



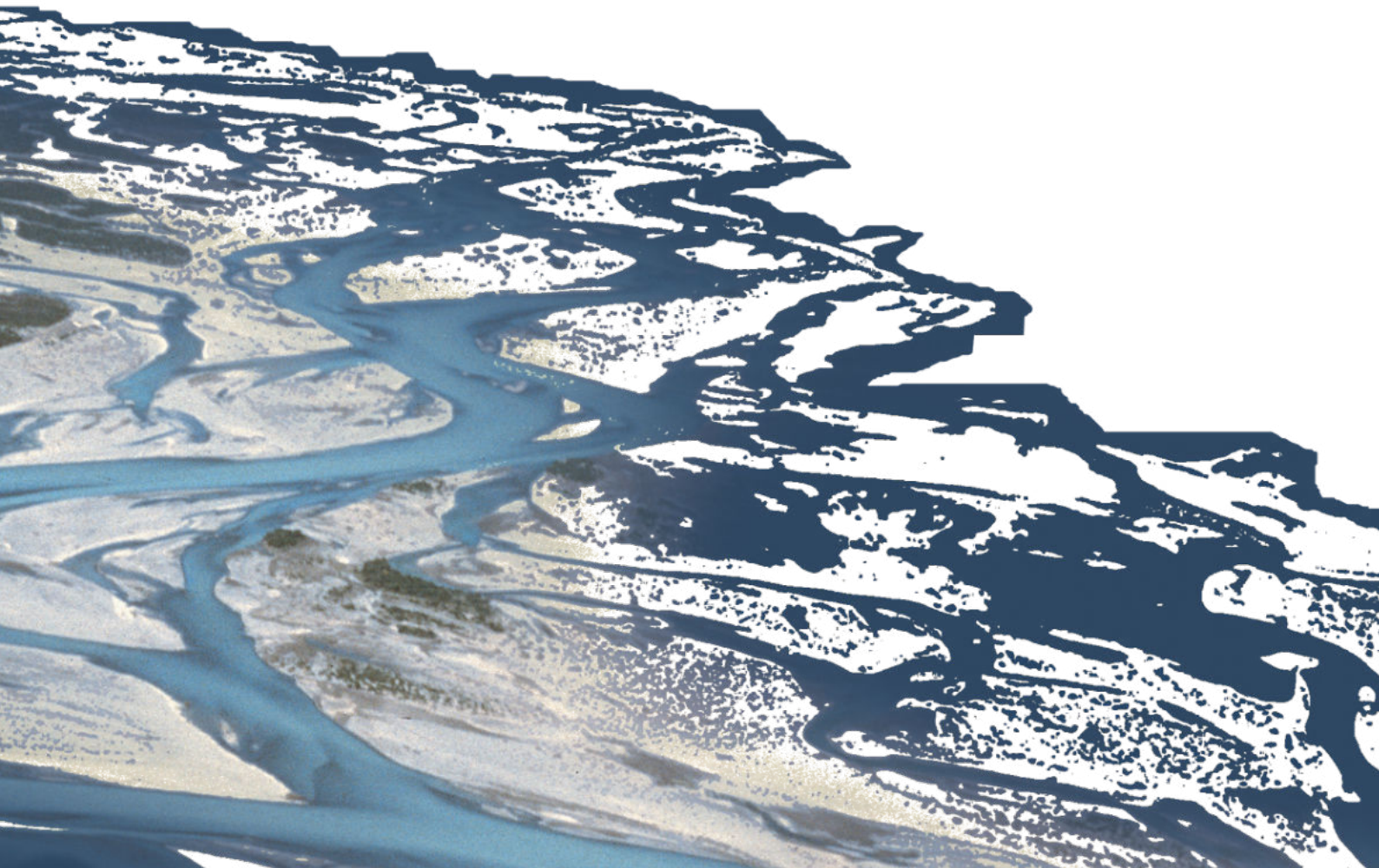
**BASEMENT**

**BASIC SIMULATION ENVIRONMENT  
FOR MODELLING OF ENVIRONMENTAL  
FLOWS AND NATURAL HAZARDS**

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# **SYSTEM MANUALS**

**VERSION 4.0.2  
OCTOBER 2023**





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# Preamble

**VERSION 4.0.2**

*October 2023*

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*For System Manuals:*

Vetsch D., Siviglia A., Bacigaluppi P., Brown, A., Bürgler M., Caponi F., Conde D., Gerke E., Halso M., Kammerer S., Koch A., Peter S., Vanzo D., Vonwiller L., Weberndorfer M. 2023. System Manuals of BASEMENT, Version 4.0.2. Laboratory of Hydraulics, Glaciology and Hydrology (VAW). ETH Zurich. Available from <https://www.basement.ethz.ch>. [date of access].

*For Website:*

BASEMENT – Basic Simulation Environment for Computation of Environmental Flow and Natural Hazard Simulation, 2023. <https://www.basement.ethz.ch>

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**BASIC SIMULATION ENVIRONMENT  
FOR MODELLING OF ENVIRONMENTAL  
FLOWS AND NATURAL HAZARDS**

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# **REFERENCE MANUAL BASEHPC**

**VERSION 4.0.2  
OCTOBER 2023**



**BASEMENT**



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# Mathematical Models

## 1.1 Hydrodynamics

### 1.1.1 Introduction

Mathematical models of the so-called *shallow water* type govern a wide variety of physical phenomena. Especially the one-dimensional (1D) de Saint-Venant equations (SVE) or two-dimensional (2D) shallow water equations (SWE) are of practical interest with regard to water flows with a free surface under the influence of gravity. Applications of the models include e.g.:

- River hydrodynamics
- Propagation of flood waves
- Dam break waves
- Flooding and inundation
- Ecological assessment based on flow quantities

The 2D SWE are based on the following set of hypotheses:

- the water is assumed to be incompressible; i.e. the water density  $\rho$  is constant
- the vertical acceleration of the water particles are assumed to be small compared to the longitudinal component of the acceleration. As a consequence the pressure distribution is hydrostatic;
- the bottom slope is small enough for the longitudinal coordinate to coincide with the horizontal axis;
- the flow regime is turbulent. As a consequence the head loss, mainly due to friction against the bottom, is proportional to the square of the flow velocity.

### 1.1.2 Governing Equations

The governing equations are obtained under shallow water conditions imposing mass conservation for the fluid and solid phases and the momentum principle to a flow in an open channel with a fixed bottom.

Introducing a Cartesian reference system  $(x, y, z)$  in which the  $z$  axis is vertical and the  $x - y$  plane is horizontal with respect to gravity  $g$ , the system of governing equations can be written as

$$\begin{cases} \frac{\partial h}{\partial t} + \frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} = S_h \\ \frac{\partial q_x}{\partial t} + \frac{\partial}{\partial x} \left( \frac{q_x^2}{h} + \frac{1}{2}gh^2 \right) + \frac{\partial}{\partial y} \left( \frac{q_x q_y}{h} \right) + gh(S_{bx} + S_{fx}) + T_x = 0 \\ \frac{\partial q_y}{\partial t} + \frac{\partial}{\partial x} \left( \frac{q_y q_x}{h} \right) + \frac{\partial}{\partial y} \left( \frac{q_y^2}{h} + \frac{1}{2}gh^2 \right) + gh(S_{by} + S_{fy}) + T_y = 0, \end{cases} \quad (1.1)$$

where:

$h$	$[m]$	water depth
$g$	$[m/s^2]$	gravity acceleration
$u$ ( $v$ )	$[m/s]$	depth averaged velocity in $x$ ( $y$ ) direction
$q_x$ ( $q_y$ )	$[m^2/s]$	discharge per unit width in $x$ ( $y$ ) direction
$S_h$	$[m/s]$	lateral inflow/outflow discharge per unit width
$S_{fx}$ ( $S_{fy}$ )	$[-]$	friction terms in $x$ ( $y$ ) direction
$T_x$ ( $T_y$ )	$[-]$	turbulent terms in $x$ ( $y$ ) direction .

The bed slope source terms

$$S_{bx}, S_{by}$$

are evaluated as follows:

$$S_{bx} = -\frac{\partial z_B}{\partial x} \quad ; \quad S_{by} = -\frac{\partial z_B}{\partial y} \quad (1.2)$$

### 1.1.3 Closure Relations

In order to solve system (eq. 1.1) we need to specify the closure relations for the friction terms  $S_{fx}, S_{fy}$  and the value of lateral inflow/outflow discharge per unit width  $S_h$ .

#### 1.1.3.1 Friction Terms

The governing equations (eq. 1.1) have been derived under the hypothesis (H3) of turbulent flow, hence the friction term  $S_f$  can be assumed proportional to the square of the depth-averaged velocity and can be written as:

$$S_{fx} = \frac{u|\vec{u}|}{gc_f^2 h} \quad ; \quad S_{fy} = \frac{v|\vec{u}|}{gc_f^2 h} \quad (1.3)$$

where  $g$  is the gravity acceleration,  $u$  and  $v$  are the depth averaged velocities in  $x$  and  $y$  direction,  $|\vec{u}| = \sqrt{u^2 + v^2}$  is the magnitude of the velocity vector and  $c_f$  is the dimensionless friction coefficient.

Several formulae are available for the dimensionless friction coefficient  $c_f$ . Here it is quantified using both a power or a logarithmic for which are described in the next sections.

### 1.1.3.1.1 Power Law

The Manning-Strickler power law is widely used in practice and it requires that either the Strickler's  $k_{str}$  [ $m^{1/3}/s$ ] or the Manning's  $n$  coefficients ( $k_{str} = n^{-1}$ ) is specified.

In this case the dimensionless friction coefficient  $c_f$  is calculated as

$$c_f = \frac{k_{str} h^{1/6}}{\sqrt{g}} \quad (1.4)$$

### 1.1.3.1.2 Logarithmic Law

The following approaches are implemented to determine the friction coefficient  $c_f$ :

Chézy:

$$\begin{aligned} c_f &= 5.75 \log \left( 12 \frac{R}{K_s} \right) & \text{for } R > K_s \\ c_f &= 5.75 \log (12) & \text{for } R < K_s, \end{aligned} \quad (1.5)$$

where  $K_s$  [ $m$ ] is the bed roughness height which is commonly taken to be proportional to a representative sediment size  $d_x$ . For rivers,  $K_s$  can be assumed  $K_s = n_k d_{90}$  where  $n_k = 2 \div 3$ .

Bezzola:

In this closure relation, proposed by Bezzola (2002),  $c_f$  is given as a function of the roughness sublayer height  $y_R$  [ $m$ ] (usually for rivers  $y_R \approx 1.0 d_{90}$  is a good approximation). This approach is also valid for small values of the relative submergence  $h/y_r$  Bezzola (2002).

$$\begin{cases} c_f = 2.5 \sqrt{1 - \frac{y_R}{h}} \ln \left( 10.9 \frac{R}{y_R} \right), & \text{for } \frac{h}{y_R} > 2 \\ c_f = 1.25 \sqrt{\frac{h}{y_R}} \ln \left( 10.9 \frac{R}{y_R} \right), & \text{for } 0.5 \leq \frac{h}{y_R} \leq 2 \\ c_f = 1.5, & \text{for } \frac{h}{y_R} < 0.5 \end{cases} \quad (1.6)$$

### 1.1.3.2 Turbulent Terms

The turbulent and viscous stresses are considered under the eddy viscosity model, following the Boussinesq hypothesis, and can be written as

$$T_{x_i} = \frac{\partial}{\partial x_j} \left( (\nu + \nu_t) \frac{\partial}{\partial x_j} h u_{x_i} \right)$$

where  $x_i$  stands for the  $x$  or  $y$  direction,  $\nu$  and  $\nu_t$  are the laminar and turbulent kinematic viscosities, respectively,  $u_{x_i}$  is the flow velocity and  $T_{x_i}$  are the commonly denominated Reynolds stresses. The turbulent viscosity is the defining parameter when employing this

type of turbulence closure, with several closures being available in fluid mechanics literature. For SWE systems, the most common options for computing the eddy viscosity are the constant, mixing-length and  $\kappa$ - $\epsilon$  models.

### 1.1.3.2.1 Mixing-length model

The mixing-length model was derived by Ludwig Prandtl and is based on a characteristic length concept where a fluid retains some of its original characteristics (namely turbulence) before dispersing them into the surrounding fluid. In this model, the turbulent viscosity is locally derived by a length scale, the strain rate and the friction velocity as follows:

$$\nu_t = l_s^2 \sqrt{2S_{ij}S_{ij} + \frac{2.34\kappa}{u_f}}$$

where  $l_s$  is the characteristic length scale,  $S_{ij} = (\partial_{x_j}u_i + \partial_{x_i}u_j)/2$  is the strain rate,  $\kappa$  is the von Karman constant and  $u_f = c_f|\bar{u}|$  is the friction velocity. The length scale can be defined as a constant or by using  $l_s = 0.267\kappa h$  (ref).

### 1.1.3.2.2 $\kappa$ - $\epsilon$ model

The  $\kappa$ - $\epsilon$  model is one of the most widely adopted models in hydraulics and general fluid mechanics. The turbulent kinetic energy (TKE)  $\kappa$  is introduced by means of an additional conservation equation, given by:

$$\frac{\partial(kh)}{\partial t} + \frac{\partial(khu_i)}{\partial x_i} - \frac{\partial}{\partial x_j} \frac{\nu_t}{\sigma_k} \frac{\partial kh}{\partial x_j} = hP_k - h\epsilon$$

where  $P_k$  is the TKE production rate and  $\epsilon$  is its dissipation rate. The  $P_k$  term is locally defined, whereas the  $\epsilon$  term is solved through another conservation equation, as:

$$\frac{\partial(\epsilon h)}{\partial t} + \frac{\partial(\epsilon hu_i)}{\partial x_i} - \frac{\partial}{\partial x_j} \frac{\nu_t}{\sigma_\epsilon} \frac{\partial \epsilon h}{\partial x_j} = hS_\epsilon$$

where  $S_\epsilon$  is an aggregate net source of  $\epsilon$ . In the Standard  $\kappa$ - $\epsilon$  model, the source terms for both  $\kappa$  and  $\epsilon$  are defined as:

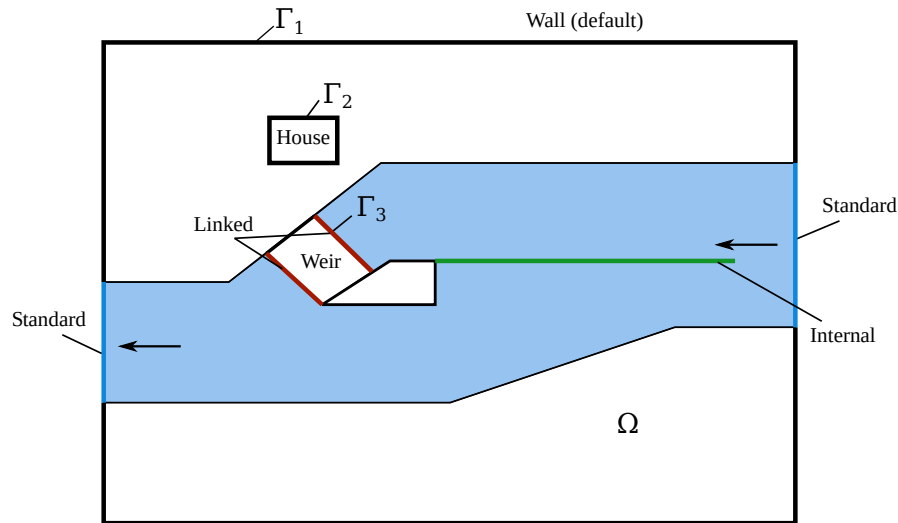
$$P_k = 2\nu_t S_{ij} S_{ij} + \frac{u_f^3}{c_f h}$$

$$S_\epsilon = 2c_{\epsilon_1} \frac{\epsilon}{\kappa} \nu_t S_{ij} S_{ij} + 3.6 \frac{c_{\epsilon_2} \sqrt{c_\mu} u_f^4}{h^2 c_f^{3/4}} - c_{\epsilon_2} \frac{\epsilon^2}{\kappa}$$

and the constants of the standard model are  $c_\mu = 0.09$ ,  $\sigma_k = 1.00$ ,  $\sigma_\epsilon = 1.30$ ,  $c_{\epsilon_1} = 1.44$  and  $c_{\epsilon_2} = 1.92$ . The turbulent viscosity is finally obtained, locally, as

$$\nu_t = c_\mu \frac{\kappa^2}{\epsilon}$$





**Figure 1.1** Modeling domain and types of boundary conditions available. The flow is from right to left and a side weir (green line) divides the channel into a lower and an upper channel through the weir. **Standard** or **linked** boundary conditions must be provided at  $\Gamma_1$ ,  $\Gamma_2$  and  $\Gamma_3$  while **internal** boundary conditions can be specified in any place within  $\Omega$

### 1.1.3.3 Lateral Inflow/Outflow

$S_h$  is used to represent additional sources of water like rainfall and springs or water abstraction (sink) and are allocated on a set of elements defined by regions. The external source can be specified as total discharge [ $m^3/s$ ] or distributed over time [ $mm/h$ ]. Different approaches are used to manage the behaviour of the external sources:

- Exact: The specified water volume is added or extracted (non conservative)
- Available: The specified water volume to extract is limited by the available water volume in the elements (conservative)
- Infinity: All available water will be abstracted (conservative)

Addition of water always follows the “Exact” behaviour as there is no upper limit. The abstraction of water could also follow the “Exact” behaviour but the simulation might end abruptly if the available water volume is smaller than the volume prescribed. Therefore, the “Available” behaviour aims to avoid this situation. The “Infinity” behaviour abstracts all available water volume.

