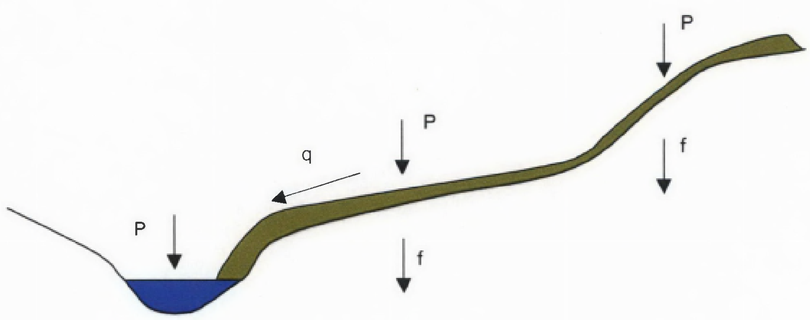
4.0 Catchment Surface Runoff

Catchment input to dams, surface runoff is a critical factor in both the physical and chemical systems within a dam and hence requires investigation as to the timing and extent of this influence. The impact that inflow has on these systems depends largely on the timing and volume of water entering from the catchment. This is itself dependant largely on the physical characteristics of the catchment and the precipitation occurring. Catchment factors such as surface retention play an important role in runoff processes whilst rainfall intensity and duration are key storm characteristics affecting the degree of surface runoff.

Surface retention includes losses of runoff due to interception from vegetal cover and depression storage (storage in puddles which form on the ground). It also includes losses due to evaporation during the storm event

4.1 Surface Runoff Mechanisms

There are a number of runoff mechanism, which may be operating independently or in conjunction with each other depending on environmental conditions. These include Hortonian Overland Flow (HOF), Saturation Excess Overland Flow and Interflow. The most widely known and evidenced is HOF. Horton (1933) insisted that the dominant mechanism for surface runoff or overland flow is that of infiltration excess runoff (Figure 4.1). Hortonian flow as it was later termed was based on observations that



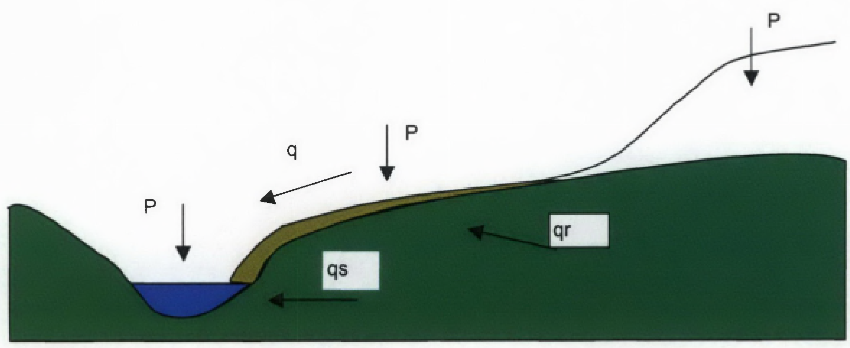
Where: P=precipitation, f=infiltration, q=overland flow

Figure 4.1 Hortonian Overland Flow.

soil has a limiting infiltration capacity after which excess water ponds on the soil surface.

When local depressions are filled and surface storage is satisfied the excess water then moves laterally filling other depressions until eventually the water freely moves across the surface as overland flow. Hortonian flow occurs predominantly in areas of sparse vegetation, compacted soils, medium to high slopes and generally locations where the infiltration capacity is naturally low or diminished by human impact (Shaw 1994).

Another method by which runoff may occur is via saturation excess overland flow (SEOF). This occurs in areas where the infiltration rate is larger than rainfall intensity. In this instance surface runoff occurs because the moisture storage capacity of the soil profile is less than the rainfall volume (Shaw 1994). Rainfall intensity does not usually exceed the site infiltration rate however sufficient water has been added to the soil profile such that it becomes saturated, allowing no more infiltration. The excess becoming overland flow (Figure 4.2).



Where: P=precipitation, q=overland flow, qs=subsurface flow, qr= interflow

Figure 4.2 Saturation Excess Overland Flow.

This type of surface runoff is common in forested areas (highly permeable soils) and locations where the water table is close to the soil surface (such as near stream beds). Runoff is influenced by geographical and topographical consideration. It is common in such areas as (Shaw 1994):

a) Where subsurface flow lines converge in slope concavities and water arrives faster than it can be transmitted down-slope by subsurface flow.
b) Concave slope breaks where the hydraulic gradient inducing subsurface flow from upslope is greater then that inducing down-slope transmission.
c) Where soil layers conducting subsurface flow are locally thin.
d) Where the hydraulic conductivity decreases abruptly or gradually with depth and percolating water accumulates above the low-conductivity layers to form perched zones of saturation that reach the surface.

Interflow is lateral movement of water within the soil profile. It can contribute to surface runoff when this water intersects the soil surface and causes localized saturation (common over the lower part of hillslopes). Under these conditions further rainfall falling on such areas cannot penetrate into the soil and thus travels across the soil surface as saturated excess overland flow with the interflow component.

4.2 Effective Rainfall

Effective rainfall is the portion of precipitation which contributes to surface runoff and hence streamflow. It is the volume of rainfall remaining after interception loss, depression storage, infiltration, groundwater recharge and evaporation have been satisfied (Equation 4).

Equation 4

$$\mathit{Effective\,Rain} = P - (I + E + \mathit{Inf} + G)$$

Where P = Precipitation, I=Interception, E=Evapotranspiration, Inf= Infiltration, G=Groundwater Recharge

4.2.1 Interception

Interception is the initial loss fraction of precipitation and will vary in magnitude in response to many factors relating to both the vegetation in the immediate area and the characteristics of precipitation producing storm event. Tree canopies, grass, shrubs, litter and moss are the primary vegetation types, which act to intercept rainfall in most rural catchments[12]. As such highly variable vegetation parameters such as leaf area, surface tension forces, size, flexibility, strength, pattern of branching, texture, and orientation of leaves directly influence the interception capacity of the vegetation type and the impact this has on runoff volume (Leonard 1965, Aston 1979, Hewlett 1982, Hutchinson et. al. 1986, Beymer 2001).

The magnitude of interception loss has been demonstrated by Linsley et al. (1988) who predicted that in a well-developed forest canopy 10–20% of annual precipitation can be intercepted. This was in conjunction with a storage capacity of 0.8–1mm. Falling in this bracket are conifers and deciduous trees, which intercept 25-30% and 15-25% of annual precipitation respectively (Linsley et al. 1988). Further work in the natural hardwood forest of Taiwan, revealed that interception can account for as much as 11.3% of total annual rainfall (Lu & Tang 1995[13]) whilst interception rates of up to 50% of total catchment water input have been observed by Cavelier $ Goldstein (1989) in the Tropical Mountain Cloud Forests in the Andes. However, in Zumbador, Venezuela this type of interception accounted for only 3.5% of water input (Cavelier & Goldstein 1989). Trimble and Weitzman (1954) established that 25% of precipitation from summer and winter storms was intercepted by 50 year old hardwoods in the Southern Appalachian Mountains.

