortonian overland flow***. For example, if the infiltration rate is f (inches/hour) and rainfall intensity is i (inches/hour), the rate of rainfall excess is (i-f), provided i>f. Of course, when i<f, all water infiltrates into the soil matrix and no overland flow is generated. Hortonian overland flow will be generally applicable to impervious surfaces in urban areas and in areas where soils have very low infiltration capacity.

In forested landscapes with soils of high infiltration capacity, Hortonian overland flow is quite rare. Instead, water infiltrates through the soil matrix and saturates the soil column (given rainfall is

sufficient). In such catchments, low lying areas of catchments nearer to the stream will get saturated earlier. Any rain falling over saturated soils will be immediately converted to ***saturation overland flow***. As the saturated area will vary with time, the areas of land contributing to overland flow are also termed as ***variable source areas***. Typically, as a rainfall increases across a catchment, the variable source areas also increase. After the rainfall ceases, the variable source areas also shrink. Thus, only a small surface area of a catchment may contribute to streamflow in a storm.
Water can reach the stream using many paths. Rainfall excess is the 'extra' water that is collected overland, and flows under gravity to reach the stream. As this water joins the stream, it flows in the stream to reach the gage location, where it is measured. It is worth pointing out here that runoff generation is a complex process as water infiltrating into the soil matrix may join the ground water and flow towards the stream constituting ***baseflow***. Baseflow contributions from water stored in a catchment may continue well after rainfall has ceased. The degree of storage in a catchment therefore has a great impact on ability of streamflow to maintain itself in non-rainfall periods. An important number that represents catchment's hydrology is the ***runoff coefficient***, which is the ratio of total runoff in a time period (generally a year or longer) to total precipitation in the same time period. At annual to larger time scales, the runoff coefficient represents how incoming precipitation is partitioned into streamflow and evapotranspiration in a catchment.

G. Rainfall Excess Quantification

Rainfall excess eventually becomes streamflow out of a catchment. Thus, its quantification is critical to get an accurate understanding of how rainfall is partitioned in a catchment. There are various ways to quantify rainfall excess, beginning with simple methods based on use of infiltration equations, to complex hydrologic models. Each method is a conceptualization, though distributed hydrologic models employing Richard's equation (along with supporting equations of mass and momentum conservation for saturated media) are often claimed to be more physically based. It is worth remembering here that the heterogeneity of even a small catchment makes it hard to solve these equations accurately. Not to mention, the inability to capture the spatial distribution of rainfall in a catchment makes the results even more approximate. Most catchments will have only a few (or just one) rain gage, and catchment average rainfall will be used for estimating rainfall excess. Thus, the onus of choosing a suitable method is on the hydrologist. It is better to use multiple methods and cross-verify the numbers. Here, a common method to estimate rainfall excess for a storm is described to give a flavor of various factors that control runoff excess.
The ***Soil Conservation Service's (SCS) Curve Number method*** is a conceptual method, which has been combined with field observations to allow application to various conditions. The SCS curve number method assumes that for any storm, the ratio of the depth of excess rainfall (P_e) to total storm precipitation (P) is equal to the ratio of water retained in the basin (Fa) to maximum storage within the catchment (S). In the form of equations:

$$\frac{F_a}{S} = \frac{P_e}{P - I_a} \tag{6}$$

Where I_a is the initial abstract or that amount of rainfall for which no rainfall excess occurs. One can conceptualize the initial abstraction as the water retained in depressions in the ground or leaves. This water remains as a storage in the catchment and no rainfall excess will be generated until these storages are filled up. Note also that this Eq. 6 is applicable to the entire catchment at once. Equation 6 is coupled with continuity equation, again applied for the entire catchment at once:

$$P = P_e + I_a + F_a \tag{7}$$

Equation 7 says that the sum of rainfall excess, water infiltrated into the soil, and any initital abstractions must be equal to storm rainfall. Combining Eqs. 6 and 7 lead to:

$$\begin{aligned} P_e &= \frac{\left(P - I_a\right)^2}{P - I_a + S}, \; if \;\; I_a = 0.2S \\ P_e &= \frac{\left(P - 0.2S\right)^2}{P + 0.8S} \end{aligned} \tag{8}$$