### 1.1.4 Boundary Conditions

After the specification of the *closure relations* there are now three equations and three unknowns, namely  $h$ ,  $q_x$  and  $q_y$ . In principle, given initial and boundary conditions, one should be able to solve system (eq. 1.1) for  $h$ ,  $q_x$  and  $q_y$  as functions of space  $x$ ,  $y$ , and time  $t$ . Given the modeling domain described in Figure 1.1, boundary conditions are required at the domain boundary  $\Gamma$  and optionally can be specified within the interior domain  $\Omega$ .

Therefore, three different types of boundary conditions can be defined:

- Standard boundary conditions: located at the domain boundary  $\Gamma_i$
- Linked boundary conditions: located at the domain boundary  $\Gamma_i$  or inside the domain  $\Omega$
- Internal boundary conditions: located inside the domain  $\Omega$

*Standard boundaries* (at  $\Gamma$ ) represent the limits of the computational domain possibly including also buildings, weirs or structures for water intake (see Figure 1.1).

#### 1.1.4.1 Standard Boundary Conditions

At the standard boundaries, two different types of boundary conditions can be specified: wall or flow boundaries. Flow boundary conditions allow the flow to enter or leave the domain while wall boundary conditions express no mass flux over the boundary. By default, the external boundaries of the domain are set as wall boundaries.

##### 1.1.4.1.1 Wall Boundaries

The *Wall* or *reflective* boundary consider the boundary at  $\Gamma_i$  and suppose it physically consists of a fixed, reflective impermeable wall. Then the physical situation is modelled imposing that:

$$\rho \vec{u} \cdot \vec{n} = 0 \quad ; \quad \frac{\partial \vec{u}}{\partial \vec{n}} = 0 \quad (1.7)$$

Where  $\vec{n}$  is the outward directed unit vector perpendicular to the wall and  $\vec{u} = (u, v)^T$  is the velocity vector. The static pressure is assumed to be zero.

##### 1.1.4.1.2 Flow Boundaries

The *Flow* boundary conditions are defined as *inflow* if they let water entering or as *outflow* if they let water leaving the domain. Flow boundaries are further distinguished into *Standard* and *Linked*. The former are applied on the boundary domain  $\Gamma$ , while the latter establish a *link* between two portions of the domain.

##### Standard

*Inflow boundaries:*

This boundary requires the specification of a value for the total volume discharge  $Q$ , [ $m^3/s$ ], which is then divided by the length of the boundary  $\Gamma$  and projected orthogonally to the boundary to obtain the values of  $q_x$  and  $q_y$ . In case of supercritical flow the following possibilities to specify the value of the water depth  $h$  are possible:

- *uniform in*:  $h$  is calculated assuming that local uniform flow conditions. The calculation proceeds as follows:

$$h = \sqrt[3]{\frac{(Q/b)^2}{gc_f^2 s}} \quad (1.8)$$

where  $c_f$  is the Chézy coefficient,  $b$  is the entire length of the boundary  $\Gamma$  and  $s$  is the value of the local bed slope that must be specified.

- *froude\_in*: In this case the flow depth  $h$  is calculated as follows:

$$h = \sqrt[3]{\frac{(Q/b)^2}{gFr^2}} \quad (1.9)$$

where  $b$  is the entire length of the boundary and  $Fr$  is the value of the local Froude number that must be specified

- *zhydrograph*: The water surface elevation (wse) at the boundary must be specified by the user. The depth is calculated as:

$$h = wse - z_B \quad (1.10)$$

where  $z_B$  is the bottom elevation at the boundary. The flow velocity at the boundary is set to zero.

#### Outflow boundaries:

At the outflow boundaries a value for the water depth  $h$  must be specified. These are the possible options:

- *uniform\_out*: the water depth  $h$  is calculated using equation (eq. 1.8) specifying a value for the total discharge  $Q$  and a local bed slope  $s$ . Uniform flow is calculated based on given slope and cell state at boundary (eq. 1.8).
- *weir\_out\_constant* and *weir\_out\_dynamic*: These boundary conditions establish a relation between the approaching discharge  $q$  constant and the water depth using the Poleni weir formula:

$$q = \frac{2}{3}\mu\sqrt{2g(h_{up} - w)^3} \quad (1.11)$$

where  $h_{up}$  is the water depth of the approaching flow and  $w$  is the weir elevation. The Poleni factor  $\mu$  can be either set as constant ( $\mu = 0.75$  by default) or dynamically evaluated as:

$$\mu = \frac{0.611}{a} \frac{0.75}{b} \frac{h_{up} - z_w}{w} \quad (1.12)$$

where  $a$  and  $b$  must be specified by the user in the case of *weir\_out\_dynamic* (default values are  $a = 0.611$  and  $b = 0.075$ ).

- *hqrelation\_out*: The discharge is determined as a function of the water surface elevation, thus a stage-discharge-relation has to be specified.
- *zhydrograph*: Sets a fixed water surface elevation (wse) at the boundary. The wse [m] at the boundary must be specified by the user. The depth is calculated as:

$$h = wse - z_B \quad (1.13)$$

where  $z_B$  is the bottom elevation at the boundary. The flow velocity is calculated with the Riemann solver (HLLC).

- *zero\_gradient\_out* (scientific use only): Transmissive, or transparent boundaries allow the passage of waves without any effect on them. This is mathematically obtained imposing over the entire length of the boundary that:

$$\rho \vec{u} \cdot \vec{n} = \text{const} \quad ; \quad \frac{\partial \vec{u}}{\partial \vec{n}} = 0 \quad (1.14)$$

In this case there is no need to specify further parameters.

*Note: This is boundary condition should **not** be used for practical problems and is intended for scientific use only.*

#### 1.1.4.2 Linked Boundary Conditions

This type of boundaries establish a *link* between within a certain region of the domain where equations are not solved. Once this domain portion is identified the two boundaries, between which the link is established, must be specified. Let us call them  $\Gamma_{in}$  and  $\Gamma_{out}$ . Then, one inflow boundary condition must be specified at  $\Gamma_{in}$  and one outflow boundary condition at  $\Gamma_{out}$  while in the remaining boundaries wall conditions are automatically assigned. Not necessarily,  $\Gamma_{in}$  and  $\Gamma_{out}$  must have the same number of elements.

Linked boundaries can describe a  $h - Q$  relation, a weir, or prescribed water surface elevation, i.e.:

- *weir\_linked\_constant* and *weir\_linked\_dynamic*: Similar to the standard weir boundary, the weir height  $w$  has to be specified.
- *hqrelation\_linked*: The discharge is determined as a function of the water surface elevation, thus a stage-discharge-relation has to be specified.
- *2way\_hqrelation\_linked*: The internal boundary works as dynamic wall that is controlled by water surface elevation thresholds. If the upper water surface elevation threshold is reached, the internal boundary is removed until the water level reaches the lower water surface elevation, where the wall is re-established.
- *hydrograph\_linked*: Sets a prescribed water surface elevation (WSE) at the upstream boundary. The flux over the upstream boundary is calculated with the Riemann solver and used as inflow in the downstream boundary.

The inflow conditions at the downstream boundary can be controlled via the tag *type\_downstream*. Four options are available as downstream inflow condition. For all options, the mass inflow flux is calculated based on outflow at upstream boundary. The difference between the options lies in the calculation of the momentum flux over the boundary:

- *no\_momentum*: No momentum flux over the boundary.
- *froude\_downstream*: Momentum flux calculated based on Froude number in the elements of the downstream nodestring.

- *froude*: Momentum flux calculated based on the Froude number specified via the parameter *froude\_number*.
- *uniform*: Momentum flux calculated based on uniform flow conditions for the slope specified via the parameter *slope*.

In BASEMENT versions without the option *type\_downstream*, the various linked boundary conditions differ in how the momentum flux at the downstream boundary was calculated. The following combinations correspond to the implementation of the boundary conditions in previous versions:

type	type_downstream
weir_linked_constant	froude_downstream
weir_linked_dynamic	froude_downstream
hqrelation_linked	no_momentum
2way_hqrelation_linked	no_momentum
zhydrograph_linked	no_momentum
zhydrograph_linked_kinE	froude_downstream

#### 1.1.4.3 Internal Boundary Conditions

The internal boundary condition allows a direct cell-cell relation due to the exact same number of elements on the left and on right side of the boundary. Internal boundary conditions can be used to specify internal walls, dynamic walls or an h-Q relation.

- *wall\_internal*: The wall conditions (eq. 1.7) are applied on both sides of the internal boundary.
- *dynamic\_wall\_internal*: The wall conditions are applied on the internal boundary until reaching a threshold value (time or water depth) after which the wall is removed.
- *hqrelation\_internal*: A stage-discharge relation is applied on one side of the internal boundary, while on the other side, wall conditions apply (unidirectional flow).

For the *hqrelation\_internal* boundary condition, the inflow conditions at the downstream boundary can be controlled via the tag *type\_downstream*. Four options are available as downstream inflow condition. For all options, the mass inflow flux is calculated based on outflow at upstream boundary. The difference between the options lies in the calculation of the momentum flux over the boundary:

- *no\_momentum*: No momentum flux over the boundary.
- *froude\_downstream*: Momentum flux calculated based on Froude number in the elements of the downstream nodestring.
- *froude*: Momentum flux calculated based on the Froude number specified via the parameter *froude\_number*.
- *uniform*: Momentum flux calculated based on uniform flow conditions for the slope specified via the parameter *slope*.

In BASEMENT versions without the option *type\_downstream*, the various linked boundary conditions differ in how the momentum flux at the downstream boundary was calculated. The following combinations correspond to the implementation of the boundary conditions in previous versions:

type	type_downstream
hqrelation_internal	no_momentum

### 1.1.5 Flood Tracking

The flood tracking aims at extracting the flood arrival time, the maximum water depth, flow velocity, specific discharge and bed shear stress along the numerical simulation and over a selected domain area. The flood tracking provides outputs within a tracking time step defined by the user.

## 1.2 Morphodynamics

### 1.2.1 Introduction

Morphodynamic models provide scientific frameworks for advancing our understanding of river systems. The research on involved topics is an important and socially relevant undertaking regarding our environment. Nowadays numerical models are used for different purposes, from answering questions about basic morphodynamic research to managing complex river engineering problems. Due to increasing computer power and the development of advanced numerical techniques, morphodynamic models are now more and more used to predict the bed patterns evolution to a broad spectrum of spatial and temporal scales. The development and the success of application of such models are based upon a wide range of disciplines from applied mathematics for the numerical solution of the equations to geomorphology for the physical interpretation of the results.

Applications of morphodynamic models include:

- Damming of river basins
- Morphological changes due to width changes (e.g. River widenings)
- Effects of sediment mining
- River straightening

### 1.2.2 Bedload Sediment Transport

#### 1.2.2.1 Governing Equations for Uniform Sediment Transport

The governing equations are obtained under shallow water conditions imposing mass conservation for the fluid and solid phases and the momentum principle to a flow in an open channel with a cohesionless bottom.

Introducing a Cartesian reference system  $(x; y; z)$  in which the  $z$  axis is vertical and the  $x - y$  plane is horizontal, the system of governing equations is described by the system of equations (eq. 1.1) for hydrodynamics coupled with one equation for the conservation of the total sediment mass. The conservation of sediment mass is ensured by the Exner equation (eq. 1.15), named after the Austrian sedimentologist Felix M. Exner (Exner, 1925). The Exner equation allows to describe the bed evolution due to erosion or deposition, which results in the elevation change of the actual bed level  $z_B$ :

$$(1 - p) \frac{\partial z_B}{\partial t} + \frac{\partial q_{B_x}}{\partial x} + \frac{\partial q_{B_y}}{\partial y} - Sl_b = S_s \quad (1.15)$$

where  $p$  is the porosity,  $Sl_b$  is the source term per unit width specifying local input or output of sediment material (e.g. slope collapse or excavation), and  $\vec{q}_B = \begin{pmatrix} q_{B_x} \\ q_{B_y} \end{pmatrix}$  is the specific bedload flux. The term  $S_s$  describes the exchange per unit width between the sediment and the suspended material (see Section 1.3). The Exner equation describes the bed evolution due to erosion or deposition processes, which results in changes of the bed level  $z_B$ .

The Exner equation is solved in a decoupled way, meaning that the shallow water equations and the Exner equation are solved in sequence. This approach makes the assumption that the bedload flux is much slower than the water flow velocity (Soares-Frazão and Zech, 2011).