The high interception capacity of forested areas due to canopy factors and sub-canopy factors is transformed into reduced runoff volume. In some instances an overall reduction of 15% has been observed (Kirby et al. 1991). Some of this may be attributed to interception by litter and grasses on the forest floor. Forest litter has been shown to have a maximum holding capacity of 215 – 263% of dry weight (Bernard 1963, Helvey 1964) whilst the capacity of some types of grasses to intercept rainfall is comparable to tree canopy interception (Table 4.1). Standard crops have also been investigated and have shown that interception

[12] Built structures also act to intercept rainfall however their importance in runoff estimation in rural landscapes is negligible due to the relatively minor land area covered by such structures.
[13] This however can be misleading due to the high seasonality of rainfall in the area. During the typhoon season months of summer interception accounted for only 3.9% of precipitation compared to 100% during the dryer winter months.

varies markedly with season. At the height of season oats, soybeans, corn and alfalfa are capable of intercepting 7, 15, 16 and 36% of precipitation respectively. However during low season or early summer these are reduced to 3, 9, 3, and 22% respectively.

Table 4.1 Interception of rainfall by common grass types.

Source	Common Name	Scientific Name	Interception % of Rainfall	Rainfall Event
Collins (1970)	Saltbush	Atriplex *argentea*	50	300 mm/hr
Collins (1970)	Burning Bush	Kochia *scoparia*	44	300 mm/hr
Thurow et al. (1987)	Shortgrass	Hillaria *belangeri*	18.1	Annual rainfall
Thurow et al. (1987)	Midgrass	Bouteloua *curtipndula*	10.8	Annual rainfall
West & Gifford (1976)	Sagebrush	Not known	4	Annual rainfall
West & Gifford (1976)	Shadescale	Not known	4	Annual rainfall
Rowe & Hammilton (1949)	Sagebrush	Artemisia *tridentate*	30	Summer rainfall events

Thurow et. al. (1987) estimated that saturation or storage capacity for midgrass ({sideouts gama}, Bouteloua *curtipndula*), and shortgrass ({curlymesquite}, Hillaria *belangeri*) was 1.8mm and 1mm respectively. They found that the time needed to achieve this capacity per unit dry weight of both was dependant on the rainfall intensity. Saturation was attained after 8 minutes for a 25mm/hr storm event and 5min for a 114mm/hr event. Once saturation has been achieved the excess water, minus slight evaporative losses is free to continue on towards the soil surface via stemflow or drippage. In reality interception loss is only a maximum for short duration events, which are spaced such that vegetation is able to dry before the next event.

The characteristics of the precipitation event influence the degree to which interception affects the magnitude of runoff. In particular precipitation amount, frequency, intensity, duration, type wind during storm and wind during evaporation are of importance (Hewlett 1982). In general, interception is proportionally larger for small precipitation events due to the relatively larger water requirements of the vegetation to attain saturation (Figure 4.3).

Intercepted water can be considered as a net loss from precipitation as it is lost through evaporation. It has been shown that intercepted water evaporates at a higher rate than transpiration water and cools wet foliage, which in turn cools the plant, further suppressing transpiration and encouraging evaporation (University of Regina 2003).

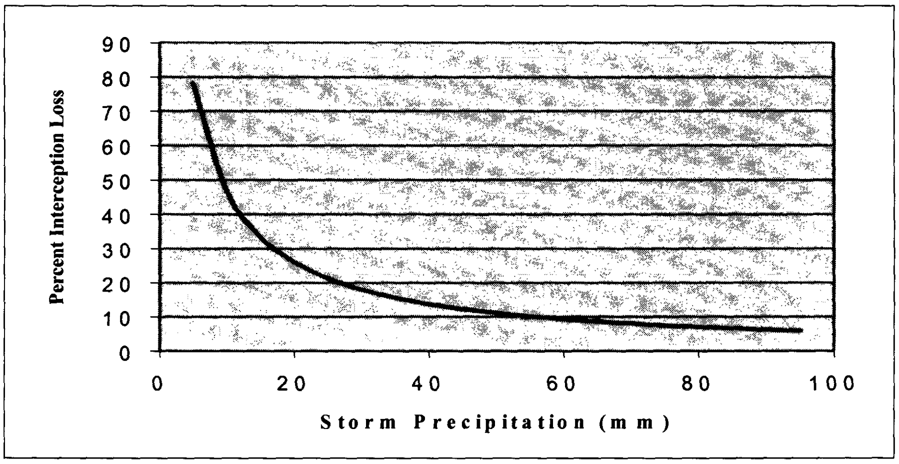


Figure 4.3 Interception loss by Oak Motte canopy as a function of storm precipitation (modified from Thurow et al. 1987).

4.2.2 Depression Storage

Depression storage is that proportion of precipitation that reaches the soil surface and does not flow laterally towards a stream or channel. Hence, a depression is any landform that retards the movement of surface water to a stream or channel (such as potholes or rabbit holes). It does not include structures such as dams.

Factors such as the nature of the terrain, slope, type of soil, time and antecedent rainfall all affect the degree of depression storage (Salas 2002). Likewise landuse and land management activities such as the creation of stock ponds, terraces or contour farming, land levelling and drainage also affect the depression storage potential of an area.

Measurement of depression storage remains difficult to achieve due to interaction with both infiltration and evaporation loss pathways. In impervious drainage areas, depression storage has been found to decrease linearly with increasing slope (Viessmann 1968 cited in Salas 2002) and exponentially with time (Salas 2002). Although these relationships are known the amount of water retained in depressions and the impact this has on runoff calculation remains difficult to ascertain. Hans and Chanasyk (2000) were unable to predict watershed runoff from two reclaimed mining sites. They found that depression storage was insignificant in the pervious sandy subsoil watershed while runoff from the other was restricted to rainfall falling directly on saturated channels. Thus they concluded that runoff magnitude and timing could not be predicted from the same parameters. Abedini (n.d.) found that in the absence of infiltration he was able to detect the effects that depressions have on runoff characteristics (R^2 = 0.9). Under these conditions parameters seemed to reflect physical catchment characteristics such as mean time of travel and mean depth of depression storage. However, when infiltration was occurring the effects of depressions on the hydrograph were effectively masked.

4.2.3 Infiltration

Infiltration is the movement of water across the soil-surface interface and represents in most cases a loss from potential surface runoff volume[14]. The factors within a catchment which influence infiltration include *slope, aspect, land use, vegetation and antecedent soil moisture.*

4.2.3.1 Slope, Aspect and Surface Area

The slope and aspect affect the velocity and routing of surface runoff and the quantity and quality of solar radiation reaching the soil surface. Likewise slope attributes such as steepness, length, width, and form all contribute to the capacity of a site to infiltrate. Areas with minimal slope steepness are conducive to infiltration due to increased travelling time for surface runoff. However, minimal slopes also receive less solar energy than steeper surfaces and thus have a larger

[14] See saturated overland flow

potential for evaporation (Nie et al. 1992, Ellanskaya et al. 1997, Nater & Bell 2002). The aspect of a site affects the distribution and intensity of solar radiation (Figure 4.4), which in turn determines the air and soil temperatures, humidity, PET, vegetation and the distribution of precipitation (Stone et al. 1998, Nater & Bell 2001).

