Determination of hydraulic features in Colombian rivers by tracer analysis

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Abstract: The importance of hydraulic, geomorphological and hydrodynamic conditions in rivers on watershed management cannot be overstated. However, in developing countries, sometimes the infrastructure is poor and these conditions are not managed properly. Therefore, this paper proposes a methodology to calculate flow, slope, Chezy's *C*, Manning's *n* and hydrodynamic conditions in rivers with a tracer. This methodology allows the calculation of these hydraulic and hydrodynamic characteristics for rivers that have velocity from 0.108 m/s to 1.93 m/s.

Keywords: hydraulica features; hydrodynamic; rivers; tracer.

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1 Introduction

The need to know the biotic factors, hydraulic, geomorphological and hydrodynamic features (Abbasova et al., 2017; Eslamian and Eslamian, 2017); in watersheds is a requirement for the integrated management and protection of water resources. These characteristics can affect social, economic, environmental, political and cultural factors in their communities (Thorne et al., 1997). However, in developing countries, those characteristics are not clearly known due to low, or non-existent, funds invested by authorities.

One of these methodologies is the use of a tracer in rivers for determining the hydraulic conditions. For example, Richardson and Carlin (2006) employed the ADZ model to investigate the bedrock channel hydraulics and their implications in solute transport. They calculated the flow, mean velocity, slope and Manning's *n*. Jiménez and Wohl (2013) used the ADZ model for the identification of the bankfull width, flow, step-to-step length, pool length and step height and generated a morphologic model for a river.

This paper proposes a new methodology based on the work carried out by Leopold and Maddock (1953), USGS scientists from the USA in the 1950s, 1960s and 1970s. They studied the physical conformation of the natural channels, especially the application of thermodynamics to the resolution of such problem. These general guidelines, adjustments and additions of a tracer were added. This paper is presented in five sections. Section 2 explains the theory of the method, Sections 3 presents the new approach, Section 4 shows the results and the calculations for five different rivers and Section 5 contains the conclusions.

2 Theory

2.1 Basic thermodynamic analysis of a test section of a channel

A test section of a natural channel is usually interpreted (in simplified form) as an isothermal closed system at a constant volume (which exchanges energy but not substances at a constant temperature). The potential energy (ΔU) comes from the gravitational force which forms an inclined plane, measured by the slope, S, with h representing the height and X the longitudinal variable:

$$S = \frac{\Delta h}{\Delta X} \tag{1}$$

This energy is immediately transformed into the energy of water movement (kinetic energy, ΔK). Because it is a totally irreversible process (without using energy into useful work), this energy is completely dissipated into heat, ΔQ, which is transmitted to the exterior:

$$\Delta U \Rightarrow \Delta K \Rightarrow \sum \Delta Q \tag{2}$$

In this case, in order to keep a constant temperature in the constant volume considered, the entropy produced by irreversible processes, $+\Delta Si$, must be expelled entirely to the exterior by entropy exchanges at the border, $-\Delta Se$. It is possible to infer, in accordance with the equations (3) and (4), that the entropy remains constant in the system, $S = cte$:

$$\Delta S = \Delta Si - \Delta Se \tag{3}$$

$$\Delta Si = \Delta Se \tag{4}$$

so:

$$\Delta S = 0 \tag{5}$$

In this instance, if an interpretation of maxima and minima is assumed, a null increase of entropy (as a first derivate of time) can be accepted as a maximum of this function. In addition to these results, it can be stated that in such totally irreversible systems, the injected energy (usable energy), U, tends to a minimum:

$$\Delta U = 0 \tag{6}$$

on the other hand, from statistical mechanics, it can be established that the entropy (as an external variable) is also related to the probability of discrete thermodynamic data treated, p_i, in the volumetric system, Ω, as follows, with k as the coefficient for adjustment of the dimensions involved:

$$S = k \times \sum_{\Omega} p_i \times \ln(p_i) \tag{7}$$

When the isothermal system (at constant volume) considers the entropy as a constant and maximum value, then $\Delta S = 0$. According to the definition of equation (7), this condition is achieved when all the probabilities in the system are equal:

$$p_1 = p_2 = p_3 = p_4 = ..p_i.. = p_\Omega \tag{8}$$

This implies that different thermodynamic variables associated with the description of the physical system in question have a uniform spatial distribution, as long as their probabilities of occurrence are similar. The practical conclusion is that if these variables are measured at different points in the system, these values will tend to a single value, if the conditions in which the analytical points of this study were developed are met.