### 1.2.2.2 Threshold Conditions for Sediment Movement

The key dimensionless parameter quantifying sediment mobility is the Shields parameter defined as:

$$\theta = \frac{|\vec{\tau}_b|}{(\rho_s - \rho)gd} \quad (1.16)$$

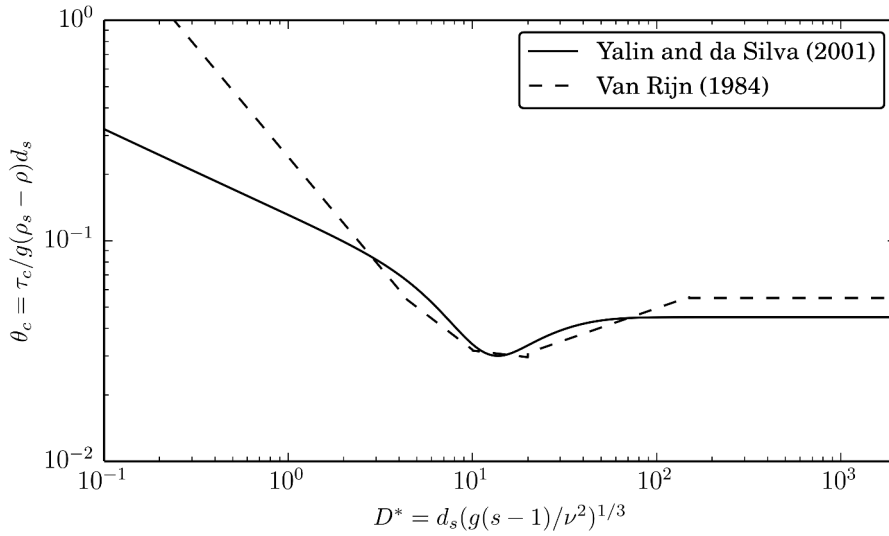
where  $vec\tau_b$  is the bed shear stress (drag force acting on the particle),  $d$  is the sediment diameter,  $\rho$  and  $\rho_s$  are the water and sediment density, respectively. The Shields parameter can be interpreted as the ratio scaling the impelling force of flow drag acting on a particle to the Coulomb force resisting motion acting on the same particle.

The bed shear stress  $\vec{\tau}_b = \begin{pmatrix} \tau_{bx} \\ \tau_{by} \end{pmatrix}$  is usually estimated by a closure condition using an empirical or semi-empirical formula. Here we use the quadratic friction law which relates the depth-averaged velocities to the bed shear stress as follows:

$$\tau_{bx} = \rho \frac{|\vec{u}|u}{c_f^2} \quad ; \quad \tau_{by} = \rho \frac{|\vec{u}|v}{c_f^2} \quad (1.17)$$

where  $\vec{u} = \begin{pmatrix} u \\ v \end{pmatrix}$  is the flow velocity vector,  $\rho$  is the density of water and  $c_f$  is the dimensionless Chézy friction coefficient as defined in Section 1.1.3.1.

When a granular bed is subjected to a turbulent flow, it is found that virtually no motion of the grains is observed below a critical value ( $\theta_{cr}$ ) of the Shields parameter. According



**Figure 1.2** Modified Shields diagram for initiation of sediment motion

to the Shields' theory Shields (1936),  $\theta_{cr}$  can be expressed as a function of the Reynolds number  $Re^* = \frac{du_*}{\nu}$ . Alternatively, the diagram of incipient motion (see Figure 1.2) can be plotted as a function of the dimensionless grain diameter  $D^*$  ( $\theta_{cr} = f(D^*)$ ), where

$$D^* = d \left[ \frac{g(s-1)}{\nu^2} \right]^{1/3}$$

The curve representing the particle incipient motion ( $\theta = \theta_{cr}$ ) can be divided into three parts in the log-log graph:

- for  $D^* \leq 3$ , can be approximated by a linear segment;
- for  $3 \leq D^* \leq 100$  this is represented by a curve with a relative minimum;
- for  $D^* > 100$  by a constant trend.

An approximation of the original Shields diagram was proposed by van Rijn (1984a):

$$\begin{aligned} \theta_{cr} &= 0.24(D^*)^{-1} & \text{for } 1 \leq D^* \leq 4 \\ \theta_{cr} &= 0.14(D^*)^{-0.64} & \text{for } 4 < D^* \leq 10 \\ \theta_{cr} &= 0.04(D^*)^{-0.1} & \text{for } 10 < D^* \leq 20 \\ \theta_{cr} &= 0.013(D^*)^{0.29} & \text{for } 20 < D^* \leq 150 \\ \theta_{cr} &= 0.055 & \text{for } D^* > 150 \end{aligned} \quad (1.18)$$

Another explicit formulation of the Shields curve was proposed by Yalin and Silva (2001). It reads

$$\theta_{cr} = 0.13D^{*-0.392} \exp(-0.015D^*) + 0.045 (1 - \exp(-0.068D^*)) \quad (1.19)$$



### 1.2.2.2.1 Influence of Local Slope on Incipient Motion

The threshold condition for incipient motion of grains developed by Shields is valid for almost horizontal bed. In case of sloped bed in flow direction or transverse to it, the stability of grains is either increased or reduced due to the gravity. The critical shear stress value can be adapted consequently to account for the influence of local slopes. One approach is to multiply the critical shear stress for almost horizontal bed  $\theta_{cr}$  with the correction factors  $k_l$  and  $k_t$  for the local bed slope in the longitudinal and transversal flow direction. In the following, the critical Shields stress corrected for arbitrary bed slope  $\delta$  is referred to as  $\theta_{c,\delta}$ , defined as:

$$\frac{\theta_{c,\delta}}{\theta_{cr}} = k_l k_t = k \quad (1.20)$$

The correction factor  $k_l$  and  $k_t$  are calculated as suggested by van Rijn (1989):

$$k_l = \cos \delta_l \left( 1 - \frac{\tan \delta_l}{\tan \gamma} \right) \quad (1.21)$$

$$k_t = \cos \delta_t \sqrt{1 - \frac{\tan^2 \delta_t}{\tan^2 \gamma}} \quad (1.22)$$

where  $\delta_l$  is the angle between the horizontal and the bed along flow direction,  $\delta_t$  is the slope angle transversal to the flow direction and  $\gamma$  is the angle of repose of the sediment material.

Other formulations are also available, as for example the one proposed by Chen et al. (2010):

$$k = \frac{1}{\tan \gamma} \left( \cos^2 \left( \frac{\pi}{2} - \delta_l \right) - 1 + \frac{1 + \tan^2 \gamma}{(1 + \tan^2 \delta_l + \tan^2 \delta_t)} \right)^{0.5} + \cos \left( \frac{\pi}{2} - \delta_l \right) \quad (1.23)$$

### 1.2.2.3 Closure Relations for Bedload Transport

In order to solve system (eq. 1.1) and equation (eq. 1.15) we need to specify the closure relations. For the friction terms  $S_{fx}$ ,  $S_{fy}$  and the value of lateral inflow/outflow discharge per unit width  $S_h$  we can use the relations already introduced in the Hydrodynamic part (Section 1.1.3). For the Exner equation we need relations quantifying the bedload discharge. Let us now introduce the dimensionless bedload transport rate  $\Phi$  also known as the Einstein bedload number, first introduced by Hans Albert Einstein in 1950, and given by

$$\Phi = \frac{q_B}{\sqrt{(s-1)gd^3}} \quad (1.24)$$

where  $s = \rho_s/\rho$ .

It is common practice to quantify bedload transport empirically relating  $\Phi$  with either the Shields stress  $\theta$  or the excess of the Shields stress  $\theta$  above some appropriately defined ‘‘critical’’ Shields stress ( $\theta - \theta_{cr}$ ). The critical Shields stress  $\theta_{cr}$  is defined so as to fit experimental or field data and provide a threshold for which the bedload transport rate is too low to be of interest. The Shields parameter, takes the following form

$$\theta = \frac{h\sqrt{S_{fx}^2 + S_{fy}^2}}{(s-1)d} \quad (1.25)$$

where  $h$  is the water depth,  $S_{fx}$  and  $S_{fy}$  the friction slope in x- and y-direction respectively,  $s = \rho_s/\rho_w$ , and  $d$  is the grain size diameter. Note that Eq. 1.16 and Eq. 1.25 are equivalent.

In what follows, we describe the bedload transport formulas that are implemented to calculate the transport capacity  $q_B = |\vec{q}_B|$ , where the specific bedload flux vector  $\vec{q}_B = (q_{B_x}, q_{B_y})$  generally has the same direction as the water flow.

For practical purposes, the bedload transport formula can be calibrated by an additional pre-factor (*factor*). The bedload transport capacity is obtained from the closure relation scaled by this pre-factor.

### 1.2.2.3.1 Meyer-Peter and Müller (1948)

The bedload transport formula of Meyer-Peter and Müller (Meyer-Peter and Müller, 1948) defines the specific bedload transport rate  $q_B$  as:

$$q_B = \alpha(\theta - \theta_{cr})^m \sqrt{(s-1)gd^3} \quad (1.26)$$

Herein,  $\alpha$  denotes the bedload coefficient (originally  $\alpha = 8$ ),  $m$  the bedload exponent (originally  $m = 1.5$ ),  $\theta$  is the dimensionless bed shear stress (Shields parameter),  $\theta_{cr}$  is the critical dimensionless bed shear stress,  $d$  is the grain diameter,  $s = \rho_s/\rho$  and  $g$  stands for the gravitational acceleration. Meyer-Peter and Müller observed in their experiments that the first grains moved already for  $\theta_{cr} = 0.03$ . But as their experiments took place with steady conditions they used a value for which already 50% of the grains were moving. They proposed the value of  $\theta_{cr} = 0.047$ . The formula of Meyer-Peter and Müller is applicable in particular for coarse sand and gravel with grain diameters larger than 1 mm (Malcherek, 2001).

The bedload coefficient  $\alpha$ , the exponent  $m$  and the critical Shields parameter  $\theta_{cr}$  can be adapted by the user in the MPM-like formula.

### 1.2.2.3.2 Grass Formula

The Grass formula (Grass, 1981) proposes a simple bedload transport formula, where  $q_b$  is a function of the flow velocity  $u$  and a dimensional constant  $\alpha$  and does not require the evaluation of the Shields stress:

$$q_B = \alpha(u - u_c)^m \cdot \sqrt{(s-1)gd^3} \quad (1.27)$$

where  $\alpha \in [0, 1]$  is a dimensional constant that encompasses the effects of grain size and kinematic viscosity and is usually determined from experimental data,  $u_c$  is the critical velocity and the exponent  $m$  is usually set to  $m = 3$ . The threshold condition for incipient motion of grains is typically set to zero, meaning that the bedload transport and the fluid motion start simultaneously. The coefficient  $\alpha$  characterizes the interaction between the bed and the fluid. If  $\alpha = 0$ , no sediment transport occurs. If  $\alpha = 1$  the interaction between the bed and fluid is the largest.

### 1.2.2.3.3 Engelund and Hansen (1972)

Engelund and Hansen (1972) proposed a transport formula for uniform bed material taking into account at the same time the presence of both bed- and suspended-load. This formula is commonly used as a bulk load formula and reads

$$q_B = 0.05 \sqrt{(s-1)gd^3} \cdot c_f^2 \theta^{2.5} \quad (1.28)$$

where  $d$  denotes the median sediment size of the bed material,  $c_f$  the dimensionless Chézy friction coefficient and  $\theta$  is the dimensionless bed shear stress (see eq. 1.25). The Engelund and Hansen formula for bedload transport does not consider the critical shear stress as threshold condition for incipient motion.

### 1.2.2.3.4 Smart & Jäggi (1983)

Smart and Jaeggi (1983) developed a bedload transport formula for steep channels using their own experimental results and the results of Meyer-Peter and Müller (Meyer-Peter and Müller, 1948). The specific bedload transport rate  $q_B$  is defined as:

$$q_B = \frac{\alpha}{(s-1)} \left( \frac{d_{90}}{d_{30}} \right)^{0.2} J^{0.6} |\vec{q}| (J - J_{cr}) \quad (1.29)$$

where  $s$  is the sediment density coefficient ( $s = \rho_s/\rho$ ),  $|\vec{q}|$  is the magnitude of the specific discharge and  $d_{30}$  and  $d_{90}$  are the characteristic grain size diameters, i.e. 30 % resp. 90 % (by weight) of the bed material are smaller. The energy slope  $J$  and the critical slope for the initiation of the bedload transport  $J_{cr}$  calculated as

$$J = \frac{\theta(s-1)d_m}{h} \quad (1.30)$$

$$J_{cr} = \frac{\theta_{cr}(s-1)d_m}{h} \quad (1.31)$$

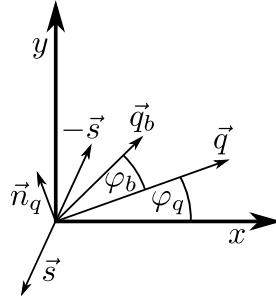
where  $\theta$  is the dimensionless bed shear stress (see eq. 1.25),  $\theta_{cr}$  the critical dimensionless bed shear stress,  $d_m$  the mean grain size diameter and  $h$  the water depth. Smart and Jaeggi (1983) recommend values of  $\alpha = 4$  and  $\theta_{cr} = 0.05$ . The scope of application is for bed slopes  $0.005 \leq J \leq 0.2$  (Smart and Jaeggi, 1983).

### 1.2.2.4 Correction of Bedload Direction

The 2D projection of the solid discharge along  $x$  and  $y$  is obtained through standard procedures, that are mostly based on empirical basis and which account for the downward effect of gravity on sediment particles due to local bed slope and the presence of spiral flow motion in curved reaches.

#### 1.2.2.4.1 Lateral Bed Slope Effect

Empirical bedload formulas were originally derived for situations where bed slope equals flow direction. However, in case of lateral bed slope with respect to flow direction, the



**Figure 1.3** Bed load transport deviation angle  $\varphi_b$  from the flow direction  $\vec{q}$  due to the lateral bed slope  $\vec{s}$  (Vonwiller, 2017)

bedload direction differs from the flow direction due to gravity acting on the bed material. Figure 1.3 illustrates the deviation of the bedload transport direction due to lateral bed slope in a Cartesian coordinate system.

The bedload direction is corrected for lateral bed slope based on the following approach (e.g. see Ikeda (1982) and Talmon et al. (1995)):

$$\tan \varphi_b = -f(\theta) \cdot \vec{s} \cdot \vec{n}_q \quad \text{for } \vec{s} \cdot \vec{n}_q < 0 \quad (1.32)$$

$$f(\theta) = N_l \left( \frac{\theta_{cr}}{\theta} \right)^{M_l} \quad (1.33)$$

where  $\varphi_b$  = bedload direction with respect to the flow vector  $\vec{q}$ ,  $N_l$  = lateral transport factor ( $0.75 \leq N_l \leq 2.63$ ),  $M_l$  = lateral transport exponent (typically  $M_l = 0.5$ ),  $\vec{s} = \left( \frac{\partial z_B}{\partial x}, \frac{\partial z_B}{\partial y} \right)$  bed slope (positive uphill, negative downhill),  $\vec{n}_q$  = unit vector perpendicular to  $\vec{q}$  pointing in downhill direction ( $\vec{s} \cdot \vec{n}_q < 0$ ),  $\theta$  = effective dimensionless shear stress and  $\theta_{cr}$  = critical dimensionless shear stress of sediment.