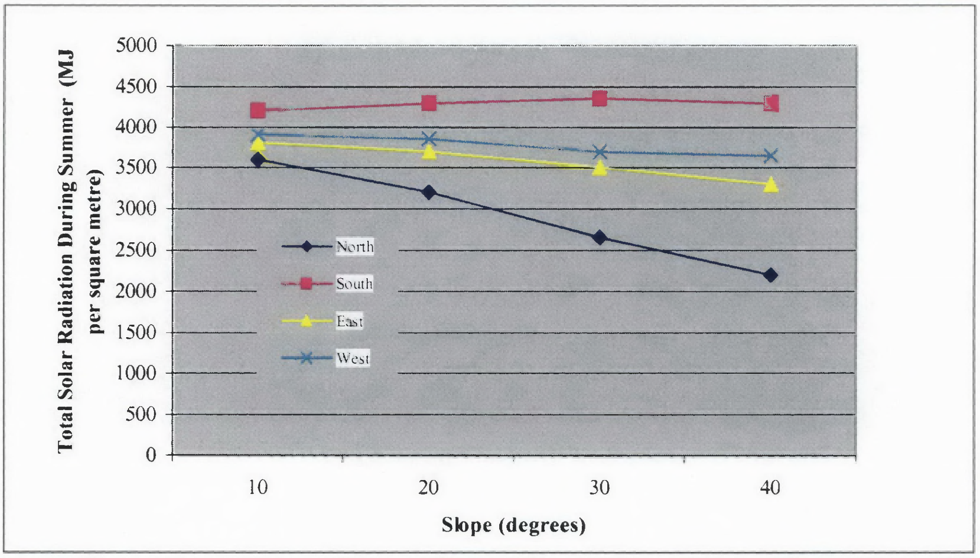


Figure 4.4 Solar radiation as a function of slope and aspect in the Okanagan area of Southern British Columbia (Hawthorne Mountain Vineyards 2002).

The incident angle of solar radiation on the earth's surface varies from a maximum in summer to a minimum in winter. In the southern hemisphere the sun reaches a maximum altitude above the horizon of 66.5° during summer solstice (December 22 when the earth is turned at an angle of 23.5° towards the sun). This is the point of maximum potential solar flux. At an angle of 60° to the horizon 95% of incoming short-wave radiation is adsorbed by water whereas at a solar angle of 5° only 40% is adsorbed, the rest being reflected. This transforms into variation in local air temperature (Holtch 1931, Wilson 1970) and soil temperatures (Cottle 1932, Dixon 1986). In the southern hemisphere north facing slopes below 23.5° S will receive greater energy than south facing slopes while in the northern hemisphere this situation is reversed.

4.2.3.2 Land-use, Vegetation and Organic Matter

Infiltration is markedly influenced by the vegetation characteristics and land-use of a catchment. .

Vegetation acts to retard surface flow, increase soil porosity and reduce soil packing from raindrops (Linsley et al. 1988). Each increases potential infiltration (Linsley et al. 1988). Vegetation also directly affects the recharge of groundwater reserves through losses via transpiration however it also provides shade, which reduces surface soil temperature and subsequently evaporation loss. Vegetation is also important in maintaining soil stability. In Texas pasture management was used as a means of increasing infiltration so as to reduce soil erosion. This was achieved by manipulating the percent ground cover (Figure 4.5). In this instance both erosion and runoff were significantly reduced with

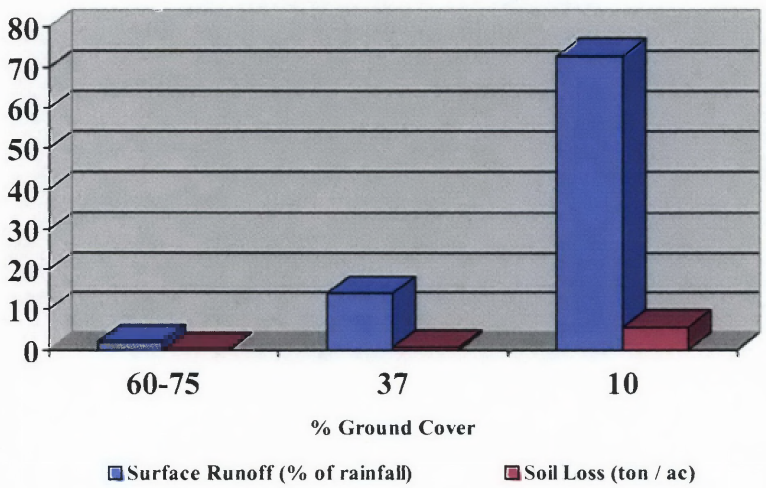


Figure 4.5 Effect of vegetation on runoff and erosion (adapted from Bailey & Copeland 1961).

increasing ground cover. Further work revealed that infiltration is also dependent on the species of vegetation (Figure 4.6). Similar results have been obtained using turf. However, for most turf lawns the effectiveness of turf to increase infiltration is limited due to the generally high compaction of normal lawn soil and the relatively shallow root depth for turf (5-10cm). In comparison to turf, native grasses and pastures tend to have deeper root systems and thus are more able to assist infiltration.

Deep rooted species are also capable of creating water deficits within the soil (Table 4.2). This attribute is important in reducing the susceptibility of an area to dryland salinity, which is an increasing concern in Australia (Bennet et al. 2002). Lucerne, the most widely used perennial pasture can root to 3 m creating a dry soil buffer of up to 270cm (Bennet et al. 2002). Annual pastures such as

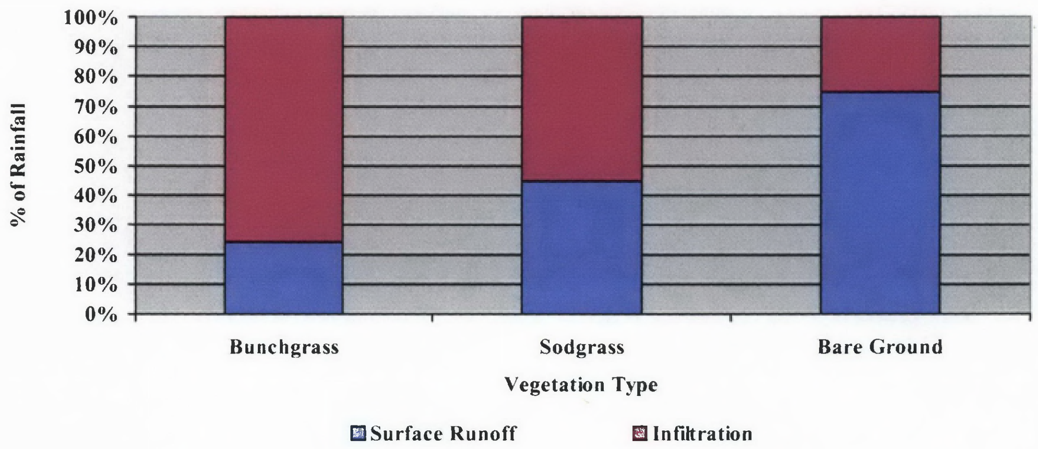


Figure 4.6 The influence of vegetation type on surface runoff and infiltration (adapted from Blackburn et al. 1986).

serradella (O. *compressus*) have also been known to deep root (195 cm) on deep sandy soil, however in general annuals tend to average around 50 –70 cm root depth (Hamblin & Hamblin 1985).