This condition of the study of entropy (second law) on the spatial distribution of thermodynamic variables in a section of the channel will later help define certain laws in the application of dye tracers to the study of the great rivers. The uniformity of the channel parameters, derived from total irreversibility of the process in the natural channel considered, manifests along the length, area or volume of the section studied.

2.1.1 Dynamic equilibrium in natural channels

Dynamic equilibrium of channels is a principle that has been studied extensively and is established in terms of the geomorphological characteristics (slope, width, level and velocity). The flow rate is maintained over time in a natural channel, despite the occurrence of specific changes, large and small.

It is now necessary to analyse how the slope of a section, defined by the principle of uniformity described in the preceding paragraph, enables the occurrence of a dynamic equilibrium in the section. This means that the geomorphological stability of the channels is derived from the irreversible nature of the processes that occur there.

2.1.2 Formation of the slope of a channel according Leopold-Maddock

According to the development of Leopold and Maddock (1954), it can be said that the entropy production in a section of the channel can be expressed as:

$$\frac{\Delta S_i}{\Delta t} \approx \left(-\frac{Q}{h}\right)\frac{dh}{dX} \tag{9}$$

Here Q is the flow rate and the derived longitudinal of the depth, h, is the slope. The minus sign results from the fact that an increase in X implies a decrease of the slope, h. As it is considered that U is the potential energy by height, the slope is also the gradient of the loss of useful energy in the section. According to equation (7):

$$\frac{\Delta S_i}{\Delta t} \approx \frac{k}{\Delta t}\sum p \times \ln(p) \tag{10}$$

Therefore, it can be expressed that in the unit of time and considering the principle of uniformity in the section:

$$\left(\frac{-1}{h}\right)\frac{dh}{dX} \propto cons\tan te \tag{11}$$

Therefore:

$$\int \frac{dh}{h} = -aX + Co \tag{12}$$

so, if Co is properly interpreted:

$$h = he^{-aX} \tag{13}$$

and the slope, S, is

$$S = \frac{dh}{dX} = -be^{-aX} \tag{14}$$

This result is particularly interesting since this decaying exponential profile of the channel's height in the longitudinal direction corresponds to a multitude of experimental measurements, and perhaps more importantly, determines conclusively that the main element of the fluvial geomorphology is a consequence of the nature irreversible transformations of the potential energy in the section.

If the slope thus defined is a manifestation of the principle of uniformity (spatial equiprobability) then it will tend to have a certain value (not arbitrary), in the sense that the velocity of the current will adjust optimally to the restrictions of energy loss defined by equation (14). This restriction can be seen as certain stability in its value which manifests itself as a dynamic equilibrium of the channel. A direct consequence of this statement is that the channel will seek a special way to build its meanders, so that the resulting roughness would be compatible with the average velocity required by the slope of equilibrium.

Any natural or artificial change of geometric or dynamic parameters associated with the slope of equilibrium will try to be reversed by the channel through a reaction that tends to restore the initial condition. This is another interpretation of the well-known principle of Le Chatelier-Braun, but this time applied to hydraulics.

2.1.3 The formation of the channel

The slope first adjusts the average velocity, which at the same time defines the capacity of erosion in the streambed, consistent with the nature of the soil. The eroding effect will also depend on the area that it acts on.

The current causes erosion to the channel through which the liquid, dissolved and suspended substances will be transported. The process by which the formation of the channel is performed corresponds to how much of the banks and streambed area is subject to scour. These mechanisms are balanced to provide a channel that has an ideal configuration, as both extremes drive to the opposite action.