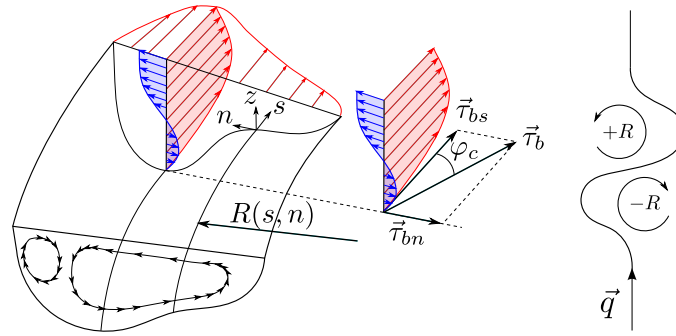
The direction of the bedload transport under the influence of lateral bed slope is written as:

$$\frac{q_{B_y}}{q_{B_x}} = \tan(\varphi_b + \varphi_q) \quad (1.34)$$

#### 1.2.2.4.2 Curvature Effect

Curvature in rivers may cause deviation of the bedload direction from the depth averaged flow direction. Due to three dimensional spiral flow motion, the bedload direction tends to point towards the inner side of the curve, while the flow direction points towards the outer side (Figure 1.4). This curvature effect is taken into account according to an approach proposed by Engelund (1974), where the deviation angle  $\varphi_c$  of the bottom shear stress  $\vec{\tau}_b$  (positive counterclockwise and vice versa) from the main flow direction is determined as

$$\tan \varphi_c = \frac{|\vec{\tau}_{bn}|}{|\vec{\tau}_{bs}|} = -N_* \frac{h}{R} \quad (1.35)$$



**Figure 1.4** Effect of spiral motion in river bend on bed shear stress  $\vec{\tau}_b$  with deviation angle from main flow direction  $\varphi_c$  (Vonwiller, 2017)

where  $\vec{\tau}_{bn}$  and  $\vec{\tau}_{bs}$  are the bed shear stress normal to and in the flow direction respectively,  $h$  denotes the water depth,  $N_*$  is a curvature factor, and  $R$  denotes the radius of the river bend (positive for curvature in counterclockwise direction and vice versa).

Note that the curvature factor  $N_*$  mainly depends on bed roughness. Therefore,  $N_* \approx 7$  for natural streams (Engelund, 1974), and values up  $N_* \approx 11$  for laboratory channels (Rozovskii, 1961).

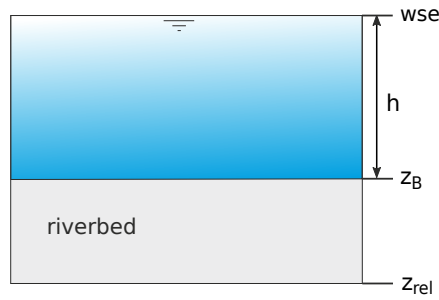
### 1.2.2.5 Fixed Bed Concept

Morphodynamic simulations generate deposition and erosion patterns of the riverbed. Erosion processes, if not limited, can proceed indefinitely in the vertical direction. This limit can be imposed by defining a non-erodible fixed bed elevation  $z_{rel}$ , below which the river bed is considered as *fixed*. This threshold also determines the amount of sediment available for transport (see Figure 1.5). The fixed bed elevation can either be assigned via regions or via a separate .2dm mesh file. When the fixed bed elevation is specified via regions, the fixed bed elevation must be provided relative to the initial bottom elevation  $z_B$  with  $z_{rel} \leq 0$ . When the fixed bed elevation is specified via a mesh, the elevation in the separate .2dm mesh file must correspond to the absolute fixed bed elevation  $z_{fix}$  [m]. Moreover, the fixed bed mesh file must have the exact same topology as the original computational mesh. In case the specified mesh only has elevation information on the nodes, the elevation is interpolated with the same method as specified in the INTERPOLATION block (default: mean). If the fixed bed elevation of the fixed bed mesh exceeds the bottom elevation of the computational mesh, the fixed bed elevation is defined at the elevation of the computational mesh.

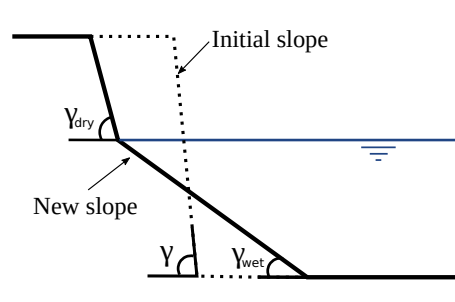
The accuracy of the fixed bed correction is guaranteed by defining the maximal overshoot below the fix bed elevation and the maximal number of iterations required for the correction.

### 1.2.2.6 Gravitational Transport

Gravitational induced riverbank or sidewall failures are significant aspects concerning erosion and transport modelling. Such processes may play an important role in many situations, such as meandering streams, river widenings or failures of erodible embankment



**Figure 1.5** Fixed bed concept and definition



**Figure 1.6** Critical failure angles for slope collapse

structures due to overtopping waters. Such slope failure processes take place mostly discontinuous and can deliver significant contributions to the total sum of transported material. The modes of slope failures can differ largely (falls, topples, slides, etc.) and depend on the soil material, the degree of soil compaction and the pore pressures within the soil matrix. Here, a simplified, geometric approach is applied to be able to consider some aspects of this purely gravitational induced transport. The main idea of the implemented geometrical approach is to assume that a slope failure takes place if the local bed slope  $\gamma$  becomes steeper than a critical slope  $\gamma_{cr}$  (Figure 1.6).

$$q_{B,grav} = \begin{cases} 0 & \text{if } (\gamma \leq \gamma_{cr}) \\ f(\gamma, \gamma_{cr}) & \text{if } (\gamma > \gamma_{cr}) \end{cases} \quad (1.36)$$

The sliding material is moved from the sediment element with higher elevation to the lower situated element until the stable condition:  $\gamma \leq \gamma_{cr}$  is reached. Two characteristic critical slope angles are defined in this approach to have some flexibility in modelling the complex geotechnical aspects. The critical angles can be characterized as:

- critical angle for dry or partially saturated bank material  $\gamma_{dry}$ , which may greatly exceed the material's angle of repose (up to nearly vertical walls) due to negative pore pressures,
- critical angle for fully saturated and over flown material  $\gamma_{wet}$  which is in the range of the material's angle of repose

#### 1.2.2.6.1 Calculation Procedure

The flux due to gravitational transport  $q_{B,grav}$  is calculated with the following procedure, by looping over each element of the computational grid:

1. In a first step, the local bed slope  $\gamma_i$  is calculated with respect to each neighbouring element  $i$ , where  $z$  is the bed elevation of the main element,  $z_i$  is the bed elevation of the neighbour element  $i$  and  $d_i$  is the distance between the element centers.

$$\gamma_i = \arctan\left(\frac{z - z_i}{d_i}\right) \quad (1.37)$$

2. For local bed slopes  $\gamma_i$  exceeding the critical slope  $\gamma_{cr}$ , the new bed elevations  $z_{new}$  and  $z_{i,new}$  are determined such that the stable condition  $\gamma_i = \gamma_{cr}$  is reached for all neighbouring cells  $i$ . The critical angle  $\gamma_{cr}$  is selected according to eq. 1.38, where  $h$  is the water depth in the main element and  $h_{min}$  is the user-specified minimum water depth.

$$\gamma_{cr} = \begin{cases} \gamma_{wet} & \text{if } h \geq h_{min} \\ \gamma_{dry} & \text{if } h < h_{min} \end{cases} \quad (1.38)$$

3. If the calculated bed elevation change of the main element  $\delta_z = z - z_{new}$  is smaller than the user-specified parameter  $min\_bed\_change$  (default: 0.001 m)  $\delta_{z,min}$ , no gravitational transport occurs to avoid oscillatory behaviour and to reduce the computational effort. If the minimum bed elevation change  $\delta_{z,min}$  is exceeded, the specific gravitational flux  $q_{B,grav,i}$  [ $m^2/s$ ] to each neighbour element  $i$  is calculated from the bed elevation change in element  $i$  according to eq. 1.39, where  $A_i$  is the area of element  $i$ , and  $l_i$  is the length of the edge connecting the main element to its  $i^{th}$  neighbour element. If the calculated bed elevation change of the main element  $\delta_z = z - z_{new}$  is larger than the user-defined maximum bed elevation change  $\delta_{z,max}$ , the gravitational flux is limited, such that  $\delta_z = \delta_{z,max}$ . The maximum bed elevation change  $\delta_{z,max}$  is calculated by eq. 1.40, where  $r_{b,max}$  is the user-specified parameter  $max\_bed\_change\_rate$  and  $\Delta t$  is the current update time step.

$$q_{B,grav,i} = \begin{cases} 0 & \text{if } \delta_z < \delta_{z,min} \\ \frac{(z_{i,new} - z_i) \cdot A_i}{l_i} & \text{if } \delta_z \geq \delta_{z,min} \\ \frac{(z_{i,new} - z_i) \cdot A_i}{l_i} \cdot \frac{\delta_{z,max}}{\delta_z} & \text{if } \delta_z \geq \delta_{z,max} \end{cases} \quad (1.39)$$

$$\delta_{z,max} = r_{b,max} \cdot \Delta t \quad (1.40)$$

4. If a non-erodible fixed bed elevation  $z_{fix}$  is specified, the gravitational transport flux is corrected the same way as the bedload transport flux (see Section 1.2.2.5).
5. Finally, the balancing of the gravitational fluxes and the determination of the new bed elevations  $z$  is achieved by solving the Exner equation using the same numerical approaches as outlined for the bed load transport. This procedure ensures that fixed bed elevations are taken into account and the mass continuity is fulfilled.

### 1.2.2.6.2 Time Scale of the Gravitational Transport Process

Since the sediment movement due to gravitational transport during one update time step is limited to adjacent elements, it may take many update time steps to reach a stable condition on a larger scale, e.g. a bank slope spanning over multiple elements. The time scale until a globally stable condition is reached, is influenced by the update time step (`update_time`), the maximum bed change rate (`max_bed_change_rate`) and the grid resolution. The parameter `update_time` (default: 0.0 s) determines at which frequency the gravitational transport procedure (steps 1-5 above) is executed. Generally, a smaller update time step reduces the time scale until a globally stable condition is reached. The default behaviour is to set the update time step value to 0.0 seconds, which results in the gravitational transport procedure being executed at the same time step as determined from the hydraulic CFL-criterion (see Section 2.3.4).

Furthermore, the speed of the gravitational transport process can be limited with the parameter `max_bed_change_rate` (default: 1.0 m/s). This parameter represents a maximum rate at which the bed elevation of a cell can be lowered due to gravitational transport and determines the maximum bed elevation change during one update time step. Generally, a smaller value may increase the time scale until a globally stable condition is reached.

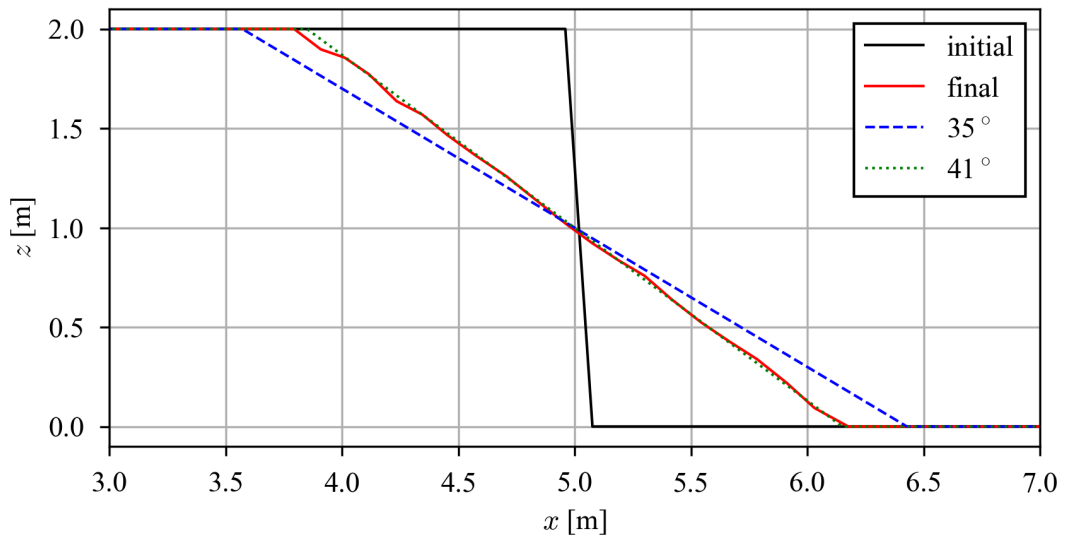
Since sediment movement due to gravitational transport is limited to adjacent elements, also the grid resolution effects the time scale until a globally stable condition is reached. A finer grid resolution (smaller elements) increases the necessary number of update cycles the reach a stable slope over a specific length. However, a finer grid resolution may also decrease the hydraulic time step and therefore may the increase frequency at which the gravitational transport procedure is executed.

### 1.2.2.6.3 Bed slope at Stable Condition

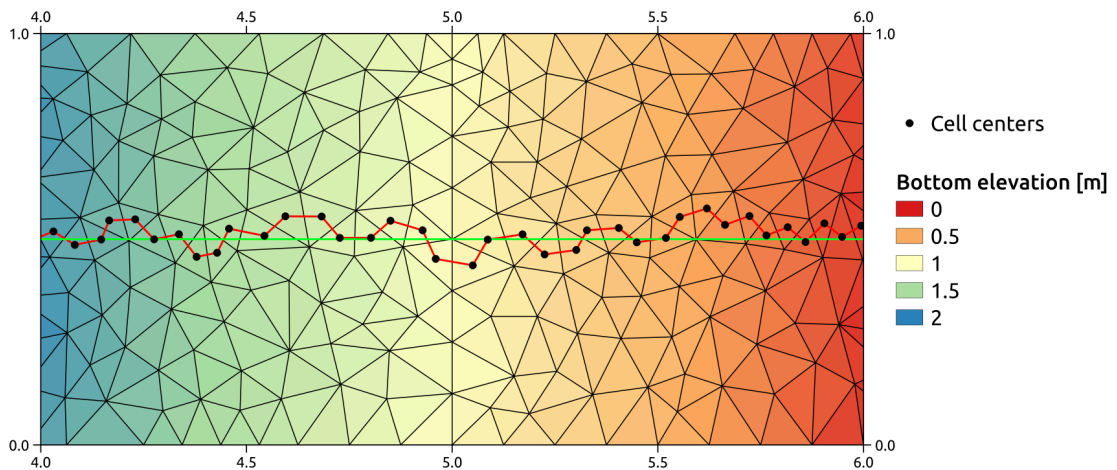
The implemented approach for modelling gravitational transport processes ensures that the local bed slope does not exceed the user-specified critical angle. Due to the spatial discretization with constant bed elevations within an element, the local bed slope is calculated between two element centers. On a larger scale (e.g. a bank slope spanning over multiple elements), the bed slope may deviate from the user-specified critical angle after reaching stable conditions. The stable bed slope over multiple elements, generally exceeds the critical bed slope.

To illustrate this effect, a simple test case was simulated. The rectangular computational domain is initially split into two parts: the left side with a bed elevation of 2 meters and the right side with a bed elevation of 0 meters. The domain is completely dry and the critical angle for dry material is set to  $\gamma_{dry} = 35^\circ$ . The simulation only considers only gravitational transport and is stopped after a stable condition is reached. The initial and final bed elevation profile along the centerline are illustrated in Figure 1.7. The initial profile exhibits the nearly vertical bed step. The final profile at stable conditions exhibits a slope of approximately  $41^\circ$ , which exceeds the critical angle of  $\gamma_{dry} = 35^\circ$ . The reason for this deviation on a larger scale is that the distance which is considered for the calculation of the slope is not a straight line, but a line connecting the element centers. This is illustrated in Figure Figure 1.8. Calculating the large scale slope considering the length of the green line results in a slope of approximately  $41^\circ$ , while considering the length of the red line results in a slope of approx.  $35^\circ$ .





**Figure 1.7** Initial and final bed elevation profiles along the centerline are compared to profiles with angles of  $35^\circ$  and  $41^\circ$ . The slope of the bed profile at stable conditions (final) exceeds the critical angle of  $\gamma_{\text{dry}} = 35^\circ$ .



