2. GENERALITIES OF THE STATE OF THE ART

As mentioned in chapter 1, the one-dimensional *Transient Storage* (TS) and the *Aggregated Dead Zone* (ADZ) models are here described, as well as some of their extensions to consider uncertainty related with the set of variables and parameters in which they are based. On the other hand, generalities of stream classifications are mentioned, emphasizing in the method posed by *Flores et al*. [2006], which is based on stream gradient and the specific stream power, and can be taken to regional application by using geospatial data.

2.1 TRANSIENT STORAGE -TS- MODEL

Figure 2-1 displays a prismatic channel having a length L [L] and a cross sectional area A [L^2], which transports a discharge Q [L^3T^{-1}] moving with a mean velocity U [LT^{-1}]. Using a scheme based on the Advection-Dispersion equation (ADE), these variables, together with the longitudinal dispersion coefficient D [L^2T^{-1}], allow performing solute transport simulations along the reach segment. However, to consider transient storage processes within the reach, *Bencala and Walters* [1983; in *Lees et. al*, 2000] proposed and extension of the ADE model by introducing an storage zone characterized in terms of the contact surface, A_s [L^2], through which the solute mass can be exchanged either from or to the main channel. The exchange rate is set proportional to the concentration difference in the main channel, c [ML^{-3}], and the storage zone, c_s [ML^{-3}], where α [T^{-1}] denotes the proportionality constant.

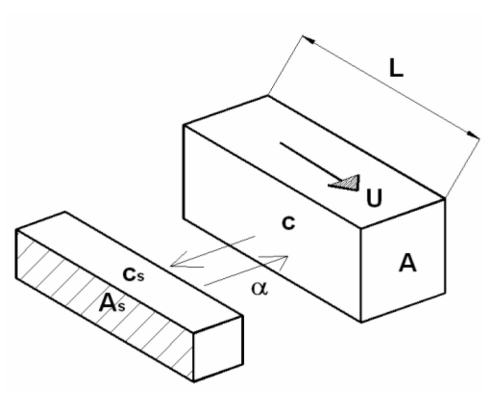


Figure 2-1. Topological representation of the TS model

The simplest form of the governing equations for the TS model is given by equations (2-2) and (2-3) [*Kazezyılmaz-Alhan y Medina*, 2006; *Meier y Reichert*, 2005; *Ge y Boufadel*, 2006; *Schmid*, 2008]. For steady and uniform flow the model parameters, *i.e*, those which must be estimated for simulation purposes, are A_S, α, , D, and U. Several shortcomings have been mentioned regarding the TS model structure, but maybe the more relevant is the lack of physical meaning of the parameters A_S and α, in terms of the impossibility to connect them with any field measurement.

$$\frac{\partial c}{\partial t} = -U\frac{\partial c}{\partial x} + D\frac{\partial^2 c}{\partial x^2} + \alpha\left(c_s - c\right) \qquad \text{2-1}$$

$$\frac{dc_s}{dt} = \alpha\frac{A}{A_s}\left(c - c_s\right) \qquad \text{2-2}$$

Some extensions of the TS model account for considering storage processes operating at different time scales. Such diversity could be driven by either surface mechanisms in large reaches featuring several stream types or a dominant surface process combined with high rates of hyporheic exchange. The second case gets significance in stream systems featuring rhythmic bed forms, where is likely to have coarser sediment, but also in those cases having large alluvial deposit extensions [*Buffington and Tonina*, 2009]. *Kazezyılmaz-Alhan and Medina* [2006] posed an extension where the storage zone account for surface and sub-surface exchange and where advection is allow between the storage reservoirs representing the hyporheic zone. Additionally, the improve model represents the mass transport due to the mass flux between the main channel and the storage zone, instead of the mass flux due to the gradient concentration between them. Despite of the improvements in the physical representation of the storage mechanisms, the set of additional parameters becomes a limitation of the approach.

An alternative approach was previously proposed by *Choi et al.* [2000] including a second storage zone completely independent of the other, and behaving according with the original model characteristics. For all the analyzed cases, excepting those they called “competitive” between the two storage zones, only one appear to be useful taking into account not only the model performance but also its parsimony.

2.2 AGGREGATED DEAD ZONE -ADZ- MODEL

Beer and Young [1983; en *Lees et al.*, 2000] proposed a reach-scale model where the storage processes spread out within the stream reach are lumped into a unique aggregated dead zone. The model uses the concept of two zones, a first region where the solute is transported during a period τ by pure advection and is therefore incompletely mixed, and a second well-stirred region where the solute is dispersed before being released at the end of the stream reach [*Richardson y Carling*, 2006]. Figure 2-2 illustrates the model topology.