In areas where the majority of the friction is in the banks of the channel, there will be more scour that will expand the width. Moreover, in areas where the friction is the horizontal line, there will be higher scour that will expand the depth. Finally, the equilibrium profile is achieved in both cases. This implies that the average width and depth (and therefore the transverse section) essentially will depend on the adjustment made by the slope on the velocity and in the nature of the streambed material.

2.2 Model of natural channel in dynamic equilibrium proposed by Leopold and others

Based on the concepts developed in the previous sections, in the 1950s works began on analytical 'models' to relate the different variables of interest, especially hydraulics and geomorphological. This model is based on potential relationships between the width, W, the level (depth), h, the mean velocity of the channel, U, depending on the flow.

$$W = aQ^b \tag{15}$$

$$h = cQ^f \tag{16}$$

$$U = kQ^m \tag{17}$$

Based on the classical Chezy-Manning equation (Singh, 2003), it has the following development for the slope and the Manning roughness number:

$$U \approx h^{\frac{2}{3}} \times S^{\frac{1}{2}} / n \tag{18}$$

$$S \approx pQ^{z} \tag{19}$$

$$n \approx qQ^{y} \tag{20}$$

Here, the following identities are true:

$$1 = k \times c \times a \tag{21}$$

$$k \approx p^{\frac{1}{2}} \times c^{\frac{1}{6}} \times q^{-1} \tag{22}$$

and

$$\frac{1}{4} \approx \frac{z}{2} - y \tag{23}$$

Leopold led several USGS work groups, which by exploring many different channels, obtained approximate numerical values for potential exponents such as:

$$W \approx a\sqrt[4]{Q} \tag{24}$$

$$h \approx c\sqrt[2]{Q} \tag{25}$$

$$W \approx k\sqrt[3]{Q} \tag{26}$$

Normally, the exponent for the slope model is $z \approx 0.5$, according to Leopold's results. These developments are important because they can be compared with the results of velocity, slope and flow that result from the dye tracer data.

2.3 A state function for the evolution of a dye tracer cloud in natural channels

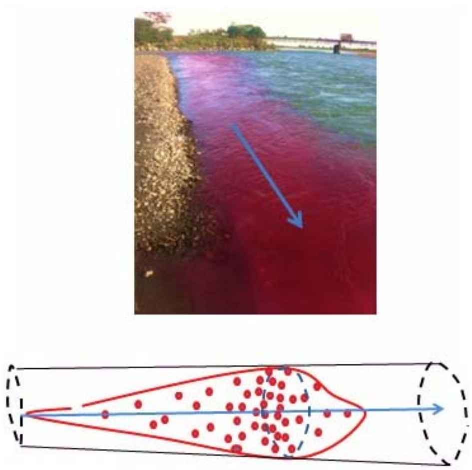
When a conservative solute (dye tracer) is poured into a natural flow, this substance intimately permeates with particles of water and 'simulates' the various movements of these. If there is turbulence, tracer particles will move alike, propitiating a large scattering effect (random expansion) and mixing of the solute. This is very interesting because the dye tracer participates with the flow's thermodynamic properties listed in the preceding paragraphs.

2.3.1 Definition of 'pipe flow' for the dye tracer's evolution

Pipe flow is simplistically defined by the ideal cylinder inside of the fluid containing the dye tracer particles from the injection point to the point where its evolution is measured. Although it is a vague definition, it conveys the concept. The dye tracer gradually extends lengthwise and it expands width-wise. In the top picture in Figure 1, the actual appearance of the 'ideal pipe' is shown. In the lower diagram, an ideal schematic is shown. As seen in the lower diagram, the outermost particles of the contamination cloud of the dye tracer are on the border of the pipe, although the spot is not uniform and does not cover some parts of the solute. It is a useful idealisation to put identifiable borders to the spread of solute in the flow (Constain-Aragón and Peña-Guzman, 2015).