Equation 8 provides an expression to estimate rainfall excess, given knowledge of storm precipitation, and maximum possible storage in the catchment, S. S is a property of catchment and varies from one to another. For example, catchments with highly porous soils covered by forests may store more water than those covered by impervious concrete surfaces. Thus, experiments were carried out across different land use and soil types to relate S to catchment properties. To simplify the interpretation of results, S was recast as curve number, CN, such that:

$$S = \frac{1000}{CN} - 10 \tag{9}$$

In simple terms, curve number, CN, is the slope (in percentage) of the relationship between cumulative rainfall excess and cumulative rainfall falling over a plot. A CN value of 100 implies all rainfall is converted to runoff, while CN<100 indicates that some rainfall is retained in the catchments and the remaining becomes rainfall excess. CN was derived experimentally for different land use and soil types. In addition, corrections to CN were suggested if ***antecedent moisture conditions (AMC)*** in the catchments differed from normal. If the catchment is much wetter than normal before a rainfall event, it will generate more excess rainfall than expected and vice-versa. Note that the curve number method as presented here is applicable only to storms, and cannot be applied to estimate rainfall excess at time scales greater than storm time scales, such as monthly or annual. This is because, no where in this method the estimation of actual evapotranspiration out of the catchment is included. For time scales such as annual, the total precipitation is divided into streamflow, actual evapotranspiration, and ground water fluxes out of a catchment.

H. Flow routing

Once rainfall excess is generated at a location in a catchment, it takes some time to flow over the land and reach the stream. From the point where this rainfall excess joins the stream, it travels in the stream to the location where flow is measured. Remember that rainfall excess is generated across a catchment in a very heterogeneous manner, thus water *parcels* may join the stream at various locations along it, and travel different distances to reach the observation location. This introduces a time lag between generation and observation of surface runoff. Among other factors, this lag will be governed by the size of the catchment. Water parcels generated across a very small catchment will take much less time to reach catchment outlet, while those generated in large catchments may take days. Average annual water availability in a basin may not be affected by this lag, as the total surface water is still the same. But this lag is very important to get the timing and magnitude of floods right. Thus, additional methods are needed to conceptualize and quantify this process. These suite of methods fall under the category of routing methods.

Unit hydrograph are the earliest of these methods. It is defined as the direct runoff hydrograph resulting from 1 inch of rainfall excess generated uniformly over a catchment at a constant rate for an effective duration. The effective duration refers to the time step of the excess rainfall event, whether excess rainfall is simulated every 30 minutes or every hour or other time steps. Generally, unit hydrograph is derived from available rainfall and runoff data, and then used to predict future response of the catchment under different rainfall conditions. There are other methods that aim to derive unit hydrograph for a catchment in absence of streamflow data using catchment properties. While unit hydrographs are useful to capture the movement of water parcels from within the catchment to measurement location at the catchment outlet, other routing methods are used to simulate how water flows from one point in a stream to another downstream. The Muskingum method is one such method. In case there are reservoirs in the path of a river, level pool routing method is used to simulate flow downstream of a reservoir. Unit hydrograph, Muskingum, and level pool routing all fall under the broad category of hydrologic routing as they do not simulate streamflow as a function of space, but only as a function of time. Hydraulic routing methods such as Kinematic wave, diffusion wave, or the dynamic wave routing methods, are different forms of the Saint-Venant's equations that allow simulation of streamflow as a function of both space and time.