The second region is precisely the ADZ element for the reach, whose water volume V is only a fraction DF (Dispersive Fraction) of the overall reach volume V_{Total}. Besides, the solute entering the *ADZ* zone resides a time T_r. Thus, the entire residence time, t_m, of the solute within the reach is given by equation (2-4).

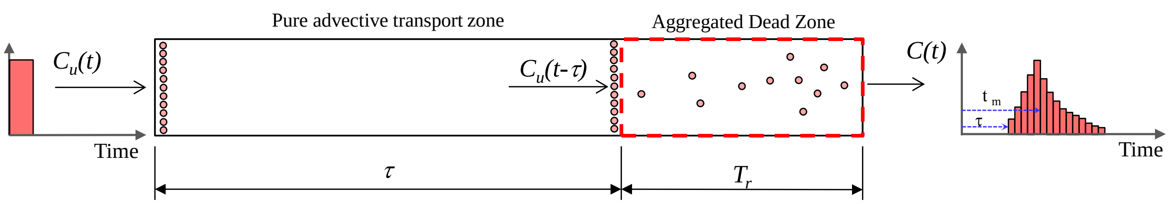


Figure 2-2. Topological dicretization of the ADZ model

$$t_m = T_r + \tau \qquad 2\text{-}3$$

$$V = DF\left(V_{Total}\right) \qquad 2\text{-}4$$

The fundamental equation of the ADZ model for steady flow is given as follows:

$$\frac{dC(t)}{dt} = \frac{1}{t_m - \tau}\left[C_u\left(t-\tau\right) - C\left(t\right)\right] \qquad 2\text{-}5$$

$$DF = 1 - \frac{\tau}{t_m} \qquad 2\text{-}6$$

where t represents time, $c_u(t)$ [ML^{-3}] is the upstream concentration entering the reach, $c(t)$ [ML^{-3}] is the output concentration at the downstream boundary, t_m [T] is the mean travel or residence time and τ [T] is the time delay or first arrival travel time.

Unlike the TS model parameters, the travel times t_m and τ are readily measured in the field by using the breakthrough curves obtained performing tracer experiments. This aspect has been highlighted as one of the advantages of the ADZ model.

Using two or more *ADZ* zones within the same reach allow extension of the model to deal with more complex stream settings, which can be set as serial ADZ regions having different characteristic time scales, or even parallel systems resembling anastomosing or braiding streams.

2.3 PROBABILISTIC APROACHES

The structure of the TS and ADZ models has proved to be efficient for solute transport simulations for a wide range of morphological settings and hydrological stages lower than the bankfull stage. Nonetheless, *Ge and Boufadel* [2006] showed for the TS model that the parameters obtained for a reach through calibration procedures, are not necessary representative at subreaches given the lack of connection between the "*smoothed"* parameters obtained, with those effectively representative at shorter scales. This is pointed as one of the reasons for the non-identifiability or equifinality in the TS model, which refers to obtaining similar model fits using different parameter sets.

As previously mentioned, the more simple form of the ADZ model given by equation 2-6 (a first order approach) allow describing stream segments having a dominant retention timescale, but longer reaches, featuring more complexities, likely require more than one aggregated dead zone.

Among the strategies to overcome the uncertainty arising in those cases some probabilistic approaches seem to be promissory. Using a general distribution of residence times (RTD), equations (2-2) y (2-3) in the original TS model by *Bencala and Walters* [1983] can be generalized as shown in equations (2-8) y (2-9) [*Zarnetske et al*., 2007; *Gooseff et al.*, 2003]. There, instead of using the first order transfer coefficient α, the transfer probability distribution function $g^*(t)$ is introduced to represent the probability that a solute molecule entering the storage zone at t=0 remains there after a time t.

$$\frac{\partial(c)}{\partial t} = -U\frac{\partial(c)}{\partial x} + D\frac{\partial^2 c}{\partial x^2} - \beta_{tot}\frac{\partial c_s}{\partial t} \qquad \text{2-7}$$

$$\frac{\partial c_s}{\partial t} = \frac{\partial}{\partial t}\int_0^t c(\tau)g^*(t-\tau)d\tau \qquad \text{2-8}$$

The probability function $g^*(t)$ for an exponential and a power law RTD are given by equations (2-10) y (2-11), respectively.