The particles advance towards the central portion where the dotted blue circumference (transverse section of pipe) changes from having many highly concentrated particles in the central axis (blue arrow), to having many particles distributed almost uniformly in the transverse section. At that time, the dye tracer uniformly fills that area and it meets 'full mix' in such pipe. Naturally, a relatively long time passes from the injection of the tracer, so that a significant transverse dispersion can occur. The corresponding distance is called 'mixing length' for the pipe in question.

Figure 1 Pipe flow for contaminated cloud of the dye tracer and 'full mix' in transverse section (see online version for colours)



There are two situations that can occur depending on the flow to be measured, the mass of dye tracer that is poured and the distance travelled by the dye tracer. For small or narrow rivers, the 'pipe flow' covers the entire current, meaning that its borders are mistaken with both banks of the channel ('sufficient' pipe). Another is when there is a very broad or very large channel and the borders of the pipe are not able to make this coverage ('insufficient' pipe). Both situations are very significant in the practice of dye tracers and will be discussed later on, especially for the characterisation of large rivers as they require a different treatment.

2.3.2 State function of the evolution of the dye tracer

When a conservative solute is injected at a current, multiple irreversible processes take place in which the energy of formation (potential chemical energy) of the compound gradually diminishes, while its particles spread in the turbulence of the liquid. From the thermodynamic point of view, this energy evolves like gravitational energy represented in the height of the channel, per the analysis proposed by Leopold. However, it is convenient to explore alternative ways for a state function, closed integral, as shown in equation (27), to describe this phenomenon, which can be represented as 'Φ'.

$$\oint_c d\phi = 0 \tag{27}$$

In previous articles, some of the authors defined this function by relating two velocities. The velocities included the dispersion of the dye tracer, V_{disp}, of irreversible nature and measured by their displacement *random walk* and the other of advection, U, as an integral factor. Here, Δ and τ are characteristic parameters of the displacement and phase of the mono-dimensional Gaussian movement of the dye tracer's contamination cloud (Constaín et al., 2014; Constain-Aragón and Lemos, 2011):

$$\phi = \frac{V_{disp}}{U} = \frac{\left(\frac{\Delta}{\tau}\right)}{U} = \frac{\left(\frac{\sqrt{2E}}{\tau}\right)}{U} \tag{28}$$

This rearranged equation allows a second equation to be formed that is related to the average velocity of the channel, in addition to the classic expression of Chezy. Here, C is Chezy's resistance coefficient, R is the hydraulic radius and S is the slope:

$$U = \frac{1}{\phi}\sqrt{\frac{2E}{\tau}} \tag{29a}$$

$$U = C\sqrt{RS} \tag{29b}$$

Although Φ is assigned to describe the evolution of the dye tracer in a specific 'pipe flow', thanks to the principle of uniformity (equiprobability) valid for the thermodynamic variables of a channel in dynamic equilibrium, its expression can be used as a general descriptor for any other pipe in the same section of the channel (Constaín and Corredor, 2016).

This uniformity for Φ simply implies that in the section of the channel studied, the pattern of dispersion *will be the same* anywhere in the section regardless of the specific details related to concrete point of injection. This can be understood based on a further development for the state function as shown below, where M is the mass of conservative solute and α is a coefficient of 'force' in the creation of the dye tracer's concentration along its evolution.

$$\phi(t) \approx \frac{M}{Q\alpha 1.16} \times \frac{1}{\sqrt[3]{t}} \tag{30}$$

The ratio '$M / (Q\alpha)$' describes the absorption ability of the excess mass in the current and depends on the electrochemical nature of the fluid in the section considered. This ratio is certainly a thermodynamic characteristic of the current on the section studied, that is to say that its order of magnitude corresponds to what is meant by a uniform distribution (according to the equiprobability for Φ) at all points of the volume in question.