By creating a water deficit, deep rooted plant species provide available pore space for surface water to continue to move into the soil profile thus reducing runoff and recharge. Land-use practices such as tilling, ripping, cropping, and stocking can also affect the infiltration of an area. Schumacher (1994) found that infiltration rates under tillage were greater than for no-till. Further studies showed that adding manure to the soil increased macro-porous flow as a result of an increase in the number of earthworms and loosening of the soil on application.

Organic matter increases the porosity of a soil. The USEPA (1999) investigated the effects of compost addition to soil on infiltration rates and hydrograph attenuation. It was found that compost amended soil could hold double the moisture as unamended soil and subsequently reduced the hydrograph peak. The infiltration rate was also 1.5 - 10.5 times the normal rate. Unfortunately the runoff from the compost amended sites contained 5 - 10 times the concentrations of nutrients and sediments although these concentrations rapidly reduced with time (USEPA 1999).

Table 4.2 Maximum dry soil buffer created by annual and perennial pasture and crop plants (from Bennet et al. 2002).

Species	Location	Dry Soil Buffer (cm)	Reference
Annual crop	Rutheglen	84	Ridley et. al. 2001
Annual pasture	Junee	89	Sandral, Dear, Virgona (unpub data)
Annual crop	Tatura	91	Whitfield 2001
Annual pasture	Tatura	100	Whitfield 2001
Austrodanthonia	Junee	140	Sandral, Dear, Virgona (unpub data)
Consol lovegrass	Junee	146	Sandral, Dear, Virgona (unpub data)
Cocksfoot	Junee	155	Sandral, Dear, Virgona (unpub data)
Lucerne	Junee	157	Sandral, Dear, Virgona (unpub data)
Phalaris	Junee	162	Whitfield 2001
Cocksfoot	Tatura	170	Lolicato2000
Birdsfoot trefoil	Tatura	200	Lolicato 2000
Phalaris	Tatura	210	Lolicato 2000
Lucerne 2m	Ruthergler	228	Ridley et al. 2001
Lucerne 2m	Tatura	230	Lolicato 2000
Lucerne 3m	Ruthergler	306	Ridley et al. 2001

Land use such as farming, residential and forestry activities also influence the infiltration capacity of an area. Farming, in particular livestock impact directly on the infiltration capacity of an area. Livestock are prone to trampling the soil which increases soil compaction, disrupts soil structure and stability of water stable soil pores and destroy cover provided by algae and lichens through contact and grazing (McGinty et al. 1991). Each of which decreases the infiltration capacity of a soil (McGinty et al. 1991).

4.2.3.3 Soil Moisture

Soil profile and acting forces

The soil profile is divided into four main compartments (Figure 4.7). Within each of these the forces acting on soil moisture vary in nature and importance. In general soil moisture is subject to gravity, capillary and hydroscopic forces. Under the influence of gravity water is drawn deep into the soil (percolation) and laterally (lateral / inter flow). Gravitational water is usually not available to plants and in some case may result in death of plants through saturation of pore spaces and denial of necessary oxygen to root systems. Capillary forces act to retain water in small pore spaces in unsaturated and capillary fringe areas.

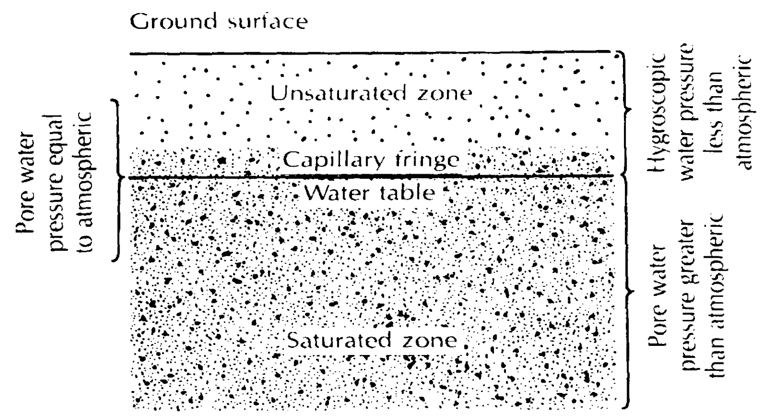


Figure 4.7 Zones within the soil profile (Northern Arizona University 2002).

Capillary forces include those of adhesion and cohesion (Figure 4.8). Adhesions is the force which initially bonds moisture to soil particles whereas cohesion is the bond between water molecules. Water held by adhesion and cohesion is available at varying water tensions determined by the water layer thickness with regard to the soil particle (Figure 4.9).

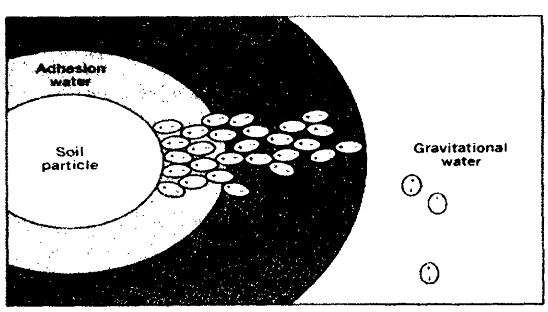


Figure 4.8 Water held around a soil particle by forces of adhesion and cohesion (Northern Arizona University 2002).

Hydroscopic forces are responsible for maintaining an ultra thin layer of unavailable moisture to soil surfaces. Essentially it is very strong adhesive forces which result in this. In soils with high surface areas to volume ratios (fine grained) have comparatively more water held as hydroscopic than coarse grained soils (NC University 2002).

Soil Moisture Content

The amount of moisture potentially present within a soil profile is determined by the permeability and porosity of the soil. These in turn are a reflection of the soil structure, grain size and distribution (Table 4.3). It follows that infiltration rate or

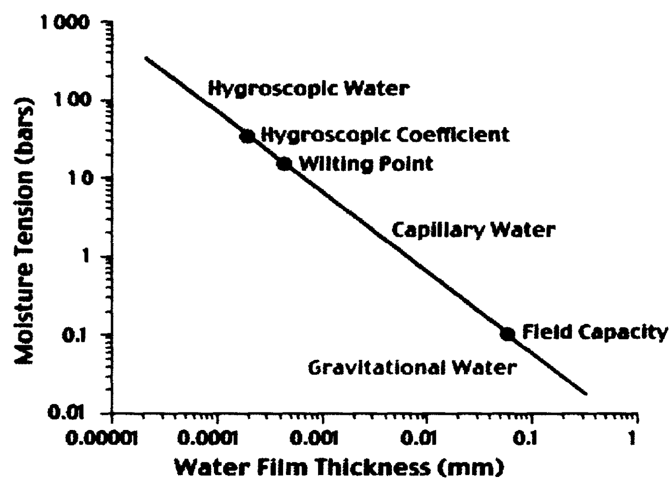


Figure 4.9 Relationship between soil water film thickness and moisture tension (University of Regina 2002).

hydraulic conductivity in unsaturated media is determined by the antecedent moisture conditions and tension forces within the soil profile (University of Regina 2003) according to Figure 4.10.