**Figure 1.8** The red line connects the element centers and represents the distance, which is taken into account for the **local slope** calculation and thus for the gravitational transport. The green line represents the direct distance, which is taken into account for the **large scale slope**. The slope along the red line corresponds to  $35^\circ$ , while the slope along the green line corresponds to  $41^\circ$ .

### 1.2.2.7 External Sources Terms

The source term  $Sl_b$  represents additional sediment mass input or output (sink) that occurs locally on the computational domain on a set of elements defined by regions. The source can be specified as total volume flux including porosity [ $m^3/s$ ]. Different approaches are used to manage the behaviour of the external sources in case of a negative flux (sink):

- Exact: The specified sediment volume is added or extracted (non conservative)
- Available: The specified sediment volume to extract is limited by the defined fixed bed elevation of the elements (conservative)
- Infinity: All available sediment will be abstracted (conservative)

Addition of sediment always follows the “Exact” behaviour as there is no upper limit. The abstraction of sediment could also follow the “Exact” behaviour but the simulation might end abruptly if the available sediment volume is smaller than the volume abstracted. Therefore, the “Available” behaviour aims to avoid this situation. The “Infinity” behaviour abstracts all available sediment volume.

### 1.2.2.8 Boundary Conditions

After the specification of the *closure relations* for the sediment transport, the system of governing equations (eq. 1.1) and (eq. 1.15) can be solved within the modeling domain described in Figure 1.1, provided boundary conditions (morphologic boundary conditions) are specified at the domain boundary  $\Gamma$ . For the sediment transport only *external boundaries* that allow sediment flowing into or out of the domain can be specified. A morphologic boundary condition can be co-located with a hydraulic boundary condition. In case no hydraulic boundary condition is specified, the boundary will behave as a wall and sediment transport will not occur.

### 1.2.2.9 Upstream Boundary Condition

- *equilibrium\_in*: After erosion or deposition up to a user specified reference bed elevation (*reference\_bed\_elevation*) this upstream boundary condition grants a equilibrium condition, i.e. the same amount of sediment leaving the first computational cell in flow direction enters the cell from the upstream boundary. This leads to a constant bed elevation at the boundary condition.
- *sedimentograph*: based on a sediment hydrograph describing the bedload inflow as function of time (constant or variable). The bedload is defined as a volumetric flow rate  $Q_b = \frac{\mu_s}{\rho_s} [m^3/s]$ , where  $\mu_s$  is the sediment mass flow rate [ $kg/s$ ] and  $\rho_s$  the sediment density [ $kg/m^3$ ]. Notice that the porosity is not considered in the bedload input and is specified separately as own parameter value. The volumetric flow rate is either distributed using a geometrical weighting (*sedimentograph*), using wetted area weighting (*sedimentograph\_warea*) or using wetted conveyance weighting (*sedimentograph\_conveyance*). **Note:** When using the pre-factor (*factor*) described in Section 1.2.2.3, it is automatically applied to the volumetric flow rate at the boundary for these types of boundary conditions, i.e. the sediment hydrograph is scaled by the pre-factor.

- *transport\_capacity*: the sediment inflow is defined by calculating the equilibrium transport capacity according to the hydraulic state at the boundary. The bedload is defined as a compact volumetric flow rate (without porosity)  $Q_b$  [ $m^3/s$ ]. The volumetric flow rate is either distributed using a geometrical weighting (*transport\_capacity*), using wetted area weighting (*transport\_capacity\_warea*) or using wetted conveyance weighting (*transport\_capacity\_conveyance*). **Note:** When using the pre-factor (*factor*) described in Section 1.2.2.3, the pre-factor is implicitly included in the volumetric flow rate at the boundary for this type of boundary condition, i.e. the transport capacity at the boundary is scaled by the pre-factor. Additionally, an independent scaling factor can be specified (*boundary\_factor*), only applying to these types of the boundaries.

For the sediment discharge and transport capacity boundary condition types, the specific sediment discharge  $q_b$  is distinguished by three weighting schemes:

1. Geometrical weighting with respect to the total nodestring length  $L_n$ .

$$q_b = \frac{Q_b}{L_n} \quad \left[ \frac{m^3}{s \cdot m} \right] \quad (1.41)$$

2. Wetted area weighting

$$q_b = \frac{Q_b}{A_{w,tot}} \cdot h \quad \left[ \frac{m^3}{s \cdot m^2} \right] \quad (1.42)$$

3. Conveyance weighting

$$q_b = \frac{Q_b}{K_{tot}} h \sqrt{c_f h} \quad \left[ \frac{m^3}{s \cdot m} \right] \quad (1.43)$$

with  $K_{tot} = A_{w,tot} \sqrt{c_f h}$  the total conveyance and  $c_f$  the friction coefficient.

### 1.2.2.10 Downstream Boundary Condition

One downstream boundary condition is available:

- *equilibrium\_out*: After erosion or deposition up to a user specified reference bed elevation (*reference\_bed\_elevation*) this downstream boundary condition grants a equilibrium condition, i.e. all sediment entering the last computational cell will leave the cell over the downstream boundary. This leads to a constant bed elevation at the boundary condition.

### 1.2.2.11 Linked Boundary Condition

One linked boundary condition is available:

- *equilibrium\_linked*: At the upstream boundary, erosion or deposition is possible up to a user specified reference bed elevation (*reference\_bed\_elevation*). After reaching the reference elevation, this boundary condition grants an equilibrium condition, i.e. all sediment leaving the computational cells on the upstream side is entering at the downstream boundary with a lag of one timestep. This leads to a constant bed elevation at the upstream boundary.

## 1.3 Suspended Sediment Transport

### 1.3.1 Governing Equations for Uniform Suspended Sediment Transport

Similarly to bedload, the governing equations are also derived under shallow water conditions ensuring mass conservation extends to the suspended sediment phase. In a Cartesian reference system  $(x; y; z)$  in which the  $z$  axis is vertical and the  $x - y$  plane is horizontal, the conservation of suspended sediment mass is ensured by an advection-diffusion equation with source terms, describing the bed interaction and elevation changes arising from resuspension and deposition.

$$\frac{\partial}{\partial t} C_s h + \frac{\partial}{\partial x_i} \left( C_s h u_i - h D_{s_{ij}} \frac{\partial C_s}{\partial x_j} \right) = S_s + Sl_s$$

where  $C_s$  is the concentration of suspended sediment,  $D_{s_{ij}}$  is its diffusivity (tensor),  $Sl_s$  are local sources and  $S_s$  aggregates the source terms that parameterize vertical exchanges with the bed.

### 1.3.2 Closures for Suspended Transport

For suspended sediment transport, the source term  $S_s$  represents the exchange with the bed and is calculated through the difference between the deposition rate  $q_d$  and the resuspension rate  $q_e$ .

$$S_s = q_e - q_d$$

#### 1.3.2.1 Deposition rate

The deposition rate is expressed as the a sink flux:

$$q_d = w_s \alpha_s C_s$$

where  $w_s$  is the settling velocity and  $\alpha_s$  is an adaptation coefficient. The adaptation coefficient can be determined through two distinct approaches, the first being provided by Lin (1984) as

$$\alpha_s = m \left( 3.25 + 0.55 \ln \left( \frac{w_s}{\kappa u_*} \right) \right)$$

where  $\kappa$  is the von Karman constant and  $m$  is a tuning parameter (default value 1.0). The second is based on the critical shear stress approach of Xu (1998)

$$\alpha_s = \begin{cases} \left( 1 - \frac{\tau}{\tau_{c,d}} \right) & \tau < \tau_{c,d} \\ 0 & \tau \geq \tau_{c,d} \end{cases}$$

with

$$\tau_{c,d} = \frac{\rho_s - \rho}{\rho_s} \frac{ghw_s}{mU}$$

and where  $m$  is a calibration parameter which has been suggested to take a value of 0.0018.

### 1.3.2.2 Erosion rate

The erosion (or resuspension) rate,  $q_e$ , is given by

$$q_e = w_s \beta_s C_r$$

where  $w_s$  is the settling velocity of particles (see Section 1.3.2.3),  $\beta_s$  is a calibration constant and  $C_r$  is a reference concentration which can be determined by multiple options available in the literature. According to van Rijn (1984b), the reference concentration can be calculated as

$$C_r = a \frac{d_g T_g(\theta_c)^c}{b D^{*d}}$$

where:

$T_g$  is the dimensionless characteristic number for the bottom shear stress,  $b$  is the reference height above the mean bed bottom,  $D^*$  is the dimensionless diameter and  $a$ ,  $c$  and  $d$  are calibration constants with default values  $(a, b, c, d) = (0.015, 0.05, 1.5, 0.3)$ . Another approach is the one of Zyserman and Fredsøe (1994), calculated as

$$C_r = \frac{a(\theta_s - \theta_c)^c}{1 + b(\theta_s - \theta_c)^d}$$

where  $a$ ,  $b$ ,  $c$  and  $d$  are calibration constants with default values  $(a, b, c, d) = (0.331, 0.72, 1.75, 1.75)$ .

### 1.3.2.3 Settling Velocities of Particles

The settling velocity  $w_s$  of sediment particles is an important parameter when determining suspended load. Many different empirical or semi-empirical formulas have been suggested in literature, with the following being implemented:

**1.3.2.3.1 van Rijn**

The sink rate can be determined according to the grain diameter after van Rijn (1984b):

$$w_s = \frac{(s-1)gd^2}{18\nu} \quad \text{for } 0.001 < d \leq 0.1\text{mm}$$

$$w_s = \frac{10\nu}{d} \left( \sqrt{1 + \frac{0.01(s-1)gd^3}{\nu^3}} - 1 \right) \quad \text{for } 0.1 < d \leq 1\text{mm}$$

$$w_s = 1.1\sqrt{(s-1)gd} \quad \text{for } d \geq 1\text{mm}$$

where  $d$  is the diameter of the grain,  $\nu$  is the kinematic viscosity and  $s = \rho_s/\rho$  is the specific density.

**1.3.2.3.2 Wu and Wang**

Another approach for the computation of the sink velocity is the one of Wu et al. (2000):

$$w = \frac{M\nu}{Nd} \left[ \sqrt{\frac{1}{4} + \left( \frac{4N}{3M^2} (D^*)^3 \right)^{1/n}} - \frac{1}{2} \right]^n$$

where:

$$M = 53.5e^{-0.65S_p}$$

$$N = 5.65e^{-2.5S_p}$$

$$n = 0.7 + 0.9S_p$$

$S_p$  is the Corey shape factor, with a value for natural sediments of about 0.7 (0.3 - 0.9).

**1.3.2.3.3 Zhang**

The Zhang formula (Zhang, 1961) is based on many laboratory data and was developed for naturally worn sediment particles. It can be used in a wide range of sediment sizes in the laminar as well turbulent settling regions:

$$w_s = \sqrt{\left(13.95\frac{\nu}{d}\right)^2 + 1.09(s-1)gd} - 13.95\frac{\nu}{d}$$

**1.3.3 External Source Terms**

The source term  $Sl_s$  represents additional sediment mass input or output (sink) that occurs locally on the computational domain on a set of elements defined by regions. The source can be specified as total volume flux including porosity [ $m^3/s$ ] or by an imposed concentration [-]. Different approaches are used to manage the behaviour of the external sources in case of a negative flux (sink):

- Exact: The specified sediment volume is added or extracted (non conservative)

- Available: The specified sediment volume to extract is limited by the defined fixed bed elevation of the elements (conservative)
- Infinity: All available sediment will be abstracted (conservative)
- Concentration: This sink type allows to impose a specified concentration.

Addition of sediment follows the “Exact” behaviour as there is no upper limit. The abstraction of sediment could also follow the “Exact” behaviour but the simulation might end abruptly if the available sediment volume is smaller than the volume abstracted. Therefore, the “Available” behaviour aims to avoid this situation. The “Infinity” behaviour abstracts all available sediment volume.

### 1.3.4 Boundary Conditions

For the suspended sediment transport only external boundaries that allow sediment flowing into or out of the domain can be specified. Boundary conditions of suspended transport type can be co-located with hydraulic or bedload boundary conditions.

#### 1.3.4.1 Upstream Boundary Condition

Two upstream boundary conditions are available:

- *discharge\_in* and *discharge\_in\_warea*: based on a suspended sediment discharge inflow, either as a constant or a function of time, the prescribed volumetric flow rate [ $m^3/s$ ] is imposed at the boundary. The volumetric flow rate is either distributed using a geometrical weighting or a wet-area weighting.
- *concentration\_in*: the suspended sediment inflow is defined by forcing a target concentration. The suspended sediment input rate is thus given by the target concentration paired with the total hydrodynamic mass flux.

#### 1.3.4.2 Downstream Boundary Condition

- *zero\_gradient\_out*: this downstream boundary allows the free outflow of any tracer quantities in the flow. **Note:** If this condition is not prescribed, a wall condition is assumed and the tracer quantities will be retained (no outflow).

## 1.4 Passive tracers

### 1.4.1 Introduction

A multitude of dissolved species are present in environmental flows. In the context of hydraulic and environmental engineering, numerical modelling of scalar transport becomes a relevant tool mostly because of its versatility. In terms of advective phenomena, the most common applications include

- Pollutant fate and transport
- Accumulation or depletion of nutrients
- Calculation of water residence times
- Flow visualization

## 1.4.2 Transport of passive species

### 1.4.2.1 Governing Equations for passive specie transport

The governing equations are obtained under the shallow water framework and impose mass conservation for both fluid and dissolved phases.

In a Cartesian frame of reference  $(x; y; z)$  in which the  $z$  axis is vertical and the horizontal lies in the  $x - y$  plane, the system of governing equations is formed by (eq. 1.1) for hydrodynamics and is coupled with multiple equations for the conservation of the total tracer masses. The conservation of each tracer mass is ensured by the scalar continuity equation (eq. 1.44), which is tightly coupled to the shallow water equations. This equation allows to describe the evolution of the specie concentration as:

$$\frac{\partial}{\partial t} C_k h + \frac{\partial}{\partial x_i} \left( C_k h u_i - h D_{k_{ij}} \frac{\partial C_k}{\partial x_j} \right) = Sl_k \quad (1.44)$$

where  $C_k$  is the concentration of specie  $k$ ,  $D_{k_{ij}}$  is it's diffusion tensor and  $Sl_k$  is the local source term, specifying additional sources or sinks of the specie  $k$ .

### 1.4.2.2 External Sources Terms

The source term  $Sl_k$  conveys an input or output (sink) that occurs locally on the computational domain over elements limited by regions. It can be specified as total volumetric flux [ $m^3/s$ ] or by an imposed concentration [-]. Different approaches define the behaviour of external sources if it imposes negative fluxes (sink):

- Exact: The specified volume is added or extracted (non conservative)
- Available: The specified volume to extract is limited by the available tracer volume (conservative)
- Infinity: All available tracer volume will be extracted (conservative)
- Concentration: This sink type allows to impose a specified tracer concentration.

Additionally, there is the option of forcing a target concentration  $C_k^f$  homogeneously across all cells defined by the region.



### 1.4.2.3 Boundary Conditions

The system of governing equations (eq. 1.1) and (eq. 1.44) can be solved within the modeling domain described in Figure 1.1, provided that boundary conditions (hydrodynamical and tracer boundary conditions) are specified at the domain boundary  $\Gamma$ . For the tracer transport only *external boundaries* that allow tracer flowing into or out of the domain can be specified. A tracer boundary condition should be co-located with a hydraulic boundary condition. Otherwise, the boundary will behave as a wall and tracer transport will not occur.