2.3.3 E as a function of time

The transport coefficient on the X axis can be cleared from equation 29(a) as follows, considering that the characteristic time $\tau \approx 0.215\ t$, according to Poisson's distribution, and assuming that the advection time is formed from the discrete count of the characteristic transportation times of the dye tracer particles (Constaín et al., 2014):

$$E = \frac{\phi^2 U^2 \times 0.215 \times t}{2} \tag{31}$$

A special feature of this definition is that unlike current approaches of dispersive transport, this coefficient is a function of time.

2.3.4 Definition of a modified Fick function and flow calculation

There is the classical Fick's equation for the conservative dye tracer's concentration, where A_h is the area of the transverse section (Fischer, 1967):

$$C \approx \frac{M}{A_h \sqrt{4\pi E t}} e^{-\frac{(X-Ut)^2}{4Et}} \tag{32}$$

If equation (31) is replaced with (32) it has a modified definition for Fick:

$$C \approx \frac{M}{Q\phi t 1.16} e^{-\frac{(X-Ut)^2}{2\times 0.215\times(\phi\times U\times t)^2}} \tag{33}$$

At the point of measurement, the flow can be calculated:

$$Q \approx \frac{M}{C_p \phi t 1.16} \tag{34}$$

This flow value is subject to a number of systematic errors that will be explained in the experimental development of Meta River.

2.4 Elder's equation as a function of time and its relation to the slope

Elder's equation for the longitudinal coefficient of dispersion was proposed in 1959, after being successfully verified in small channels in a laboratory. Here, h is the level (depth) of the channel and g is the acceleration of gravity (Elder, 1959). Its great advantage is that it links the basic characteristic of geomorphology, the slope, with the coefficient of dispersive transport, achieved with a compact expression (Constain-Aragon, 2014):

$$E = 5.93 h \sqrt{hgS} \tag{35}$$

Its development lies in the fact that the dispersion of particles of conservative solutes in turbulent flows is mainly due to the action of random separation of these particles caused by velocity distributions in a real fluid (shear). It is interesting to note that if equations (31) and (35) are equivalent, they define the same phenomenon, and then in reality Elder's definition *must also be a function of time*. It is then important to find a method to locate at which point on the curve of $E(t)$ this definition is found, $E(to)$, to be able to give coherent information about the slope.

2.5 Estimate function of optimal adjustment of Elder

To find the correct point where Elder's definition is located, it is necessary to combine equations 29(a) and 29(b) and define an estimation function F, so that when its value equals 5.93, it is located at the optimal point:

$$F = \phi^2 \tau \left(\frac{C^2}{2}\right)\left(\frac{h}{R}\right)\sqrt{\frac{S}{hg}} \tag{36}$$

The definition of the slope in this context is as follows:

$$S \approx \frac{E^2}{35.2h^3 g} \tag{37}$$

This procedure consists of trying different slope values compatible with E values within a curve and Φ values compatible with equation (10). Or of trying an approximate curve which contributes to the estimation function F, to be equal to 5.93. The resulting optimum value will help determine the optimum slope for the case considered. These procedural details will become clearer later on when the case study of the Meta River is analysed.

2.6 Characteristics of an extrapolation method to establish the geomorphology in large channels

The following section will analyse the various considerations leading to the establishment of an extrapolation method to measure large or wide channels. To perform this analysis, it is necessary focus on equation (31) in order to identify the sensitivity of this relationship in terms of the changes in variables on which it depends.

2.6.1 Effect of the state function, Φ, in the calculation of longitudinal dispersion coefficients

In order to be able to use equation (31) to determine the geomorphology, it should be extrapolated to the entire section of the channel that is considered. This is valid when the method is applied to large (or wide) channels; in the case of pipe flow of 'insufficient' dye tracers. In this important case, the dye tracer for practical reasons is applied in limited areas close to the banks, over short distances and thus for natural reasons, they are 'local' measurements. It is therefore necessary to establish a procedure for extrapolation to provide scientific information assignable to 'the entire' section of the channel.

A first aspect that emerges from the analysis of the equation is that as explained in detail, in the case of natural channels, the Φ parameter, state function, being by definition an isothermal thermodynamic potential, can be considered 'non-local' (valid for the entire section).