$$g^*(t) = \sigma e^{-\sigma t} \qquad \text{2-9}$$

$$g^*(t) = \frac{(k-2)}{\left(\sigma_{\max}^{k-2} - \sigma_{\min}^{k-2}\right)}\int_{\sigma \min}^{\sigma \max} \sigma^{k-2} e^{-\sigma t} d\sigma \qquad \text{2-10}$$

On the other hand, the structure of the ADZ model is indeed stochastic since the estimation of its parameters (*DF* and t_m, *DF* and τ, or t_m and τ) is data-based through the exploration of their parametric space [*Osuch et al.*, 2008], using methods as the GLUE (Generalized Likelihood Uncertainty Estimation) or the SCE-UA (Shuffled Complex Evolution-University of Arizona) [*González*; 2008]. Moreover, *Smith et. al* [2006] proposed an ADZ model framework which allow making probabilistic representations of *DF*, τ and the Steady State Gain (*SSG*), through normal, log-normal, and normal pdf, respectively, in order to account for errors corresponding with tracer data and the uncertainty associated with discharge changes within an analyzed stream reach.

2.4 STREAM FLOW CONSIDERATIONS

According with *Smith et al.* [2006], the major goal when defining the ADZ for a specific stream reach, is to relate one of the parameters of the model with discharge. Both timescales, t_m and τ, display clear relations with discharge which have received empirical treatment with reasonable results. The most common fitting curves are given by equations (2-12) y (2-13), where t is the characteristic temporal parameter, Q represents discharge, and k_1, k_2, k_3 y k_4 are constants.

$$t = k_1 + \frac{k_2}{Q} \qquad \text{2-11}$$

$$t = k_3 Q^{-k_4} \qquad 2\text{-}12$$

Using tracer data, *González* [2008] defined potential time-discharge relations for 9 stream reaches in Colombia, and showed the reliability of the obtained models for making new predictions. In his work, it is suggested to have at least one set of three data pairs (Q, t) spanning low, mean and high hydrological conditions. *Richardson and Carling* [2006] carried out 54 tracer experiments in a plane bedrock stream and they also highlighted the use of a potential fitting to represent the non-linearities on travel times. Besides, it is shown the low variability of these parameters at the higher stages because of the low interaction between the solute and the dead zones, which was defined in terms of a dispersive fraction near to cero.

In spite of the empirical form of the mentioned relationships, their application for simulation purposes is especially useful for cases where the stream flow regimes differ from those at which the fieldworks were performed. However, since not much is known regarding the assessment of the k_i constants based on morphological descriptors of a stream reach, tracer experiments must be implemented. In this way, to consider variations in the dispersion mechanisms driven by stream flow changes, *Camacho* [2000] posed the flow routing and solute transport scheme ADZ-MDLC, where the extension MDLC (Multilinear Discrete Lag-Cascade) is an aggregated flow routing model characterized by temporal parameters similar to those for the ADZ model. Topologically, a stream reach is represented by an element where the hydrograph is transported during a period τ_{fl} without suffering attenuation and then it enters to n identical linear reservoirs characterized by a routing coefficient K analogous to the resident time in the ADZ model.

One of the advantages of the MDLC model structure lies on the estimation of the model parameters based on hydraulic and geometric properties of a reach for any stream flow conditions. Equations (2-14) to (2-17) allow the estimation of parameters, where c_0 is the wave celerity, U_0 is the mean flow velocity, F_0 is the Froude number, L is the reach length, y_0 is the normal depth, and S_0 is the reach gradient.

$$m = \frac{c_0}{U_0}; \qquad 2\text{-}13 \qquad\qquad K = \frac{3}{2m}\left(1+(m-1)F_0^2\right)\left(\frac{y_0}{S_0 L}\right)\left(\frac{L}{mU_0}\right) \qquad 2\text{-}14$$

$$n = \frac{\frac{4m}{9}\left(1-(m-1)^2 F_0^2\right)}{\left(1+(m-1)F_0^2\right)^2\left(\frac{y_0}{S_0 L}\right)} \quad \text{2-15} \qquad \tau_{fl} = \left(\frac{L}{mU_0}\right)\left[1-\frac{\frac{2}{3}\left(1-(m-1)^2 F_0^2\right)}{\left(1+(m-1)F_0^2\right)}\right] \quad \text{2-16}$$

Additionally, given the reference discharge *Q*, the normal depth y_0 and the mean velocity U_0 can be estimated by:

$$Q = U_0 A = U_0\left(y_0 W\right) \quad \text{2-17}$$

$$Q \approx \left(\frac{1}{n_e} y_0^{2/3} S_0^{1/2}\right) y_0 W = \frac{1}{n_e} y_0^{5/3} S_0^{1/2} W \quad \text{2-18}$$

where *W* corresponds to the reach-average channel top width, and n_e is a reach-representative Manning coefficient, which takes into account the overall factors contributing to flow resistance.