High soil porosity does not necessarily mean water will pass through the soil readily (as is the case with clay soils where moisture is held in capillary pores). Fine-grained soils tend to retain water whereas coarse-grained soils tend to facilitate freer movement of water through its layers and thus have a relatively higher specific yield. The presence of macropores (from burrowing animals and root systems etc.) will increase the specific yield of a soil.

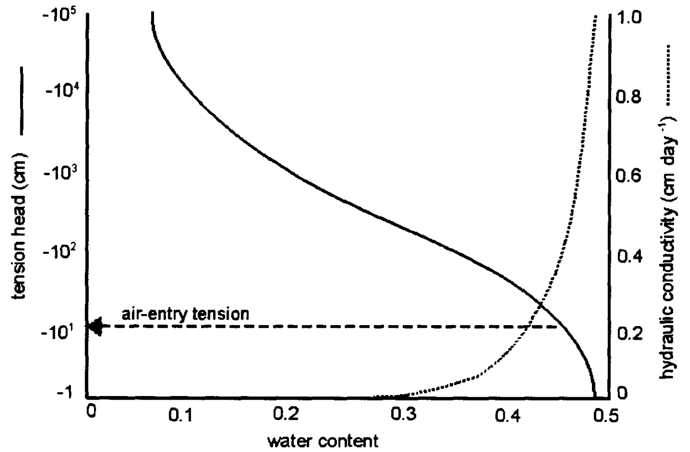
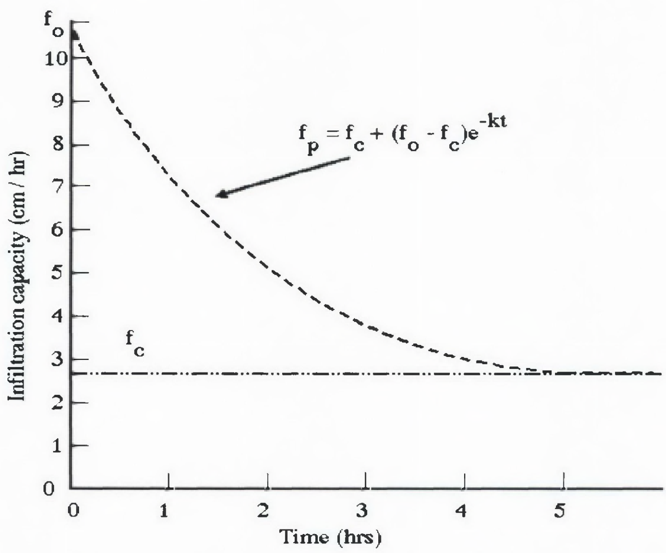


Figure 4.10 Relationship between soil water content, hydraulic conductivity and tension head for unsaturated soil conditions (University of Regina 2003).

Table 4.3 Approximate average porosity, specific yield, and permeability of various material (from Linsley et. al., 1988).

Material	Porosity %	Specific Yield %	Permeability $m^3\ day^{-}m^{-2}$
Clay	45	3	0.0004
Sand	35	25	41
Gravel	25	22	41000
Gravel and Sand	20	16	410
Sandstone	15	8	4.1
Dense limestone and shale	5	2	0.041
Quartzite, granite	1	0.5	0.0004

In a homogenous soil infiltration decreases gradually until the zone of aeration of the soil is saturated (Figure 4.11). For non-homogenous soils the infiltration rate will be limited by the least permeable soil layer. Unless the soil layer is thin or highly permeable infiltrated water rarely passes below 150cm below the soil surface (Linsley et. al. 1988).



f_o = Initial infiltration capacity, f_p = Infiltration capacity, f_c = Equilibrium infiltration capacity, e = naparian base, k =constant, t

Figure 4.11 Infiltration Vs time for homogenous soil (Northern Arizona University 2002).

Affecting those parameters in Table 4.4 is the degree of compaction of the soil. The USEPA (1999) tested the infiltration rate of a range of soils from non-compacted sandy soil to compacted dry-clayey soils (Figure 4.12). The sandy soil was most affected by compaction and little affected by the initial moisture content. Although the infiltration rate for clay was reduced on compaction it was more influenced by the initial moisture content.

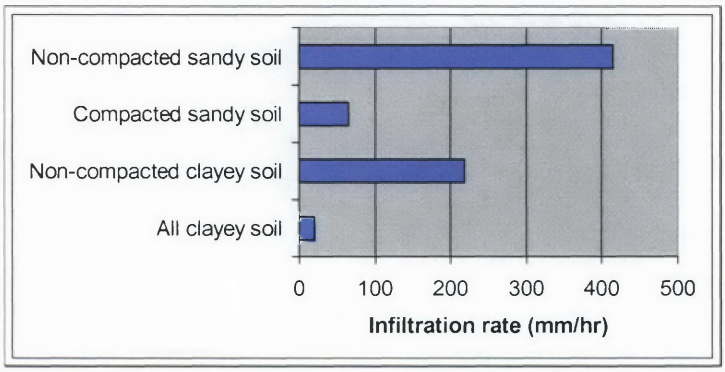


Figure 4.12 Average infiltration rates for different categories of soil (USEPA 1999).

Soil moisture is typically expressed as percent water on a volume basis (Equation 5) or percent weight of water (Equation 6). Some common soil moisture characteristics are given below (Table 4.4).

Equation 5

Percent water on volume basis (θ) = $\dfrac{Volume\ of\ water(mm) \times 100}{Volume\ of\ soil(m^3)}$

Equation 6

Percent mass of water (θ_m) = $\dfrac{Mass\ of\ water(gm) \times 100}{Mass\ of\ solids(gms)}$

Table 4.4 Soil moisture content, weight fraction (Linsley et al. 1988)[15].

	Percent Dry Weight(%)			
	Field Capacity	Wilting Point	Available Water	Density (kg/m^3 dry)
Sand	5	2	3	1520
Sandy Loam	12	5	7	1440
Loam	19	10	9	1360
Silt Loam	22	13	9	1280
Clay Loam	24	15	9	1280
Clay Loam	36	20	16	1200
Peat	140	75	16	400

Soil moisture is commonly measured using quantitative and qualitative methods. Of the quantitative methods gravimetric sampling is the most widely utilized as it is a relatively uncomplicated process. It is this method against which neutron scattering and di-electric constant measurement (other quantitative methods) are calibrated. Unfortunately, the gravimetric method is labour and time intensive and requires large sample numbers to overcome the inherent spatial variability in soil characteristics. These types of methods give feedback on how much water is present in the soil from which water availability may be inferred. On the other hand qualitative methods specialize in indicating the state of availability of soil moisture and give no definite numbers on water content. Under this heading are tensiometers and porous blocks.