2. <u>Methods for estimating decision relevant metrics</u>

For decision analysis related to water resources, a variety of information is needed depending upon the context. For example, when planning for water supply of a city that depends primarily on a surface water source, information regarding mean annual streamflow as well as seasonality of streamflow may be required. In addition, water managers may also need to prepare for hydrologic extremes such as floods and droughts. They will be interested to learn about various metrics such as duration of droughts, peak floods, frequency of exceeding certain thresholds, etc. All these metrics have to be abstracted from a time series of streamflow. Thus, depending upon the context, daily to monthly streamflow time series would be needed for planning purposes. We group the methods to obtain this information into two categories:

A. When streamflow data is available

It is possible that the region for which one is planning already has good quality measurements of streamflow for the last few years or decades. In the past, when climate change impacts were not a concern, planners would use these past observations to estimate relevant metrics. Under the assumption that the physio-climatic characteristics of a catchment do not change significantly in the future, the same metrics can be used for planning. Thus, various tools of hydrologic statistics would be used to extract relevant information from the time series, such as return periods, exceedance probabilities, etc. The only source of uncertainty would then come from our measurements themselves, for example, from ***gage-discharge relationships*** that are used to convert river stage to discharge in practice. Additional uncertainties can arise from natural variability of streamflow, such that any metric should be reported with appropriate ***confidence bounds***. These required careful analysis of the measurement data and advanced statistical tools, but was quite achievable.

This approach worked quite well and standard practices were (are) in place for most planning problems. However, in the past few years, extensive land use change and long term climate change have become key drivers of hydrologic response. The assumption that the past will represent the future (in terms of statistical summaries) does not work anymore (see Milly et al. 2008). This forces us to now understand catchment processes and estimate how the future may differ from the past. Wagener et al. 2010 recommend how past practices may be altered to deal with these new challenges.

B. When streamflow data is not available

Most hydrology today falls under this category. This is because even if data is available, impending environmental changes imply that we need to project how the catchment's response might change under changing drivers. In addition, large parts of the world do not have good quality streamflow measurements. Thus, data is not available when there are no measurements and when we need to project into a future that is quite different from the past. The future is data scarce as we have not witnessed it yet. In these circumstances, two approaches are common.

B1. Rainfall runoff modelling

Rainfall runoff models translate inputs such as rainfall, temperature, land use type, etc. to streamflow response for a catchment. Thus, if inputs change in the future, models should be able to project the streamflow simulations under those changed conditions. A model is an abstraction of reality, we identify dominant processes in a catchment and represent them in mathematical forms in a model. Any model is basically a set of equations that comprise ***variables*** (that change with time) and ***parameters*** (that do not change with time). Parameters represent the time invariant properties of the catchment. They are a complex function of the geologic, topographic, vegetative, and perhaps even climatic characteristics of a catchment. Parameters govern how a catchment will respond to an input forcing. In a broader sense, both model structure (the set of equations) and parameters (constants in those equations) together represent a catchment. But, many times, the same model structure may be applicable to a variety of catchments with variation of catchment response

represented by changing parameters. More often than not, the parameters are estimated by calibration. Calibration is the process of exploring the feasible parameter space to identify parameter combinations that best explain the past observed streamflow. However, it is also possible that in certain catchments the model structure itself fails, as it does not contain certain important catchment process(es). In summary, rainfall runoff models can be used for simulating streamflow response of a catchment but the modeler should be mindful of the underlying assumptions of this process.