Once the parameters *K*, *n* y τ_{fl} are obtained, the mean solute travel time can be estimated by equation 2-20 [*Camacho*, 2000; *González*, 2008]. Meanwhile, the first arrival time can be assessed with two different approaches using the *ADZ-MDLC*. First, τ is estimated as a function of the mean travel time and a predefined dispersive fraction, according with equation (2-21). This option is particularly useful in those systems where *DF* is likely to have wide variations when changing discharge.

$$t_m = m\left(nK + \tau_{fl}\right)\left(1+\beta\right) \quad \text{2-19}$$

$$\tau = t_m\left(1-DF\right) \quad \text{2-20}$$

Conversely, the second method is based on the assumption than *DF* has no significant variations, an annotation made in several classical studies. However, *González* [2008] showed that the dispersion fraction can vary up to a 22% within the same reach. For the latter assumption, the time τ can be estimated following equation (2-22).

$$\tau = m\tau_{fl}\left(1+\beta\right) \quad \text{2-21}$$

2.5 STREAM MORPHOLOGY

Several researches highlight as temporal storage processes both, those taking place in the hiporheic zone, which exchange solute mass with the main surface water body, and those due to surface dead zones related with the irregularities of stream geometry, whose effect has the major impact in the dispersion represented by models as the TS and the ADZ. According to this, dispersion could be assessed through the representation of the transport mechanisms in a geomorphological context, taking into account that a morphological and topological description of a drainage network can be made at different spatial scales through the growing availability of satellite information, digital elevation models, and algorithms for extracting drainage networks automatically.

Watersheds are characterized for featuring several stream morphologies which appear as a consequence of the interactions between topographic, geological, climatic, and even anthropogenic factors. Each morphological type has its own mechanisms to dissipate energy, transporting sediment and to assimilate pollution, and they also can be distinguished in terms of the habitat variety which favors the development of biota [*Stewardson*, 2005; *Thompson et al.*, 2008].

Stream patterns have been studied by geomophologists since 1900, when *Davis* [1899; en *Knighton*, 1998] classified streams as young, mature and old, under the assumption of a continuity of the erosion cycle. Although it is known that such cycle can be disturbed, more recent classification schemes remain being supportive of Davis' concept. However, unlike the temporal criteria considered in the early classification schemes, modern geomorphology is largely based on field observations in order to quantify the shapes and forms of featuring stream corridors in different river landscapes.

Different studies pose stream classification frameworks ranging from distinctions based on stream patterns and floodplain settings, to detailed descriptions including additional descriptors as sediment size distribution, bankfull stage signatures, slope gradient of both the main channel and the downstream valley, and local lithology.

The classification method developed by *Montgomery and Buffington* [1997] is based on the assumption that fluvial channels reach a geomorphologically stable morphology for a certain condition of sediment supply relative to transport capacity. The reach types considered in this scheme are the morphological classes known as *colluvial*, *cascade*, *step-pool*, *plane-bed*, *pool-riffle* and *dunne-ripple*, which according with the underlain hypothesis of the method, follow an ideal downstream sequence along a river corridor and feature different ranges of stream gradient. *Thomson et al.* [2008] also introduced watershed area and catchment lithology to classify Australian streams, since it led to more agreements with field observations than using only slope as a discriminatory variable.

Although reach-slope appears to be a significant variable to classify channels types [*Montgomery y Buffington*, 1997; *Wohl y Merritt*, 2008; *Thompson et al.*, 2008], a method only based on this variable leads to shortcomings at a regional scale analysis. In this regard, it has been shown that different stream morphologies can overlap reach-slope values, due to the additional forcing factors which determine a stream response.