### 1.4.2.4 Upstream Boundary Condition

Two upstream boundary conditions are available:

- *discharge\_in* and *discharge\_in\_warea*: based on a tracer discharge inflow, either as a constant or a function of time, the prescribed volumetric flow rate  $Q_k$  [ $m^3/s$ ] is imposed at the boundary. The volumetric flow rate is either distributed using a geometrical weighting or a wet-area weighting, and the specific tracer discharge  $q_k$  is thus given by:

1. Geometrical weighting with respect to the total nodestring length  $L_n$ .

$$q_k = \frac{Q_k}{L_n} \quad \left[ \frac{m^3}{s \cdot m} \right]$$

2. Wetted-area weighting

$$q_k = \frac{Q_k}{A_{w,tot}} \cdot h \quad \left[ \frac{m^3}{s \cdot m} \right]$$

- *concentration\_in*: the tracer inflow is defined by forcing a target tracer concentration  $C_k^f$ . The volumetric flow rate is thus given by the target concentration paired with the total hydrodynamic mass flux  $q$  through that boundary as  $q_k = q\phi_k^f$ .

**Note:** If the hydrodynamical mass flow at the boundary is not inward directed then no tracer flux is imposed.

### 1.4.2.5 Downstream Boundary Condition

One downstream boundary condition is available:

- *zero\_gradient\_out*: this downstream boundary allows the free outflow of any tracer quantities in the flow. **Note:** If this condition is not prescribed, a wall condition is assumed and the tracer quantities will be retained (no outflow).

## 1.5 Riparian vegetation

### 1.5.1 Introduction

Riparian vegetation is known to play an important role in mediating river morphodynamic processes at various spatial and temporal scales. Predicting river morphology has to account for the feedbacks between plants, flow, and sediment transport. On one hand, vegetation can modify the flow resistance and the shear stresses acting on the riverbed, thus influencing the overall flow pattern and rate of sediment transport. On the other hand, erosion and deposition processes may cause plant mortality through uprooting and burial during floods. Including vegetation into river morphodynamic simulations has potential applications for

- estimating sedimentation and erosion processes
- restoration projects
- riparian forests dynamics
- flood hazard mapping

### 1.5.2 Description of Vegetation

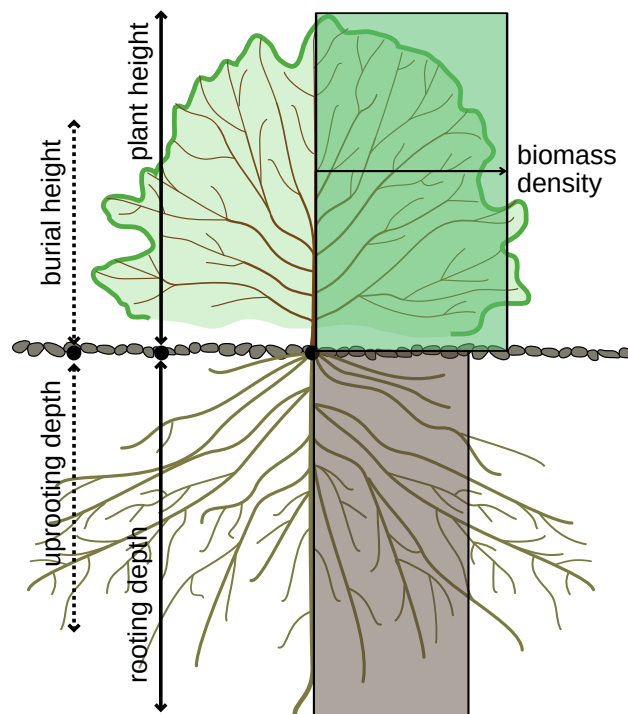
Vegetation is described by a dimensionless aboveground biomass density,  $B_c$  (subscript  $c$  stands for canopy), belowground density,  $B_r$  (subscript  $r$  stands for roots) and a rooting depth,  $D_r$  (see Figure 1.9). The sum of  $B_c$  and  $B_r$  should ideally be 1 at maximum, while there is no specific limit for the rooting depth. The model accounts also for the vegetation height,  $H$ , which is derived as  $H = aB_c^b$ , where  $a$  and  $b$  are constant parameters.  $B_c$  and  $B_r$  are considered to distribute linearly along the main vertical axis of the plant. The uprooting depth,  $D_{upr}$ , namely the depth at which uprooting occurs, is defined as  $\beta_{upr}D_r$ , with  $\beta_{upr}$  a constant input parameter that decreases the maximum erosion that vegetation can withstand. The burial height,  $H_{bur}$ , represents the effective plant height when the plant is submerged (due to bending) and is defined as  $H_{bur} = \beta_{bur}H$ , with  $\beta_{bur}$  a constant parameter (see Section 1.5.5).

### 1.5.3 Effect of Vegetation on Water Flow

The aboveground vegetation is assumed to change the bed roughness. This is included by modifying the Strickler's coefficient  $k_{str}$  [ $m^{1/3}/s$ ] (Bertoldi et al., 2014), such as

$$k_{str}(t) = k_{s,g} + (k_{s,v} - k_{s,g})B_c(t) \quad (1.45)$$

where  $k_{s,g}$  represents the roughness of the bare bed, which depends on the sediment grain size, while  $k_{s,v}$  ( $< k_{s,g}$ ) is the roughness of a completely vegetated bed assumed to vary with species-specific canopy characteristics. The Strickler's coefficient can change during the simulation as  $B_c$  changes (see Section 1.5.5) (Caponi et al., 2020).



**Figure 1.9** Schematic representation of vegetation in the model. The vegetation state is described by an aboveground, belowground biomass, and a rooting depth. The other variables (uprooting depth, burial height, and plant height) are derived through specific functions and input parameters. Adapted from Caponi et al. (2020)

## 1.5.4 Effects of Vegetation on Bedload Transport

### 1.5.4.1 Bottom Shear Stress

The presence of vegetation is also known to affect the shear stresses acting on the bed surface and responsible for sediment transport. The reduction of bed shear stress is included in the model by multiplying the total shear stress by a factor  $\gamma < 1$  and computing the sediment flux using a reduced dimensionless bed shear stress (Shields parameter),  $\gamma\theta$ . The parameter  $\gamma$  ranges between 0 and 1 and it is chosen according to  $\gamma = k_{str}(t)/k_{s,g}$ , with  $k_{str}$  evaluated as in eq. 1.45, (Caponi and Siviglia, 2018).

### 1.5.4.2 Critical Shear Stress

The role of root-enhanced riverbed cohesion is taken into account by increasing the critical Shields parameter (Bertoldi et al., 2014). Assuming an MPM-like formula (eq. 1.26), the critical Shields parameter  $\theta_{cr}$  is defined as

$$\theta_{cr}(t) = \theta_{cr,g} + (\theta_{cr,v} - \theta_{cr,g})B_r(t) \quad (1.46)$$

in which  $\theta_{cr,g}$  and  $\theta_{cr,v}$  ( $> \theta_{cr,g}$ ) represent the values used for bare bed and completely vegetated riverbed, respectively (Caponi and Siviglia, 2018).

## 1.5.5 Effects of Bed Changes on Vegetation

Bed level change causes three effects on vegetation in the model: uprooting, burial, and biomass redistribution.

### 1.5.5.1 Biomass Redistribution

As soon the bed level changes, the portion of vegetation above and below ground changes as well. The model assumes that the total biomass remains constant during a simulation, unless uprooting occurs. This means that in case of a positive bed level change,  $\Delta z_B > 0$  (deposition),  $B_c$  will decrease by  $\Delta z_B(B_c/H_{bur})$  and  $B_r$  will increase by the same amount, where  $H_{bur}$  is a fraction of the plant height. On the contrary, in case of bed level erosion,  $B_c$  will increase by  $\Delta z_B(B_r/D_r)$  and  $B_r$  will decrease by the same amount (Caponi et al., 2020). Changes in  $B_c$  and  $B_r$  during the simulation will change the effects on flow and sediment transport associated with these values following eq. 1.46 and eq. 1.45.

### 1.5.5.2 Plant Burial

Plant burial is the mechanism by which a plant get covered by sediments, that is, when the bed level increases in a cell until reaching the burial plant height  $H_{bur}$ . This value takes into account the bending of the plant when submerged, which reduces the effective height of the plant. In case the bed level change reaches  $H_{bur}$ ,  $B_c$  will go to zero, causing the  $k_{str}$  value to equal  $k_{s,g}$ . In case the bed level decreases again during the simulation,  $B_c$  can increase as well (Caponi et al., 2020). Plants can in fact resist well sedimentation processes, thanks to the high flexibility of the stems that prevents breakage.

### 1.5.5.3 Plant Uprooting

Plant uprooting occurs when the pulling forces applied on the plant by the water flow equal the resisting forces of the plant, which are given by the root system anchoring. We consider that this can occur only when part of the plant roots are already exposed to the flow, that is, when the bed level has decreased during the simulation (type II of uprooting, following the conceptual model of Edmaier et al. (2018)). Therefore, the uprooting is modelled by defining a critical depth at which the plant is removed. This depth is currently defined as a percentage of the rooting depth as  $D_{cr} = \beta_{upr} D_r$ , where  $\beta_{upr}$  is a parameter that modulates the resistance of the plant and depends on soil characteristics and plant species. When uprooting occurs,  $B_c$  and  $B_r$  are set to zero and vegetation cannot influence flow and sediment transport anymore during the simulation.



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# Numerical Models

## 2.1 General View

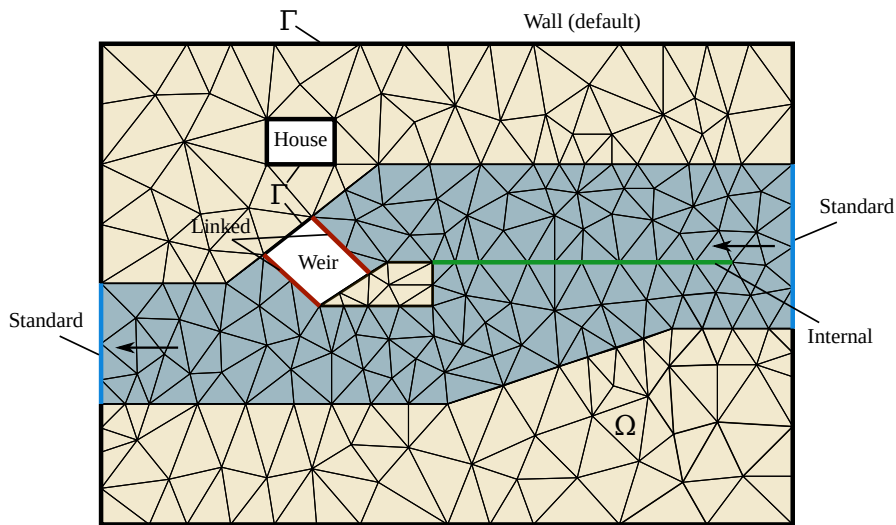
The governing equations of hydro- and morphodynamics are conservation laws expressing conservation of mass and momentum. The aim of the numerical simulation is to solve these equations over the computational domain and for a given time. The computational domain is discretized by a computational mesh consisting of elements and conservation equations are applied on each domain element. In order to numerically solve the conservation equations, the mathematical model is approximated by numerical schemes, i.e. the numerical approximation consists of the spatial and temporal discretization of the conservation equations including an algorithm that solves the discretized equations.

For BASEHPC::BASEplane, the spatial discretisation of the domain is based on an unstructured mesh made of triangular elements (Figure 2.1). For the conservation equations, the spatial discretisation follows the finite volume scheme, while for the temporal discretisation an explicit first order Euler scheme is used. The numerical model processes the hydro- and morphodynamic equations in a decoupled way (Figure 2.2).

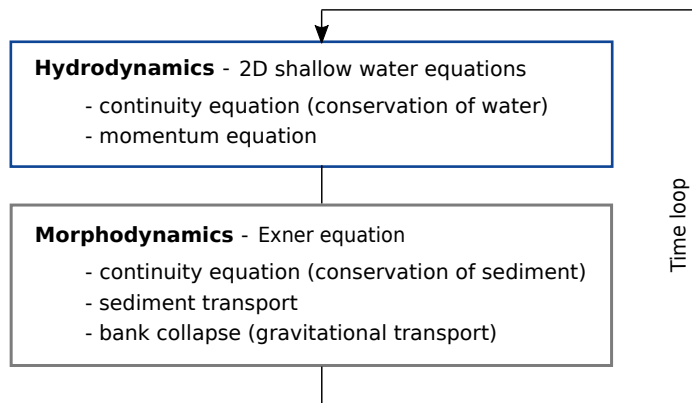
The discretization and the solution method for the hydro- and morphodynamic equations will be presented in the following sections.

## 2.2 Discretization

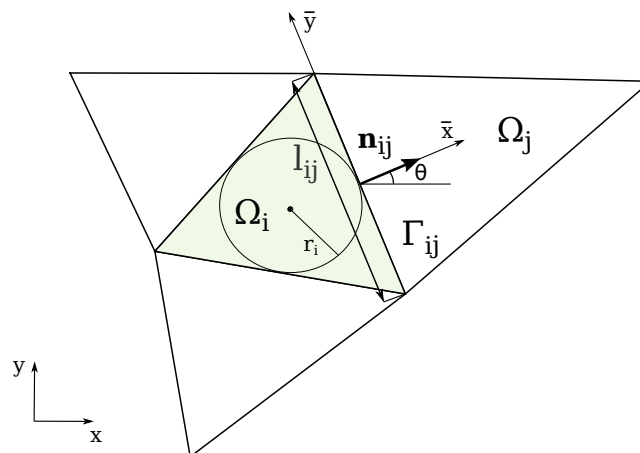
The problem is discretised adopting a finite volume approach over unstructured triangular meshes. A conforming triangulation  $T_\Omega$  of the computational domain  $\Omega \subset \mathbb{R}^2$  by elements  $\Omega_i$  such that  $T_\Omega = \bigcup \Omega_i$ , is assumed. Hereafter we will use the following notation: given a finite volume  $\Omega_i$ ,  $j = 1, 2, 3$  is the set of indexes such that  $\Omega_j$  is a neighbour of  $\Omega_i$ ;  $\Gamma_{ij}$  is the common edge of two neighbour cells  $\Omega_i$  and  $\Omega_j$ , and  $l_{ij}$  its length.  $\mathbf{n}_{ij} = (n_{ij,x}, n_{ij,y})$  is the unit vector which is normal to the edge  $\Gamma_{ij}$  and points toward the cell  $\Omega_j$  (see Figure 2.3). Data are represented by cell averages  $U_i^n$  and the numerical solution sought at time  $t^{n+1} = t^n + \Delta t$ , is denoted by  $U_i^{n+1}$ .