Rainfall runoff models available to a planner range from purely data driven, to conceptual, to physics based. ***Data driven models*** assume that the hydrologist has no knowledge of the system and the model structure itself along with any parameters should be derived from the data. Common among data driven approaches are artificial neural networks, model trees and other machine learning (and now deep learning) techniques. ***Physics based models*** build hydrologic response from first principles. They apply major governing equations for unsaturated (recall the Richard's equation) and saturated flows across the catchment and solve these partial differential equations using a numerical scheme. Ideally, a physics based model would not need calibration as by definition all parameters should be derivable from observable catchment properties. Despite the search for a calibration free physics based model being the holy-grail of rainfall runoff modelers, catchment's heterogeneity and the inability to measure subsurface properties have been major hindrances in obtaining calibration free physics based models. In most cases, a few parameters of these physics based models are tuned to ensure that past behavior is adequately represented. ***Conceptual models*** fall in the middle of these two opposite paradigms. Their basis is the ***systems approach*** to model the catchment. The systems approach says that each catchment is a system and instead of trying to model all processes within the system, the system can be abstract at the scale of interest. For example, take flow in unsaturated porous media for which the Richard's equation is often employed. However, at scales much greater than lab scale soil columns, other phenomenon such as plant roots creating macropores, burrowing animals, etc. may significantly alter the way water flows through the catchment. The issue here is of scale, as we move from the scale of a soil column to entire catchments, the dominating processes may change and it may not be possible to scale up equations and principles valid at the lower scale without accounting for these additional processes and also their interactions. Thus, the systems approach tries to abstract the system processes at the scale of interest. Surprisingly, hydrological processes tend to become relatively more tractable at larger spatial and temporal scales (Sivapalan 2003). Conceptual rainfall runoff models therefore are quite simpler as compared to their physics-based counterparts, but still need calibration as their parameters are ***effective parameters*** embedding within themselves many physio-climatic characteristics, and not easily related to catchment properties.

Modelling complex environmental systems is challenging and readers are referred to Beven (2011) for a detailed reading.

B3. Methods for prediction in ungaged basins

In many catchments across the world, streamflow measurements do not exist or are available for very short period of record, or at a temporal resolution that is not sufficient for decision making.

Thus it is not possible to do statistical analysis with the data or calibrate rainfall runoff models. In such cases methods related to predictions in ungaged basins are employed. There are three broad categories of these methods:

(i) Transfer of parameters from another catchment(s): If one can identify gaged catchment(s) similar to the ungaged catchment, parameter sets calibrated on the gaged catchment(s) may be transferred to the ungaged catchment. In this way, streamflow simulations can be obtained at the ungaged site. The challenge in this approach is that hydrologic similarity is hard to define and is a complex function of catchment's physio-climatic setting. In addition, recent work has shown that the success of parameter transfer is not bi-directional, i.e., if one donor catchment successfully transfers parameters to a recipient catchment, it is not guaranteed that the recipient's parameter sets will be equally well performing on the donor catchment. This happens because of observational errors and different sensitivity of parameters across catchments.

(ii) Regionalization of model parameters: Another way to run a rainfall runoff model at an ungaged site is to relate model parameters with observable catchment properties. Many past efforts at this have shown that these relationships are regional in nature, i.e., they work in the region where they are developed and may not generalize well. Also, the relationships are quite uncertain, which is expected based on our discussion earlier regarding ***effective parameters***. Nevertheless, in some parts of the world, parameter regionalization has been attempted and it would be useful to scan past literature/ reports to see if such relationships can be utilized.

(iii) Regionalization of catchment signatures: A third approach is based on developing relationships between catchment signatures and physio-climatic properties. Signatures are metrics that can be estimated from streamflow time series, such as the runoff ratio (ratio of long term mean annual streamflow to precipitation), slope of the flow duration curve, flashiness index, etc. It has been shown that some signatures exhibit a strong relationship with catchment physio-climatic properties. For example, runoff ratio is highly correlated with aridity index (ratio of long term mean annual precipitation to potential evapotranspiration). If a large of gaged catchments are available in a region, relationships between catchment signatures and physio-climatic characteristics can be developed (techniques such as step-wise linear regression, principle components analysis, etc. are useful for this). Then, for the ungaged site, the signature value can be obtained by using the regression relationship. Following this, model parameters can be identified that are consistent with the streamflow signature values obtained from the regionalized model.

This was a brief summary of various methods for obtaining decision relevant streamflow metrics for a site. By no means is this summary exhaustive, and readers are referred to the references already listed before and Singh et al. (2014) for more details on regionalization.

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