[15] For classification see p84-85, Shaw (1988).

In recent times improvements in technology have presented various other methods for measuring soil moisture. These include methods using passive and active radar from both remote and local sensors (Barros & Bindish n.d., Calvet et al. n.d.) and methods using scatterometers (Romshoo et al. 1999).

4.2.4 Evapotranspiration

Evapotranspiration includes water lost to the atmosphere through evaporation of moisture from the surface of various sources such as soil and plants. It also includes the moisture lost from the stomatal openings of plants during metabolism (transpiration).

4.2.4.1 Evaporation

Evaporation occurs from the surface of precipitation as it falls. However, the relative significance of this loss is inversely proportional to the magnitude of the event. Such that for rainfall-runoff analysis of extreme or large precipitation events it may be ignored (Allen et al. 1988).

Evaporation rates vary depending mainly on meteorological conditions the nature of the evaporating surface and to a lesser degree, water quality[16] (Kohler 1951, Allen et al 1988).

Meteorological factors

The most important meteorological factors controlling *potential evaporation*[17] are the amount of energy being supplied *i.e.* solar radiation (and hence air temperature), the moisture content of the atmosphere (humidity) and the rate of movement of air across the surface (windspeed) (Allen et al. 1998) (Figure 4.13).

For liquid water to change into vapour at 20°C requires 2.45MJ of heat energy per square metre for every mm of water (Allen et al. 1998). This energy is

[16] Essentially water quality effects on evaporation can be ignored as they tend to result in very little overall change in evaporation however they will impact on the potential transpiration from plants growing in poor water quality conditions (Allen et al. 1998).

[17] Potential evaporation is the amount of evaporation that would occur from the soil if it were never short of water (Farnsworth et al.1982).

supplied by the sun in the form of solar radiation. The amount of which varies with latitude, season, time of day and sky conditions (see Section 4.2.3). However *actual evaporation* at a site will depend on local air temperature, vapour pressure, atmospheric pressure and of course available moisture (Chayttopadhyary & Hulme 1997, Allen et al. 1998).

Evaporation increases as wind increases however storage capacity on the vegetative surface is reduced. In general high wind speeds increase interception during long storms and decrease it during short storms.

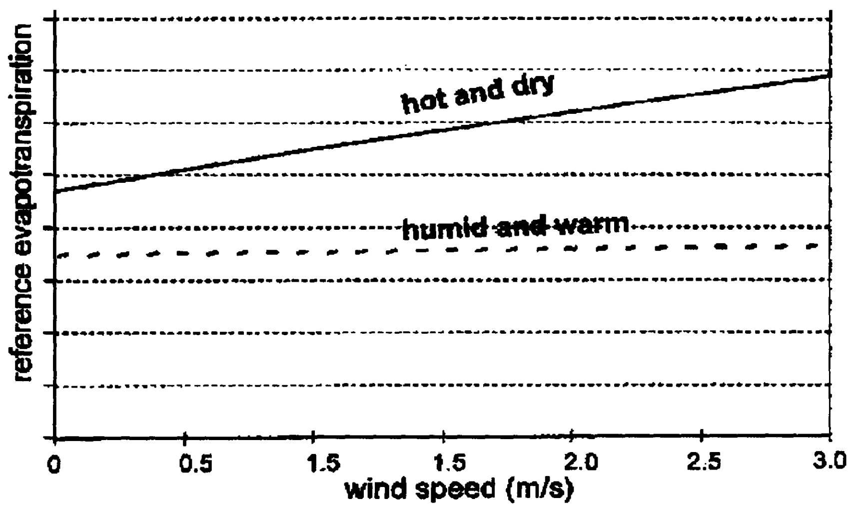


Figure 4.13 The effect of wind speed on evapotranspiration in hot-dry and humid-warm weather conditions (from Allen et al. 1998)[18].

Nature of the Evaporative Surface

All surfaces are potential sources of evaporation. The rate and degree of which will depend upon many factors including the contact angle at which the water adheres to surface (Bachmann et al. n.d, Chandra et al. 1996). Within soil primary factors include existing moisture content, soil composition and vegetative cover and root characteristics.

Saturated soil has the same initial evaporative potential as a water body under the same meteorological conditions but will rapidly decrease as soil moisture is removed and the mechanisms for moisture movement breakdown (Allen et al. 1998). Murphy & Lodge (2001) found that litter affected this relationship by

[18] Although the graph is for evapotranspiration the relationship between air temperature, windspeed and humidity are representative of those for evaporation also.

increasing the albedo of the soil surface and depressing evaporation. This was more marked for wet soil conditions where evaporation was the major mode of soil water loss (being 3 times greater than for dry conditions). Under dry conditions most water was lost to the atmosphere via transpiration which depended on plant root density whilst bare plots exhibited evaporation rates equal to potential and greater than pan measured values for wet surface conditions but was relatively insignificant for dry surface plots.

4.2.4.2 Transpiration

Most transpiration occurs during daylight hours (95%) and decreases in the presence of salinity or increasing water tension (Allen et al. 1998). It ceases altogether when the soil water temperature reaches 4°C. In building plant material (photosynthesis) a plant transpires 800 times more water than it uses to build plant biomass (Allen et al. 1998). Plants control transpiration by the closing and opening of the stomata. This action is itself dependent on meteorological conditions and the stage of growth, root depth, cover density and type of plant in question. Some authors have even suggested that the amount of CO_2 present in the atmosphere may also modify this relationship (Morison 1987, Field et al. 1995, Gifford et al. 1996, Amthor 1999).

Dunlop and Shaykewich (Dunlop & Shaykewich 1982 cited in AFRD 2002a) found that evapotranspiration was 30% of potential for the early stages of growth for some wheat species. Following complete cover evapotranspiration reached potential but again declined when the crop matured. A rate of transpiration from a 100% vegetative cover with unlimited water will transpire approximately the same amount of water as an open body of water under the same meteorological conditions (Shaw 1994). However, in reality soil attains saturation on very few occasions throughout any given year.

Hurtalova et al. (2001) found that transpiration from different crops was governed by the ability to capture solar radiation and not so much by soil moisture conditions. Given drier soil conditions (above wilting point) they found that a wheat crop was able to transpire more moisture than both maize and sugar beet due to a larger Leaf Area Index (LAI). In conditions where moisture may be limiting the ability to provide sufficient moisture for plant growth is governed by root characteristics of the plant. The majority (50-60%) of transpired water comes from the top 30 cm of soil with lesser amount being taken from an average maximum soil depth of about 1.5m (AFRD 2002b).