Flores et al. [2006] evaluated the influence of the scale and hydroclimatic regime on the morphological structure of reaches of the drainage network, and presented a classification framework that includes, in addition to the longitudinal slope, the specific power as a discriminatory element between channels limited by supply and those limited by transport capacity, according to the classification method suggested by *Montgomery and Buffington* [1997]. In their work it was also found that a classification based only on the slope created difficulties in differentiating step-pool and cascade morphologies, which correspond to the general case of channels supply-limited, and plane-bed and pool-riffle morphologies, which correspond to channels limited by transport capacity but high sediment supply. For this reason, the method introduced the specific power, ω, as a metric of the transport capacity , which depends on the slope of the channel, S_0, the specific weight of water, γ, the discharge, Q, and the surface top width, W, (equation 2-23).

$$\omega = \frac{\gamma Q S_0}{W} \qquad \text{2-22}$$

For bankfull conditions, responsible for the bed forms and shapes of the channel of a reach [*Vianello & D'Agostino*, 2007], the connection between discharge and watershed area (A) is such that the latter is used as an alternative for estimation, since generally there is not enough information to estimate discharge. Several studies show a relationship of the form:

$$Q = aA^b \qquad \text{2-23}$$

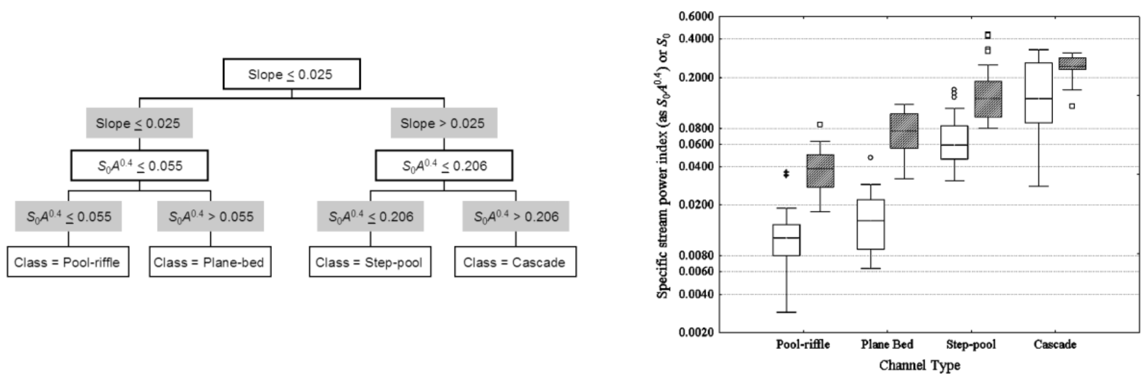
Where the exponent *b*, it was found to vary between 0.7 and 1. Similarly, a relationship with the same structure is already well known for the width, *W*.

$$W = cA^{d} \tag{2-24}$$

Considering that in equation (2-23) both the discharge, *Q*, and the width, *W*, exhibit scaling properties with the catchment area, *A*, the specific power can be written only in terms of area and slope in the form:

$$\omega \propto SA^{(b-d)} \tag{2-25}$$

where *b* and *d* denote the scaling exponents for discharge and width, respectively. *Flores et al.* [2006] used a value (*b-d*) of 0.4, defining an index of specific power as $S_0A^{0.4}$. Thus, the discriminatory tree shown in Figure 2-3a is proposed as a method of classification, which, as shown in Figure 2-3b, highlight more clearly the differences between supply-limited and transport-limited channels.



(a) (b)

Figure 2-3. a) Tree of morphological classification proposed by Flores et al. (2006), and b) effect of the differentiation of morphological types when the specific power of the current is introduced.

Although the classification schemes described have, to some extent, solved the problem of the spatial variability of the channel forms through their grouping into types or classes, its scope is limited to a macro view of the basin, and fails to quantify the internal variability of the channel geometry. Parallel lines of research continue to work on theories of hydraulic geometry, *HG*, beyond the pioneering work of *Leopold and Madock* [1953], motivated by the high variability found in the exponents that describe *HG* variables as the width of channel, *W*, the average depth, *H*, and the mean velocity, *U*, (equation 2-27).

$$W = a_1 Q^{b_1}; \quad H = a_2 Q^{b_2}; \quad U = a_3 Q^{b_3} \qquad \text{2-26}$$

where a_1, b_1, a_2, b_2, and a_3, b_3 are constants such that $a_1.a_2.a_3=1$ and $b_1+b_2+b_3=1$, to satisfy flow continuity.

One of the more recent works aimed at understanding the variability of the exponents of HG, was introduced by *Dodov and Foufoula-Georgiou* [2004a], who based on the premise that the at-station *HG* systematically depends on the scale (catchment area) and *downstream* HG depends on the recurrence of a characteristic flow, proposed a generalized model of HG from a statistical point of view based on theories of multi-scaling. Similarly*, Dodov and Foufoula-Georgiou* [2004b] showed that the variability of the HG with scale can be explained by the dependence on scale of the instability of fluvial processes.