**Figure 2.1** Modeling domain, types of boundary conditions and computational mesh. The flow is from right to left and a side weir (green line) divides the channel into a lower and an upper channel through the weir. External boundary conditions must be provided at  $\Gamma_1$ ,  $\Gamma_2$  and  $\Gamma_3$  while internal boundary conditions can be specified in any place within  $\Omega$



**Figure 2.2** Overview of the numerical model



**Figure 2.3** Element (shaded triangle) of unstructured triangular mesh and used notation.



## 2.3 Numerical solution of Hydrodynamics

### 2.3.1 Vectorial Form of the Governing Equations

For numerical convenience, the system of governing equations (eq. 1.1) is rewritten in vectorial form in terms of the water surface elevation  $H = h + z_B$ . It now reads:

$$\frac{\partial \mathbf{U}}{\partial t} + \frac{\partial \mathbf{F}_x}{\partial x} + \frac{\partial \mathbf{F}_y}{\partial y} = \mathbf{S} \quad (2.1)$$

where the vector of unknowns is

$$\mathbf{U} = \begin{pmatrix} H \\ q_x \\ q_y \end{pmatrix} \quad (2.2)$$

the vector fluxes are

$$\mathbf{F}_x = \begin{pmatrix} q_x \\ uq_x + \frac{1}{2}g(H^2 - 2Hz_b) \\ uq_y \end{pmatrix} ; \quad \mathbf{F}_y = \begin{pmatrix} q_y \\ vq_x \\ vq_y + \frac{1}{2}g(H^2 - 2Hz_b) \end{pmatrix} \quad (2.3)$$

and the vector of source terms is

$$\mathbf{S} = \begin{pmatrix} S_h \\ gHS_x \\ gHS_y \end{pmatrix}. \quad (2.4)$$

The motivation of using  $H$  instead of  $h$  lies in the fact that it is easier to develop numerical schemes which preserve depth positivity and satisfy the well-balanced property.

### 2.3.2 Spatial Discretisation

In order to discretise the system of governing equations, the domain is meshed by a set of triangular elements. The spatial discretization of the conservation equations is carried out by the finite volume method, where the differential equations are integrated over the single elements, i.e. control volumes. The water surface elevation is defined at the element center and is equally distributed over the element.

By integrating the governing system of equations eq. 2.1 in the control volume  $V = [\Omega_i] \times [t^n, t^{n+1}]$ , we obtain

$$\mathbf{U}_i^{n+1} = \mathbf{U}_i^n - \frac{\Delta t}{|\Omega_i|} \sum_{j=1}^3 l_{ij} [\mathbf{F}_{ij}] + \Delta t \mathbf{S}_i. \quad (2.5)$$

### 2.3.3 Flux Estimation

#### 2.3.3.1 Rotational Invariance of the Shallow Water Equations

The flux  $\mathbf{F}_{ij}$  are evaluated taking advantage of the rotational invariance property of the shallow water equations. According to this property the two-dimensional homogeneous shallow water equations satisfy the following equality (Toro, 2009):

$$\mathbf{n}_{ij} \cdot [\mathbf{F}_x(\mathbf{U}), \mathbf{F}_y(\mathbf{U})] = \mathbf{T}^{-1}(\theta) \mathbf{F}_x[\mathbf{T}(\theta)\mathbf{U}] \quad (2.6)$$

where  $\theta$  is the angle between the vector  $\mathbf{n}_{ij}$  and x-axis, measured counter clockwise from the  $x$ -axis (see Figure 2.3) and

$$\mathbf{T}(\theta) = \begin{pmatrix} 1 & 0 & 0 \\ 0 & \cos \theta & \sin \theta \\ 0 & -\sin \theta & \cos \theta \end{pmatrix} \quad (2.7)$$

being

$\mathbf{T}^{-1}(\theta) = \text{inverse of } \mathbf{T}(\theta)$ .

#### 2.3.3.2 Computation of the Flux

The flux  $\mathbf{F}_{ij}$  is obtained at every edge of the finite volume mesh, as the solution of the one-dimensional projected Riemann problem along the normal direction of the two conservation laws eq. 2.1. The computational steps can be summarized as follows:

- First, the vector of conserved variables  $\mathbf{U}$  is transformed into the local coordinate system  $(\bar{x}, \bar{y})$  (see Figure 2.3) at the edge with the operation  $\mathbf{T}(\theta)\mathbf{U}$ .
- A one-dimensional, local Riemann problem is formulated and solved in the normal direction of the edge. From this calculation the new flux vector over the edge  $\mathbf{F}[\mathbf{T}(\theta)\mathbf{U}]$  is defined.
- The flux vector, formulated in the local coordinate system is transformed back to the global coordinates (Cartesian) with  $\mathbf{T}^{-1}\mathbf{F}[\mathbf{T}(\theta)\mathbf{U}]$ . The sum of the fluxes of all edges of an element gives the total fluxes in the  $x$ - and  $y$  directions.

The fluxes are calculated in the normal direction of the element edges. The normal direction of the edge is defined positive from element  $i$  (L) to element  $j$  regarding the edge direction.

#### 2.3.3.3 The HLLC approximated Riemann Solver

The HLLC approximate Riemann solver (Toro, 1994) is a modified HLL (Harten, Lax and van Leer) approximate Riemann solver that includes the shear wave.

The numerical flux at the cell interface is computed as follows:

$$\mathbf{F}_{ij}^{HLLC} = \begin{cases} \mathbf{F}_i & \text{if } 0 \leq S_i, \\ \mathbf{F}_{*i} = \mathbf{F}_i + S_i(\mathbf{U}_{*L} - \mathbf{U}_i) & \text{if } S_i \leq 0 \leq S_*, \\ \mathbf{F}_{*j} = \mathbf{F}_j + S_j(\mathbf{U}_{*R} - \mathbf{U}_j) & \text{if } S_* \leq 0 \leq S_j, \\ \mathbf{F}_j & \text{if } 0 \geq S_j. \end{cases} \quad (2.8)$$

The wave speed velocities are estimated as:

$$S_i = u_i - \sqrt{gh_i}\xi_i; \quad S_j = u_j + \sqrt{gh_j}\xi_j \quad (2.9)$$

where  $\xi_{K=(i,j)}$  is defined as:

$$\xi_K = \begin{cases} \sqrt{\frac{1}{2} \left[ \frac{(h_* + h_K)h_*}{h_K^2} \right]} & \text{if } h_* > h_K, \\ 1 & \text{if } h_* \leq h_K. \end{cases} \quad (2.10)$$

with  $h_*$ , an estimate for the exact solution of the water depth in the star region obtained using the depth positivity condition. It reads as

$$h_* = \frac{1}{2}(h_L + h_R) - \frac{1}{4}(u_R - u_L)(h_L - h_R)/(\sqrt{gh_L} + \sqrt{gh_R}) \quad (2.11)$$

In case of dry-bed conditions, the wave speeds are estimated as the exact dry front speed, i.e.:

$$S_i = \begin{cases} u_i - 2\sqrt{gh_i} & \text{if } h_i = 0, \\ \text{usual estimate} & \text{if } h_i > 0, \end{cases} \quad (2.12)$$

$$S_j = \begin{cases} u_j + 2\sqrt{gh_j} & \text{if } h_j = 0, \\ \text{usual estimate} & \text{if } h_j > 0. \end{cases}$$

And the middle estimated wave speed  $S_*$  corresponds to the front wave speed in case of dry-bed problem.

The expression of the states  $\mathbf{U}_{*i}, \mathbf{U}_{*j}$  and the middle wave speed  $S_*$  can be found in the book of Toro (2009).

### 2.3.4 Numerical Stability

Numerical stability is assured by choosing the time step  $\Delta t$  for time integration such that it obeys the Courant-Friedrichs-Lewy (CFL) condition. In 2-D the Courant number (CFL) can be defined as follows:

$$CFL = \frac{(\sqrt{u^2 + v^2} + c)\Delta t}{r_i} \quad (2.13)$$

where  $r_i$  is the radius of the inscribed circle that defines the element center (Figure 2.3),  $u, v$  are the corresponding velocities of the element and  $c = \sqrt{gh}$ . The HLLC scheme is stable for

$$0 < CFL \leq 1 \quad (2.14)$$

### 2.3.5 Discretisation of Source terms

#### 2.3.5.1 Bed Slope Source Term

The bed slope source term (eq. 1.2) is discretized using the robust modified-state approach proposed by Duran et al. (2013). The discretization presents a motionless steady states-preserving scheme:

$$\mathbf{S}_{b,i} = \sum_{j=1}^m l_{ij} \mathbf{S}_{b,ij} = \sum_{j=1}^m l_{ij} \begin{pmatrix} 0 \\ gH_{ij}^*(z_i - \bar{z}_{ij}) \vec{\mathbf{n}}_{ij} \end{pmatrix} \quad (2.15)$$

where  $\bar{z}_{ij} = \check{z}_{ij} - \Delta_{ij}$  with  $\check{z}_{ij} = \max(z_{bi}, z_{bj})$  the maximum bed elevation between cells  $i$  and  $j$  and  $\Delta_{ij} = \max(0, \check{z}_{ij} - H_i)$ .  $H_{ij}^*$  is the approximated value of the water surface elevation  $H$  at the cell interface  $\Gamma_{ij}$ .

#### 2.3.5.2 Friction Source Term

We handle the inhomogeneous character of system eq. 1.1 due to the presence of frictional source terms by adopting a robust splitting technique Toro (2001). We initially consider the initial value problem (IVP)

$$\left. \begin{array}{l} PDE : \mathcal{A}(\mathbf{U}) = \mathcal{S}(\mathbf{U}) \\ IC : \mathbf{U}(x, y, 0) = \mathbf{U}_i^n \end{array} \right\} \text{IVP} .$$

where  $\mathcal{A}$  represents the advective operator

$$\mathcal{A}(\mathbf{U}) = \frac{\partial \mathbf{U}}{\partial t} + \frac{\partial \mathbf{F}_x}{\partial x} + \frac{\partial \mathbf{F}_y}{\partial y} = \mathbf{0} ,$$

and  $\mathcal{S}$  represents the frictional source term operator.

The numerical solution is then obtained by subsequently integrating *two* initial value problems (IVPs):

$$\left. \begin{array}{l} ODEs : \frac{d\mathbf{U}}{dt} = \mathcal{S}(\mathbf{U}) \\ ICs : \mathbf{U}(x, y, 0) = \mathbf{U}_i^n \end{array} \right\} \xrightarrow{\Delta t} \bar{\mathbf{U}}_i \quad \text{IVP1} ,$$

$$\left. \begin{array}{l} PDEs : \mathcal{A}(\mathbf{U}) = 0 \\ ICs : \mathbf{U}(x, y, 0) = \bar{\mathbf{U}}_i \end{array} \right\} \xrightarrow{\Delta t} \mathbf{U}_i^{n+1} \quad \text{IVP2} ,$$

The initial condition (IC) for IVP1 is  $\mathbf{U}_i^n$ , corresponding to the initial condition of the full problem IVP. The solution of IVP1 is obtained solving a system of ordinary differential equations (ODEs) after integration by a time step  $\Delta t$  and is denoted by  $\bar{\mathbf{U}}_i$ . IVP2 is then integrated by a time step  $\Delta t$ , with initial condition given by the solution of IVP1  $\bar{\mathbf{U}}_i$ . The solution of IVP2  $\mathbf{U}_i^{n+1}$  is obtained solving an hyperbolic homogeneous system of partial differential equations (PDEs) and represents the approximate solution of the full problem

IVP.\ Since we adopt an implicit second-order Runge-Kutta method for solving the ODEs systems IVP1 and an explicit finite volume method for solving IVP2, the integration time step  $\Delta t$  is determined accordingly with the *CFL* stability condition for IVP2.

### 2.3.5.3 External Source Term

An external source is defined as specific mass flux  $\delta$  (m/s), uniformly distributed over a number of elements of the domain with a specific surface area. The external source can either be specified as discharge ( $m^3/s$ ) or precipitation intensity (mm/h) for a specific region of the domain. The external source value is divided among the cells composing the region and converted to cell specific mass flux  $\delta_i$ . The volume allocated is characterized by different behaviors:

$$\begin{aligned}
 \text{Exact:} \quad & S_{h,i} = \delta_i \\
 \text{Available:} \quad & S_{h,i} = \delta_i && \text{if } \delta_i \cdot \Delta t > 0 \\
 & S_{h,i} = \max(\delta_i, -h_i) && \text{if } \delta_i \cdot \Delta t < 0 \\
 \text{Infinity:} \quad & S_{h,i} = \delta_i && \text{if } \delta_i \cdot \Delta t > 0 \\
 & S_{h,i} = -h_i && \text{if } \delta_i \cdot \Delta t < 0
 \end{aligned} \tag{2.16}$$

Where  $h_i$  is the water depth of the element  $i$ . The external source volume is added to the initial water volume.

$$h_i^{t+1} = h_i^t + S_{h,i} \cdot \Delta t \tag{2.17}$$

## 2.3.6 Solution Procedure

The numerical solution procedure of BASEMENT explains how the discretised shallow water equation (eq. 1.1) is solved inside a defined time step  $\Delta t$  through a sequence of loops over the edges or cells (Figure 2.4).

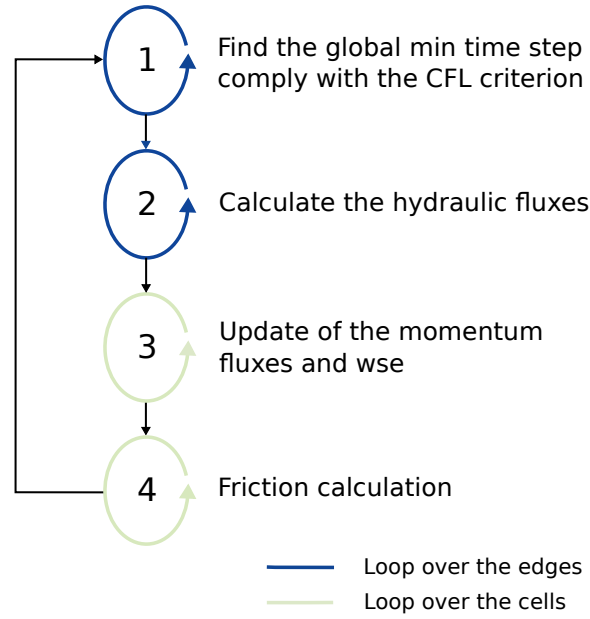
First, a global minimum time step  $\Delta t$  should be defined. Then, the hydraulic fluxes (liquid mass, x-momentum and y-momentum) are calculated with a HLLC Riemann solver at the element edges according to the initial states of the left and right cells (Section 2.3.3). Subsequently, the hydraulic state variables i.e. cell centered quantities are updated and finally, the friction (source term) is calculated using an implicit scheme, thus looping twice over the cell.

## 2.4 Numerical solution of Morphodynamics

### 2.4.1 Numerical solution of the Exner equation

#### 2.4.1.1 Fundamentals

The Exner equation assures that sediment mass is conserved in the bed and is used to model the riverbed time evolution. The rate of sediment transport is determined using a closure equation. The cell centered finite volume approach is used to discretise the Exner equation and in particular the HLL approximate Riemann solver with a wave speed



**Figure 2.4** Numerical solution procedure of hydrodynamic simulation for each time step  $\Delta t$

estimator defined in Soares-Frazão and Zech (2011) is adopted. The shallow water and the Exner equations create a system of equations that is solved in a decoupled way (Figure 2.2). This approach makes the assumption that the bed load flux is much slower than the water flow velocity (Soares-Frazão and Zech, 2011).

#### 2.4.1.2 Spatial discretization

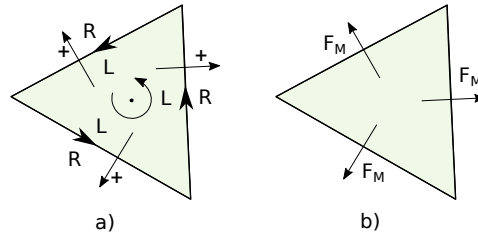
In order to discretise the the Exner equation, we use the same unstructured mesh adopted for the hydrodynamic part and the same finite volume approach. As a consequence, the bed level  $z_B$  is defined at the element center and is equally distributed over the element.