4.2.4.3 Determination of Evapotranspiration

Evapotranspiration can be determined either directly or indirectly. The choice of method is largely a function of available resources. Direct methods involve the calculation or measurement of evaporation and/or transpiration (Thornthwaite & Holtzman 1942, Penman 1948, Turc 1954, 1955, Fox 1956, Morton 1983 Cited in Linsley et al. 1988). Common direct methods include the water budget, energy budget, mass transfer, combination of these and empirical methods.

Indirect calculation usually involves at least three steps. Firstly the potential evapotranspiration of a reference crop with unlimited access to nutrients and water is determined (ET_p). This is then related to the crop in question under optimal conditions. Finally the actual environmental conditions and management factors are integrated into a stress factor (K_S) to give the actual evapotranspiration from the crop in question (Penman 1948, Thornthwaite 1948, Pruit & Laurence 1968, Priestly & Taylor 1972, McNaughton & Black 1973, Jury & Tanner 1975).

4.3 Storm Characteristics

Specific characteristics of storms affecting runoff prediction relate to the temporal and spatial variability of rainfall over the catchment and the type of precipitation occurring (Wei & Larson 1971). Temporal and spatial variability is generally a reflection of the type of storm producing the precipitation. The four main types of storms are cyclonic, frontal, convective and orographic (Linsley et al. 1988). Convective storms lose proportionally less precipitation to interception and infiltration than frontal storms and are more likely to produce runoff (Hemandez & Nachabe 2000).

The type of precipitation will also influence the runoff characteristics of a storm event. The different forms of precipitation include: drizzle, rain, snow, hail, sleet, rime, and glaze. The relative importance of each of these to the overall water budget of a catchment will be geographically dependent (Linsley et al. 1988). Research has shown that snow and other frozen forms of precipitation contribute relatively smaller proportions of water to soil moisture compared to rain but will upon melting follow similar paths as rainfall (Linsley et al. 1988, Shaw 1994). These forms of precipitation have also been shown to be more susceptible to interception.

The intensity of a storm may directly change some soil and hence catchment characteristics. Large drop sizes and or high intensity rain may compact the surface. This type of rain will also reduce surface roughness, break up soil aggregates and cause soil sealing. This will result in reduced surface retention increased runoff flow velocity and greater potential for erosion. Hail will likewise cause changes to surface conditions of the catchment similar to those caused by large raindrops.

The USEPA (2001) found that infiltration rates change throughout the duration of a storm (Figure 4.14). This variation made it difficult to select a suitable value for infiltration rate in further studies. Hence they concluded that it may be best to assume a relatively constant rate of infiltration throughout an event based on the mean rate and a Monte Carlo procedure for describing the random variation about the mean, possibly the coefficient of variation (USEPA, 2001).

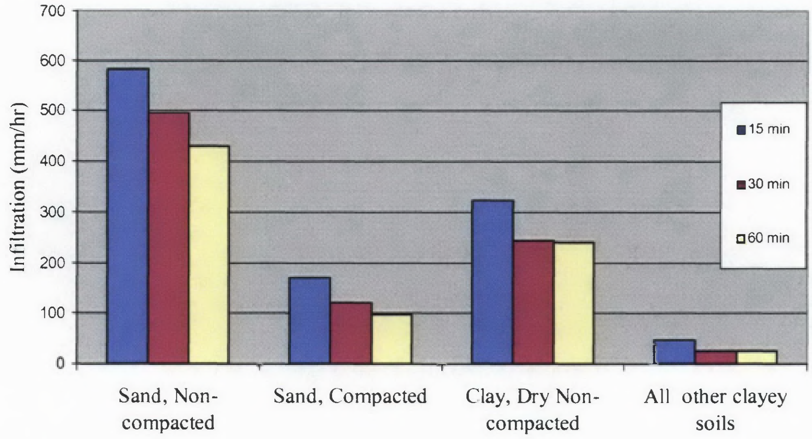


Figure 4.14 Soil infiltration rates for different categories of soil and storm duration for similar return intervals.

Hemandez & Nachabe (2000) also showed the depth of runoff for sandy soil is influenced by the depth to the water table (Figure 4.15). For deep water tables frontal storms produced a constant proportion of runoff. Rainfall variability exhibited a marked influence on the output from Hortonian type runoff models.

Hemandez and Nachabe. (2000) classified precipitation into frontal or convective events. They were then able to define temporal constraints on rainfall measurements such that the error of predicted runoff volumes was measurable and reduced. From this they concluded that a time step of 5 minutes would be optimal for convective storms. This would result in a possible 10% error (Figure 4.16). Rainfall events were classified according to Bosch and Davis (1999) (ie. 1 hour inter event for convective and 8 hours for frontal).

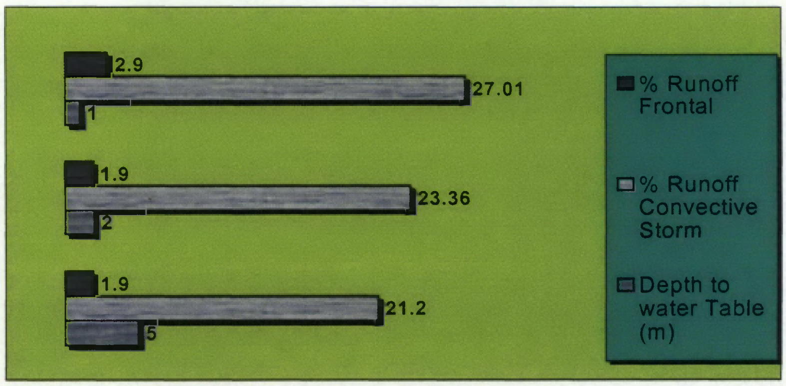


Figure 4.15 Runoff from convective and frontal storms on sandy soil (Hermandez & Nachabe 2000).

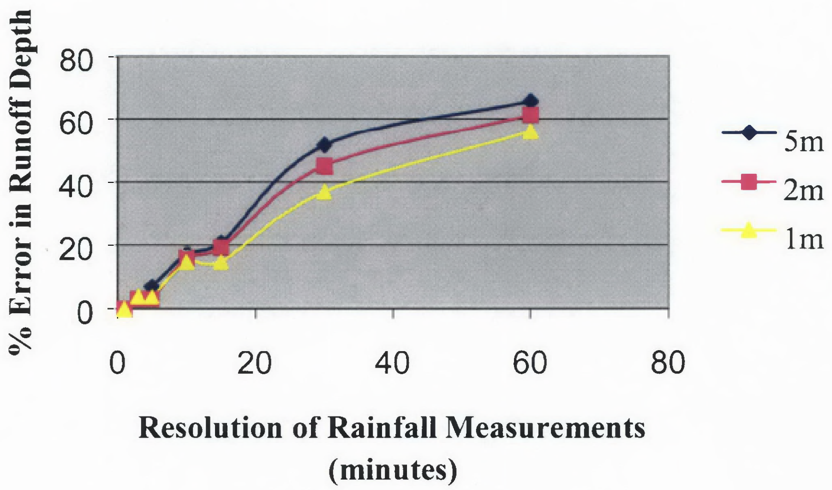


Figure 4.16 **% error of Hortonian type runoff simulations for convective storms for varying temporal rainfall measurements and depths to water table.**

4.4 Runoff Modelling and Prediction

The objective of the modelling will determine the type of method used which in turn dictates the characteristics of the data required. For example, peak discharge prediction requires small-scale measurement of rainfall (minutes to hours) while for storage assessment annual figures of rainfall, snow melt etc. may be adequate. Event based runoff prediction requires rainfall measurements over the entire basin for the duration of the event (Linsley et al. 1988, Shaw 1994).