Although channel-forming discharges or, in lack of that, the watershed area, have shown to be the most important variables in most studies of hydraulic geometry, they are not the only ones that explain such process. Other studies have explained the nonlinearity of the HG in light of the morphological variations occurring in a basin.

Vianello and D'Agostino [2007] show that changes from colluvial to alluvial morphologies, involves different mechanisms of channel width adjustment, as shown in Figure 2-4. These differences have been explained in terms of the relationship between transport capacity and sediment supply, which characterizes mountain rivers and armored stream beds [*Wohl*, 2004; *Vianello & D'Agostino*, 2007]. These types of rivers also have bank configurations more resistant than low-gradient channels subject to constant processes of lateral erosion.

In the same direction, *Parker et al.* [2007] point out that some of the aspects not considered in the traditional analysis of hydraulic geometry include the ability of a watershed to transport sediments downstream, *i.e.*, a relationship between transport capacity and sediment supply. By the term *quasi-universality* they showed that for rivers with gravel beds (D_{50}>25 mm) and well-formed alluvial plains, the exponents of *downstream HG* could be related to a shape factor of the cross section. Such factor depends on the relationship between Shields numbers at bankfull stage and the movement initiation stage, thereby, allowing the possibility of assessing changes in the transport/supply relationship.

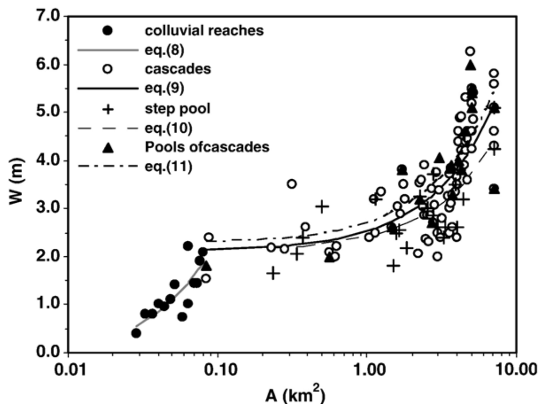


Figure 2-4. Variations of hydraulic geometry in relation to channel morphology. Taken from *Vianello and D'Agostino* [2007]

Scaling relations both physically as statistical based, used to represent HG have shown to be promissory in the conceptualization of hydrological models at the basin scale. However, at reach scale there are some geometric configurations that such scaling relations cannot represent, such as step-pool structures in high slope channels, and pool-riffle structures in low-gradient channels.

Classical works like the one by *Abrahams et al*. [1995] show that in the step-pool morphologies, the ratio $\left(\overline{H/L}\right)/S$ ranges between 1 and 2 as a natural strategy for maximizing energy dissipation. However, *Curran and Wilcock* [2005], using characteristic dimensions of step-pool morphologies from different studies, found that the theory of maximum energy dissipation is not met in all cases, and that a random process described by a Poisson distribution was better at representing bed forms. Interestingly, under both approaches, the average geometric features of such morphology also exhibit scaling relations. *Chin* [1999] showed, for different river reaches, that the ratio between the average length between pools, *L*, and the average width of the reach remains relatively invariant, suggesting that *L* also exhibits scaling relationships.

The same applies to pool-riffle morphologies in which it is found that the average riffle-width and average pool-width can be expressed independently as HG relationships [*Knighton*, 1998].

2.6 DISCUSSION

Throughout the different sections of this chapter, it is clear the close cause-effect relationship between the morphology of a stream and the dispersion mechanisms that take place along this. Among the clearest evidence is the uncertainty inherent to the *ADZ* and *TS* models when simulating scenarios over longer or shorter reaches than those for which the parameters are determined. Along with this evidence, it is important to mention the following:

- Unlike ADZ temporal model parameters, the dispersive fraction suggests no relation with discharge. However, *González* [2008] opened an important research path by showing that the higher dispersive fraction a greater percentage of the reach length is occupied by deep meso-scale units, common in morphological systems of the type step-pool and pool-riffle (Figure 2-6b).