2.6.2 Nature and calculation of the transverse distribution of longitudinal velocities

There are some parameters that do not participate fully in the principle of uniformity, at least in a holistic manner. This is the case for velocity. Although average velocity is a parameter of thermodynamic nature (and thus able to extend to the entire section), their spatial distributions are not. It is well known from the beginning of the investigations on hydraulics that velocity is null along the banks, and that in the centre of the channel the velocity is at its highest. In the case of laminar region (blue), the distribution is more evident than in the turbulent region (purple), in which it is more uniform. In one, viscosity is more significant, and in the other turbulence is.

Using a normalised graph derived from the behaviour of the velocities in a rough pipe, it is possible to make an acceptable numerical approximation for large rivers.

2.6.3 Effect of this distribution in the calculation of longitudinal dispersion coefficients

In equation (31), the average velocity, *U*, and *Φ* function act quadratically, as opposed to advection time, *t*, which acts linearly. The state function has, as mentioned before, a general (non-local) character in the section studied, so that the measured value on the bank can be used in equation (31) and extrapolated to the entire section. The average velocity does not have this character because it is subject to a transverse distribution, described for the turbulent case. equation (31) should be calculated with a corrected mean velocity, $U_{efectiva}$, according to the relationship between the width of the cloud, *Wl*, and semi-width of the channel, W_{med}, multiplied by the maximum velocity, $U_{\max}$, in the centre of the channel. The closer the width of the cloud is to the semi-width of the cloud, function *k* will be closer to the unit (i.e., at maximum velocity in the centre of the channel):

$$U_{efective} \approx k\left(\frac{Wl}{W_{med}}\right) \times U_{\max} \tag{38}$$

Likewise, the effective velocity also affects the advection time, since it has the following effective definition for this time, considering the distance of the discharge location, *X*:

$$t_{efective} \approx \frac{X}{U_{efective}} \tag{39}$$

Therefore, the extrapolated definition for *E*(*t*) corresponding to the whole section of the channel, Figure 10, is:

$$E_{general} \approx \frac{\varphi^2 \times U_{efective}^2 \times 0.215 t_{efective}}{2} \tag{40}$$

This value helps start the trial and error method to set the correct point where F ≈ 5.93, and thus it is the optimal point where equation (37) will give a consistent value for the slope.

2.7 *Determination of the hydraulic area and average depth (level) for dye tracers*

From flow's data and velocity, it is possible determines the value of the area of the average transverse section of the section:

$$A_h \approx \frac{Q}{U} \tag{41}$$

If the average width of the channel is known in the section, the average level can be found:

$$h \approx \frac{A_h}{W} \tag{42}$$

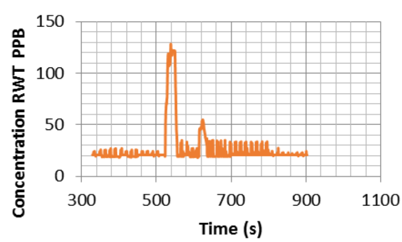
For the roughness of the channels, it is possible to apply the extended equation of Manning:

$$n \approx \frac{h^{\frac{1}{6}}}{C} \approx \frac{h^{\frac{1}{6}}}{\left(\frac{U}{\sqrt{hS}}\right)} \tag{43}$$

3 Methods and materials

For the development of this proposal, the following approximation for the development of the method arose.

Figure 2 Discharge tracer (RWT) on meta river and breakthrough curve (see online version for colours)



For the purpose of the evaluation of this methodology, five rivers from different regions and with different characteristics were selected. The historical information of flow (*Q*), level (*h*), and width (*W*), were supplied by the IDEAM (government entity that monitors water bodies).

The dye tracer used for the Meta, Negro, Las Ceibas and Guachiria Rivers was rhodamine WT (RWT) and NaCH (salt) was used for the La Vieja River. For measuring purposes, national technology Inirida deep flow (IDF) was used supported by FLUVIA, a methodology developed by Hydrocloro Tech Company from Bogota, based on real-time measurements with fluorometric dye tracers. Figure 2 shows the tracer discharge and the measurement curve of the concentration given by the IDF.