Catchment runoff models belong to one of two broad families: Those which simulate and/or describe the catchment response based on observed events, deterministic models and those which seek to reproduce catchment responses statistically, stochastic models (Linsley et al. 1988). Within both groups specific models may be conceptual or empirical, lumped or distributed, continuous or event based and complete or partial. The simplest type of deterministic model is the 'Black Box' model. A 'Black Box' model does not try to describe the individual components but just represents the transformation of the input data (rainfall) into output (runoff). Such models assume no major landuse changes which could affect the relationship which the mathematical component of a black box model is based on.

Conceptual models mathematically define the processes involved in the rainfall runoff relationship. They are mostly deterministic and simplified representations of these processes whereby the catchment is divided into linked storage components. Of the many models proposed most include surface storage (interception and depression), ground water recharge, soil water and channel storage components. Given that these models are essentially simplified representations they still however rely heavily on a large number of parameters and extensive data sets.

Empirical models rely on observed data and as such reflect the characteristics for that data set. They can be subject to large errors, particularly when the boundary conditions for the initial relationship are no longer valid (as is the case with land-use changes) (Linsley et al. 1988).

Lumped models assume homogeneity of input data over the entire area while distributed models allow for variation between locations of both storm and catchment characteristics (Linsley et al. 1988).

Continuous models produce regular output while event based models are concerned with the catchment response due to a specific storm event. The model may also be general, being able to use in other catchments, or specific, being only applicable to the region in which it was designed (Linsley et al. 1988).

Complete models detail most, if not all the hydrological components of the catchment whilst partial models deal specifically with certain aspects of the catchment components, for example, estimation of only overland flow from rainfall (Linsley et al. 1988).

4.4.1 Deterministic Models[19]

Runoff generation for deterministic models may subscribe to one or many combinations of methods including: Unit hydrograph Method (UHM), Rational Method, Horton's Equation, Variable Source Area method (VSA), Dunnean Overland Flow Variable Source Area method and soil moisture accounting. Often as already stated these may be oversimplification of the mechanisms involved. Historically, many have relied on the Unit Hydrograph Method (UHM) when data was limited such as Nash (1957, 1960) and Dooge (1959) (Nash 1957, 1960, Dooge 1959 cited in Linsley et al. 1988) or have opted for the even simpler Rational Method first proposed by Linsley et al. (1949) (Linsley et al. 1949 cited in Linsley & Nachabe 1988) to characterize catchment response.

The UHM assumes catchment characteristics are constant over time, rainfall is uniform and constant over the entire catchment and runoff is directly related to the excess rainfall falling over an area. These assumptions have proved acceptable in many situations such that the UHM remains in extensive use today amongst hydrologists and engineers (Shaw 1994). However for improved performance basins should not exceed 5km^2 when using the UHM (Hahn et al. 1982 cited in Shaw 1994).

The Rational Method as proposed by Linsley et al. (1949) (Linsley et al. 1949 cited in Linsley & Nachabe 1988) (Equation 7) is possibly the most widely used method of calculating peak discharge from rainfall events (Hann et al. 1982 cited in Shaw 1994). It assumes rainfall intensity is uniform for the duration of the event and all measured rainfall contributes to flow.

Equation 7 $$Q_p = 0.278CiA$$

Where: C is the coefficient of runoff (mmh^{-1}), i is the intensity of rainfall in time T_c and T_c is the time of concentration, A is the area of the catchment (km^2, Q is the discharge (m^{3}s^{-1}).

Hortonian models are based on Hortons equations given in Section 4.2 whereas the Variable Source Area approaches of Hewlett and Hibbert (1963) and Dunne

[19] In this instance conceptual models are categorized as deterministic due to the ever increasing level of sophistication and detail allowed the modeller through the use of computers.

and Black (1970) allow for spatial variability in the contributing area of runoff acting in an otherwise Hortonian manner.

These many and varied approaches to estimating runoff have given rise to a plethora of computer models ranging from the simple to complex. Modelling software allows most situations to be modelled to varying degrees of accuracy and sophistication. In common use are the Stanford Watershed Model (SWM) and the USDA Hydrograph Laboratory Watershed Model (USDAHL).

The SWM is a continuous hydrograph model based on hourly rainfall which is assumed to be evenly distributed over the entire catchment. The USDAHL estimates runoff for small catchments from measured values of precipitation with consideration given to infiltration, evapotranspiration and routing coefficients. Likewise it also assumes even rainfall distribution over the watershed. Other popular models relate to the SCS method which predicts both volume and rate of runoff from agricultural catchments based on catchment characteristics such as slope, soil type, cover, land-use and land cover characteristics. The early form of the SCS model NEH-4 model allowed modelling of the complete hydrograph. The later model, TR-20 specialised in part hydrograph modelling. Both models rely on 24-h rainfall.

The runoff-routing model first proposed by Laurenson (1964) is popular and has had many improvements over the years. The model is based on the routing of effective rainfall through the catchment using non-linear relationships. Originally it relied on the partitioning of the catchment by isochromes from the outfall with a relative time-area diagram. It has since been simplified by dividing the catchment into sub-regions. This has promoted the use of distributed data and allows changes in the local catchment storage and land use functions to be incorporated (Shaw 1994). In modern catchment computer models it is a component of RORB.

Some conceptual models strive to include all aspects of the hydrological cycle. These models, termed component models, require extremely large data sets and a high degree of detail. Often this limits the use of such models. With increase information and computer power however these type of models are becoming more popular (Shaw 1994). Examples of component models include the French developed SHE model and TOPMODEL (TOPography based Model).

4.4.2 Stochastic Models

Stochastic models allow time-series assessment of data, which permits the evaluation of trends, periodic, randomness and catastrophic features of the data. The basic time series can be represented as (Equation 8):

Equation 8

$$X_t = [T_t, P_t, E_t]$$

Where: X is the time series, T is the trend component, P is the periodic component, E is the stochastic component at time t.

Difficulties arise with time series when there is zero and non-zero values in the data. On such occasions multiple steps are required to represent the data correctly (Shaw 1994). An example of this would be the rainfall data. Firstly occurrence or non-occurrence of rainfall would have to be modeled and then the magnitude of any falls.

Popular stochastic models include the Thomas-Fiering model, and the family of models known as the ARMA (Auto-Regressive Moving Average) models which combine the direct correlation properties of the data with an updated data mean (Shaw 1994).

Stochastic models also introduce to the largely deterministic process of runoff generation a randomness component that can be used to represent data such as rainfall, infiltration and discharge. Detailed modelling of catchment response to rainfall should include a stochastic component within the model structure.