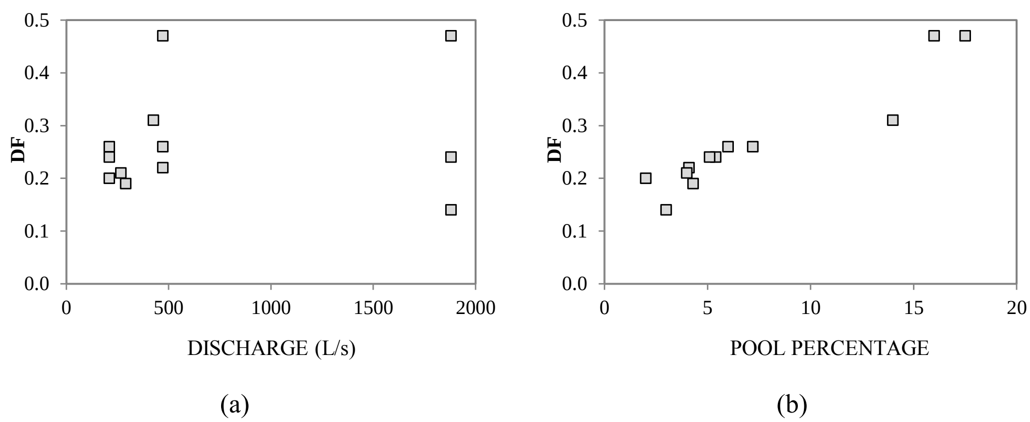


Figure 2-5. Proportionality between the dispersive fraction and the percentage of the reach length corresponding to pools. Data taken from *González* [2008]

- The mechanisms that trigger the flow of water through the hyporheic zone, recognized as one of the places for temporary storage of solutes in streams, are described by *Buffington and Tonina* [2009] in relation to different morphological attributes characteristic of fluvial systems ranked according to the classification system of *Montgomery and Buffington* [1998]. These attributes include the presence of rhythmic bed forms, the characteristics of the alluvial substrate, and the extension of the floodplains, in turn conditioned by the level of confinement. Figure 2-7 qualitatively represent the level of importance of the exchange flow rate and the spatial extent of the alluvium where such exchange takes places depending on the morphological configuration. In the graph, the assigned colors denote the level of understanding of the mechanisms of subsurface transport for each morphological type, where warm colors are the most studied.

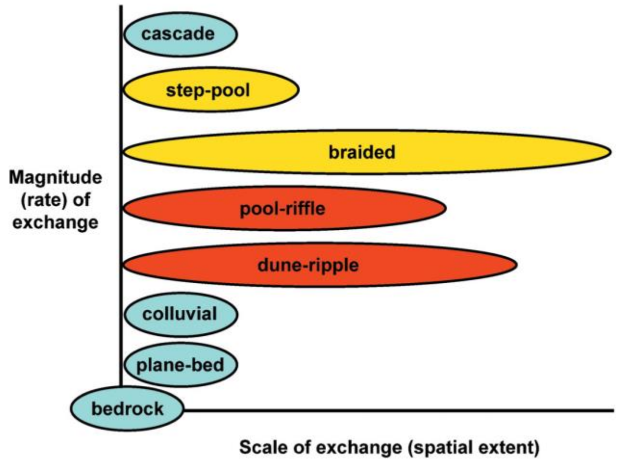


Figure 2-6. Level of importance of the exchange processes for surface and subsurface water flow according to channel morphology. Taken from *Buffington and Tonina* [2009]