Table 1 Proposed methodology for the determination of hydrodynamic and geomorphological characteristics

Step		*Observation*
1	Data collection.	From responsible government entities, dye tracers and direct observation in the field.
2	Classification, refinement and processing of information.	Acquisition of average values for the data series, government entities and dye tracers.
3	Characterisation of the statistical nature of dye tracer data taken during the month to be evaluated.	For dye tracer data, average values obtained during the transition of the month (period of measurement) were analysed in order to compare them with average annual values and thus see how representative the tracer data is.
4	Calculation of specific coefficients in the model of Leopold-Maddock for the Meta River.	Mean annual values obtained from secondary sources and values of direct observation in the field are used.
5	Characterisation of the transverse distribution of longitudinal velocities in the desired flow.	From the radial diagram for a turbulent pipe, general theoretical distribution used for the desired flow in an approximate form is adjusted. Maximum velocity data is used (centre of channel) measured with dye tracers.
6	Analysis of the dye tracer data for velocity, flow and peak concentration of the tracer.	The nature and extent of systematic errors are determined which appear when measurements are made on the banks. Equations of adjustments and corrections are determined based on the dynamics of dye tracer's cloud.
7	Resolution of data provided by government entities and dye tracer for Hydraulic.	Reliable data on mean velocity and flow for the channel. The errors involved are established and their marking.
8	Resolution of dispersive transport data and its consistency with hydraulic data.	A consistent set of extrapolated equations which can be apply to the different fields are provided. The errors involved are established and their marking.
9	Application of equations of geomorphology by Leopold and their completion and adjustment to the procedures of Elder (tracers) *E*.	The validation of the different results is done against dye tracer data. The errors involved are established and their marking.
10	Refinement of equations and adjustment of consistency for the purpose of predicting events.	Examples and applications on desired flows are performed.

Table 2 Evaluated rivers and their hydraulic characteristics

River	*Mean monthly flow (m^3/s)*	*Mean monthly height (m)*	*Mean monthly width (m)*	*Number of sections evaluated*	*Measurement date*
Meta	3,271	4.8	640	2	April–May 2016
Negro	31.3	Not reported	40	1	January 2015
Guachiría	Not reported	Not reported	84*	1	February–March 2016
La Vieja	Not reported	0.096*	1.80*	1	April 2015
Ceibas	4.92	Not reported	84	1	June 2014

Note: *Measured in field.

4 Calculations

Calculations obtained for Meta River are shown below. Using mean values to calculate the coefficients for equations (24), (25) and (26), the following values were obtained:

$$W \approx 85.5\sqrt[4]{Q}; h = 0.0874\sqrt[2]{Q} \text{ and } U = 0.134\sqrt[3]{Q} \tag{44}$$

According to the dye tracer measurements in the centre of the river, it was estimated that the maximum velocity was 2.1 m/s, For the calculation of mean velocity, the standard graphics from the theoretical pipe data that connects the transverse distribution of velocities and the ratio r/R were used, which can provide an acceptable idea of the conditions in natural channels, if sufficient turbulence exists. The mean velocity obtained was 1.93m/s. This value of mean velocity throughout the flow had a relative error that was less than 2% comparing it to the value found in the model presented by Leopold-Maddock (1.99 m/s), showing a very accurate approximation.

Table 3 Experimental data for four pouring in Meta River

Discharge	*Mean velocity U_m(m/s)*	*Travel time t (s)*	*State function Φ*	*Longitudinal dispersion coefficients E (m^2/s)*
Casanare X = 37 m M = 52 g	1.93	19.4	0.70	3.77
Casanare 1 X = 200 m M = 50 g	1.93	66.4	0.28	3.27
Vichada X = 43 m M = 100 g	1.93	430	0.33	0.97
Vichada 1 X = 100 m M = 80 g	1.93	400	0.156	0.506

Once the value of mean velocity was obtained, the next step was to calculate the longitudinal dispersion coefficients extrapolated to the entire channel. At this point, it was important to consider the error due to the advective action, which is distorted (in excess) transport time, as it is linked in the bank to velocities deflated by default. Therefore, it was necessary use the correct value of time, assuming that the longitudinal dispersion coefficient applied to 'the entire' current. This forced the use of the 'mean velocity' of the current applied to specific distances from each discharge location. Using equations (39) and (40), the following values were found.