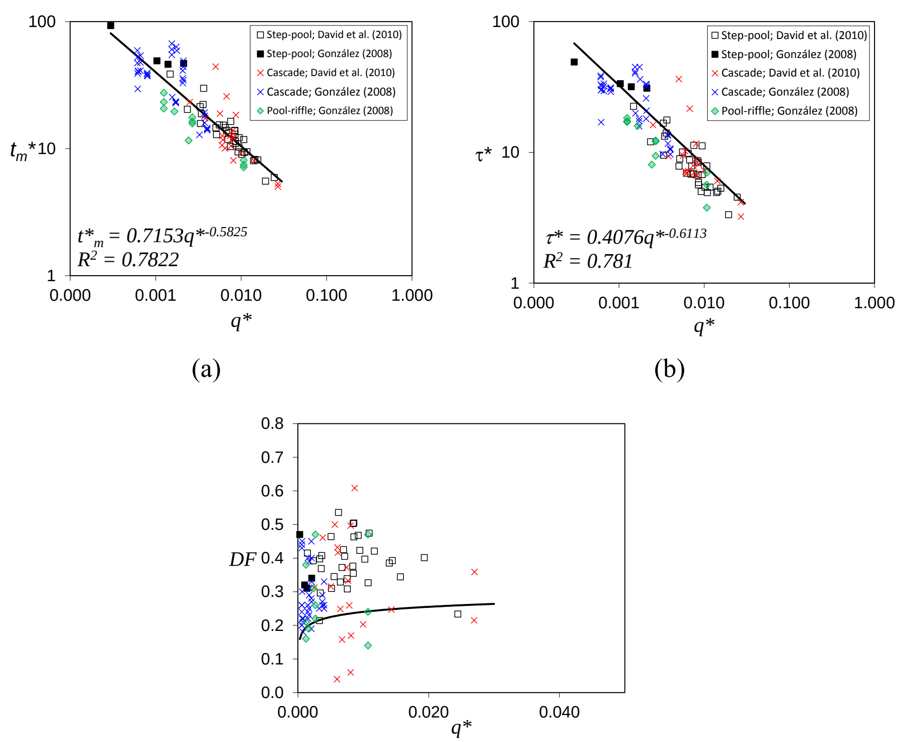
- Using the data bases *DataBogota* and *DataColorado* described in Chapter 1, a first exploration of the interrelationship between the temporal parameters of the ADZ model, t_m and τ, and morphological descriptors of the available stream reaches. For each flow condition, Q, and the corresponding times estimated in the studies that make up the databases, the dimensionless values t_m^*, τ^*, and q^* were estimated using equations (2-28), (2-29), and (2-30). The normalization strategy employs only the bankfull width, W_B, and the length of the reach, L_R, as morphological descriptors of each section. This is a limitation since the use of W_B in both dimensionless numbers can generate spurious correlation. However, the intention of this analysis is to motivate the hypothesis that under the appropriate choice of morphological descriptors in a reach of stream, the transport parameters of the ADZ model can be framed within a context of invariance of the processes or forcing factors that induce temporal storage, an aspect which will be further discussed in Chapter 3.

$$t_m^* = \frac{t_m}{L_R\left(gW_B\right)^{-0.5}} \qquad 2\text{-}27$$

$$\tau^* = \frac{\tau}{L_R\left(gW_B\right)^{-0.5}} \qquad 2\text{-}28$$

$$q^* = \frac{Q}{W_B\left(gW_B^3\right)^{0.5}} \quad \text{2-29}$$

Figure 2-8 shows the inferences made through the normalization process, where the results suggest that normalization techniques as the one used can be further analyzed in the understanding of invariance of the temporal parameters of the ADZ model. Preliminary inspections also highlight and confirm the high variability of the dispersive fraction, even for the same types or morphological classes.



(a) (b) (c)

Figure 2-7. Normalization of the solute transport parameters t_m^*, τ^*, and q^*, and their interrelation with the flow discharge and morphological descriptors of stream reaches.

Along with the evidence or motivations described above, the specific interest of this work is to address the transport of solutes in the basin scale, where the spatial and temporal variability of hydrological, geological, topographic and anthropogenic forcings lead to the genesis of various morphological configurations of streams.

In this context *Barrera et al.* [2002] developed a distributed model to assess concentrations of dissolved oxygen, biochemical oxygen demand, and total coliforms in the national drainage network, in order to provide a tool that could be used to prioritize infrastructure investment in wastewater treatment. This model represents the drainage network through different sections according to the classification of *Strahler* [1957], obtained based on a digital elevation model -DEM- with resolution of 342 m.

For the determination of dissolved oxygen and biochemical oxygen demand, the Streeter-Phelps model was implemented, along with schemes to represent first order decay of other water quality determinants. Subsequently, *Raciny and Camacho* [2003] made an extension using the model QUASAR-ADZ [*Lees et al.,* 1998] to consider hydraulic dispersion mechanisms not considered in the Streeter-Phelps model.

More recently, *Rojas* [2011] using the concept of assimilation factors, went further the contributions of predecessors studies and noted the importance of adequately represent dispersal mechanisms of solutes in the drainage network. This is particularly significant at the higher areas of the watersheds, where the dilution of pollutants in terms of the available discharge does not become dominant in the process of assimilation.

Yet, distributed models do not always give much importance to the shape of the channels, and their size is determined arbitrarily. Something similar happens with the roughness of the channel, which not only is often assumed stationary but uniform throughout the entire drainage network. However, flow routing, solute transport, sediment transport, among others, are discharge-dependent processes (nonlinear) on the reach scale and their variability is closely related to the type of channel morphology.

Similarly, the diversity of morphologies along the drainage network induces nonlinear processes in relation to the various responses that may be observed at the basin outlet according to the location of the upstream disturbances (rain, residual-water discharges, mass movements, etc.).