The basic geomorphology data in our case was the slope (*S*). It was calculated essentially with the dye tracers and then was generalised by the model of Leopold-Maddock. The extended formula from Elder [equation (37)] was used, but in the centre of the channel at mean velocity. For verification, the calculation of Chezy's expanded coefficient *C*, was necessary, along with the estimation function *F*. For each measurement point, the following was found.

Table 4 Slope values for four pouring in River Meta

Discharge	*E* (m^2/s)	*S*	*C* ($m^{1/2}/s$)	*F*
Casanare	3.77	0.00037	45.7	5.936
Casanare 1	3.27	0.00028	52.6	5.903
Vichada	0.97	0.00003	176.2	5.906
Vichada 1	0.506(*)	0.000007*	326	5.68
Mean	2.67	0.00023	N.A.	5.92

Note: *It is a gross mistake and it is not taken into account.

From the above, it is possible conclude:

1 The value of the measurement made in Vichada 1 was affected by the gross error with respect to the others. For a more homogeneous series, it was excluded from the calculation of the mean.

2 The series obtained was very homogeneous, since the estimation function in three of the four values was close to 5.93. Thus, the average value for the slope can be accepted as correct, $S \approx 0.00023 = 0.023\%$. This value was consistent with the classical data of geomorphology for lowland rivers in the development stage of the Meta River.

3 From this value, you can apply the second equation from Elder (for transverse diffusion coefficient), so that the development of the 'width of the dye tracer cloud' can be estimated with confidence.

Finally, equation (45) was used to calculate the channel roughness:

$$n \approx \frac{h^{\frac{1}{6}}}{C} \approx \frac{h^{\frac{1}{6}}}{\left(\frac{U}{\sqrt{hS}}\right)} \approx \frac{4.8^{\frac{1}{6}}}{\frac{1.93}{\sqrt{4.8 \times 0.00023}}} \approx 0.0224 \tag{45}$$

4.1 Results

The application and results of the methodology described above in different Colombian rivers are shown in Table 5.

Table 5 The results of other rivers

River	*X** *(m)*	*M* *(g)*	*Um* *(m/s)*	*t* *(s)*	*Φ*	*E* *(m^2/s)*	*S*	*F*	*C* *($m^{1/2}/s$)*	*n*
Negro	365	42	0.48	765	0.27	1.38	0.00071	5.95	17.4	0.058
Guachiría	100	40	0.79	127	0.3	0.76	0.00023	6.0	37.5	0.021
Las Ceibas	2.8	2	0.43	9.3	2.1	0.003	0.0079	5.9	6.3	0.145
La vieja	20	136	0.108	940	0.80	0.176	0.150	5.9	0.92	0.7

Note: *Distance from the discharge point.

5 Conclusions

It is important to note that the simplicity and ease of application of this method can be very useful in certain cases where the data is too difficult to obtain by other methods.

The values of the evaluated rivers are close to the expected ones under the current methodology. For example, observations by specialists in regards to the rough characteristics of these types of flows, and the values calculated by equations of Leopold-Maddock.

The results obtained in the Colombian rivers that were assessed, validate the use of dye tracers and show its usefulness when conducting measuring tasks in rivers, if the transport coefficient is a function of time applied to Elder's equation.

In developing countries where hydraulic information is scarce or not available, this methodology can be used, despite of the lack of historical references. It is able to provide a complete and consistent picture of these data for the measured sections.

Additionally, there are simple equations to determine hydrodynamic parameters that are of great importance for the application of software that allows the simulation of water quality.

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