By integrating the Exner equation in the control volume  $V = [\Omega_i] \times [t^n, t^{n+1}]$ , we obtain

$$z_{Bi}^{n+1} = z_{Bi}^n - \frac{\Delta t}{|\Omega_i|} \sum_{j=1}^3 [q_{Bij} \cdot l_{ij}] + \Delta t \mathbf{S}_i . \quad (2.18)$$

The calculation of the sediment flux at the cell interface proceeds as follows:

1. loop over the cells and calculate:
  1. correction terms for the bed-load vector directions (if selected by the user), therefore:
    - calculation of the local bed slope, for the lateral-transport correction (see section Section 1.2.2.4.1)
    - calculation of the local curvature of the flow field, for the spiral flow correction (see section Section 1.2.2.4.2)



**Figure 2.5** a) Sign convention for the edge direction: counterclockwise b) Positive morphological flux direction at edges: from left (L) to right (R)

2. loop over the cell interfaces and:

1. calculate the flux projection along the normal vector ( $n_{ij,x}, n_{ij,y}$  of edge  $\Gamma_{ij}$ , i.e.:  $q_{Bi,n} = q_{Bi,x} \cdot n_{ij,x} + q_{Bi,y} \cdot n_{ij,y}$  and  $q_{Bj,n} = q_{Bj,x} \cdot n_{ij,x} + q_{Bj,y} \cdot n_{ij,y}$  with  $j=1,2,3$ )
  2. compute the flux at the interface using the approximate HLL Riemann solver at the interface
- Evaluate the wave speeds at the interface. this is obtained following the approach proposed by Soares-Frazão and Zech (2011), for which the wave speeds can be calculated as an approximation of the smallest eigenvalue of the system of governing equations, i.e. Shallow water and Exner. They read:

$$\lambda_1 = 1/2(u_n - c - \sqrt{(u_n - c)^2 + 4a_2c^2}) \quad (2.19)$$

$$\lambda_2 = 1/2(u_n - c + \sqrt{(u_n - c)^2 + 4a_2c^2}) \quad (2.20)$$

where  $u_n = u \cdot n_{ij,x} + v \cdot n_{ij,y}$ ,  $c = \sqrt{gh}$  and  $a_2 = \frac{\partial q_{b,n}}{\partial q_n}$  which is the derivative of the bed load discharge in the normal flow direction with respect to the hydraulic flux direction. Then the speeds estimate are

$$S^- = \min(\lambda_{1,L}, \lambda_{1,R}) \quad (2.21)$$

and

$$S^+ = \max(\lambda_{2,L}, \lambda_{2,R}) \quad (2.22)$$

- Flux calculation:

$$q_{Bij}^{HLL} = \begin{cases} q_{Bi,n} & \text{if } S^- \geq 0, \\ \frac{q_{Bi,n}S^+ - q_{Bj,n}S^- + S^-S^+(z_{Bj} - z_{Bi})}{S^+ - S^-} & \text{if } S^- < 0 < S^+, \\ q_{Bj,n} & \text{if } S^+ \leq 0. \end{cases} \quad (2.23)$$

The convention for the positive bed load flux direction is the same as for the hydrodynamic flux and is presented on Figure 2.5

### 2.4.1.3 Discretization of External Source Term

The source term  $S_{l_b}$  describes a local input or removal of sediment mass into a river.

An external source is defined as specific mass flux  $\delta$  (m/s), uniformly distributed over a number of elements of the domain (region) with a specific surface area. The external source can be specified as the total volume flux ( $m^3/s$ ) for a specific region of the domain. The external source value is divided among the cells composing the region and converted to cell specific mass flux  $\delta_i$ . The volume allocated is characterized by different behaviors:

$$\begin{aligned}
 \text{Exact:} \quad & S_{b,i} = \delta_i \\
 \text{Available:} \quad & S_{b,i} = \delta_i && \text{if } \delta_i \cdot \Delta t > 0 \\
 & S_{b,i} = \max(\delta_i, -(z_{Fix} - z_i)) && \text{if } \delta_i \cdot \Delta t < 0 \\
 \text{Infinity:} \quad & S_{b,i} = \delta_i && \text{if } \delta_i \cdot \Delta t > 0 \\
 & S_{b,i} = -(z_{Fix} - z_i) && \text{if } \delta_i \cdot \Delta t < 0
 \end{aligned} \tag{2.24}$$

Where  $z_i$  is the bottom elevation and  $z_{Fix}$  the fixed bed elevation of the element  $i$ . The external source volume is added to the initial bottom elevation of element  $i$ .

$$z_i^{t+1} = z_i^t + S_{b,i} \cdot \Delta t \tag{2.25}$$

## 2.4.2 Solution procedure

The numerical solution procedure of BASEMENT explains how the discretised Exner equation (eq. 1.15) is solved through a sequence of loops over the edges or cells (Figure 2.6).

In the numerical simulation, the hydrodynamic and morphodynamic simulations are performed in a decoupled way. The morphodynamic simulation is executed after the hydrodynamic simulation, using the hydraulic fluxes to calculate the morphological fluxes. This approach assumes that the sediment transport is much slower than the water velocity, which is an accurate assumption for the numerical modelling of slow flood with morphological changes occurring over a long period (Soares-Frazão and Zech, 2011). The numerical solution procedure of Figure 2.6 is performed after the step 4 of Figure 2.4 inside the same time step  $\Delta t$ .

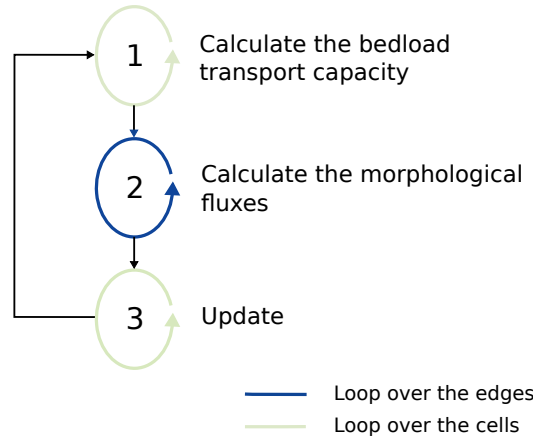
The numerical solution of the Exner equation starts with a loop over the cells in order to find the bedload transport capacity  $q_b$  with a potential correction due to a curvature effect or lateral bed slope. Then, the morphological fluxes  $F_M$  are calculated at the element edges and finally, the bed elevation  $z_b$  is updated over the cells.

## 2.5 Numerical solution of Advection-Diffusion equation

### 2.5.1 Numerical solution of the advective part

The scalar advection equation assures that scalar masses are conserved in the flow column when moving with the surrounding fluid. The same cell cell-centered finite volume approach used for the SWE is employed in scalar advection, namely the HLLC approximate Riemann solver. In order to discretize the scalar advection equation, the same unstructured mesh adopted for the hydrodynamic and morphodynamic parts is used, as is the same finite





**Figure 2.6** Numerical solution procedure of morphodynamic simulation for each time step  $\Delta t$

volume approach. As a consequence, the scalar concentration  $\phi_s$  is defined at the element center and is equally distributed over the element. This numerical approach is applied to turbulent kinetic energy quantities ( $\kappa, \epsilon$ ), suspended sediment and tracers.

By integrating the scalar advection equation

$$\frac{\partial}{\partial t} \phi_k h + \frac{\partial}{\partial x_i} \left( \phi_k h u_i - h D_{s_{ij}} \frac{\partial \phi_k}{\partial x_j} \right) = S_{\phi_k} \quad (2.26)$$

in the control volume  $V_i = [\Omega_i] \times [t^n, t^{n+1}]$  for a generic quantity defined by  $\phi_k$ , we obtain:

$$q_{\phi_k i}^{n+1} = q_{\phi_k i}^n - \frac{\Delta t}{|\Omega_i|} \sum_{j=1}^3 [f_{\phi_k, ij} \cdot l_{ij}] + \Delta t S_{\phi_k, i} \quad (2.27)$$

where  $f_{s_{ij}}$  is the intercell scalar advective flux. The scalar advective fluxes at the cell interface are defined by:

$$f_{\phi_k, ij} = \frac{q_{\phi_k i}}{h} q_m \quad \text{if } q_m > 0 \quad (2.28)$$

$$f_{\phi_k, ij} = \frac{q_{\phi_k j}}{h} q_m \quad \text{if } q_m < 0 \quad (2.29)$$

where  $q_m$  is the hydrodynamic mass flux. This approach significantly reduces the numerical effort involved in the computation of the numerical fluxes for each species, without sacrificing accuracy or stability.

### 2.5.2 Numerical solution of the diffusive part

The diffusion equation ensures that quantities are conserved in the flow column due when turbulent or molecular exchanges originate some form of mixing. By integrating the scalar diffusion equation

$$\frac{\partial}{\partial t} \phi_k h - \frac{\partial}{\partial x_i} \left( h D_{s_{ij}} \frac{\partial \phi_k}{\partial x_j} \right) = 0 \quad (2.30)$$

in the control volume  $V_i = [\Omega_i] \times [t^n, t^{n+1}]$  we obtain:

$$q_{\phi_k i}^{n+1} = q_{\phi_k i}^n - \frac{\Delta t}{|\Omega_i|} \sum_{j=1}^3 [d_{\phi_k, ij} \cdot l_{ij}] \quad (2.31)$$

where  $d_{\phi_k, ij}$  is the intercell scalar diffusive flux. To compute the solution of the diffusive terms, the SVT solver Vanzo et al. (2016) is employed. It involves the augmentation of the existing system with a set of new conservation quantities and equations for each diffusive specie  $\phi_k$ . The role of these new variables is to allow a relaxation of the diffusive part, such as

$$\lim_{\zeta \rightarrow 0} \psi_x^{\phi_k} = \partial_x \phi_k, \quad \lim_{\zeta \rightarrow 0} \psi_y^{\phi_k} = \partial_y \phi_k \quad (2.32)$$

where  $\psi$  denotes a relaxation variable, subject to the following relations

$$\partial_t \psi_x^{\phi_k} - \partial_x \frac{\phi_k}{\zeta} = -\frac{\psi_x^{\phi_k}}{\zeta}, \quad \partial_t \psi_y^{\phi_k} - \partial_y \frac{\phi_k}{\zeta} = -\frac{\psi_y^{\phi_k}}{\zeta} \quad (2.33)$$

which are formally denoted as the relaxation sub-system. The main purpose of this relaxation is to preserve the hyperbolicity of the numerical scheme, as demonstrated by Vanzo et al. (2016). To construct the diffusive numerical fluxes, the following Riemann problem is considered along some edge normal as

$$\begin{cases} \partial_t \Phi_k + \partial_\xi D_{n_{ij}} = \mathbf{0}, & \xi \in \mathbb{R}, t > 0, \\ \Phi_k(\xi, 0) = \begin{cases} \Phi_{k_i}^n & \text{if } \xi < 0, \\ \Phi_{k_j}^n & \text{if } \xi > 0, \end{cases} \end{cases} \quad (2.34)$$

where  $\Phi_k$  is the conserved variables vector and  $D_n$  is the diffusive numerical flux vector. The structure of this Riemann problem is composed of three unique waves (from a total of six). One of these is always a stationary contact discontinuity and the two remaining waves are symmetrical, with the same propagation speed in opposite directions. The problem thus becomes characterized by a single intermediate state  $\Phi_k^*$  where the fluxes are directly evaluated. The exact expression for the intermediate state  $\Phi_k^*$  and numerical  $D(\Phi_k^*)_n$  can be found in Vanzo et al. (2016).

### 2.5.3 Discretization of external source terms

The source term  $S_{\phi_k}$  describes a local input or removal of scalar mass into a river. An external source is defined as specific mass flux  $\delta$  (m/s), uniformly distributed over a number of elements of the domain (region) with a specific surface area. The external source can be specified as the total volume flux ( $m^3/s$ ) for a specific region of the domain. The external source value is divided among the cells composing the region and converted to cell specific mass flux  $\delta_i$ . The volume allocated is characterized by different behaviors:

$$\begin{array}{ll}
\text{Exact:} & S_{\phi_{k,i}} = \delta_i \\
\text{Available:} & S_{\phi_{k,i}} = \delta_i \quad \text{if } \delta_{\phi_k} \cdot \Delta t > 0 \\
& S_{\phi_{k,i}} = \max(\delta_i, -q_{\phi_{k,i}}) \quad \text{if } \delta_i \cdot \Delta t < 0 \\
\text{Infinity:} & S_{\phi_{k,i}} = \delta_i \quad \text{if } \delta_i \cdot \Delta t > 0 \\
& S_{\phi_{k,i}} = -q_{\phi_{k,i}} \quad \text{if } \delta_i \cdot \Delta t < 0
\end{array} \tag{2.35}$$

The external source volume is then added to the initial value of element  $i$  through a first-order Euler approach.

$$q_{\phi_{k,i}}^{t+1} = q_{\phi_{k,i}}^t + S_{\phi_{k,i}} \cdot \Delta t \tag{2.36}$$

### 2.5.4 Solution procedure

The numerical solution for all modules involving scalar advection and diffusive terms – turbulence, suspended sediment and tracers – is obtained as follows:

1. Loop over the cell interfaces and compute the advective flux at each interface:
  - Retrieve the hydrodynamic fluxes  $\mathbf{F}_{ij}$  at the interface between cells  $i$  and  $j$  and extract the first component of the flux vector (mass fluxes) to the variable  $q_m$ .
  - For each specie, perform the advective flux calculation through the simplified HLLC solver described above
2. Loop over the cell interfaces and compute the diffusive flux at each interface:
  - Retrieve the hydrodynamic and scalar quantities  $\mathbf{Q}_h$  and  $\Phi_k$  at the interface between cells  $i$  and  $j$  perform the flux calculation using the SVT solver
3. Define the diffusion specific timestep and update the global minimum time step  $\Delta t$
4. Loop over the cells and update the conserved quantities with the fluxes at each of its interfaces:
  - Retrieve and add the advective fluxes  $f_{\phi_{ij}}$  and diffusive fluxes  $d_{\phi_{ij}}$  at each of the three interfaces.
  - For each specie, perform the update as prescribed in Equations 2.27, 2.31, and 2.36.
5. Add the external source terms



# 3

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**BASIC SIMULATION ENVIRONMENT  
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# **APPENDIX**

**VERSION 4.0.2  
OCTOBER 2023**



**BASEMENT**





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---

Lao Word Break Dictionary Data (laodict.txt)

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```

---

Burmese Word Break Dictionary Data (burmesedict.txt)

```

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---

File: aclocal.m4 (only for ICU4C)  
Section: pkg.m4 - Macros to locate and utilise pkg-config.

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---

File: config.guess (only for ICU4C)

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#### PCRE2 JUST-IN-TIME COMPILATION SUPPORT

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#### STACK-LESS JUST-IN-TIME COMPILER

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```
[program/widget] is based in part on the work of
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```

```
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### Vtk

/\*\*\*\*\*

Program: Visualization Toolkit  
Module: Copyright.txt

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## Zlib

### ZLIB DATA COMPRESSION LIBRARY

zlib 1.2.12 is a general purpose data compression library. All the code is thread safe. The data format used by the zlib library is described by RFCs (Request for Comments) 1950 to 1952 in the files <http://tools.ietf.org/html/rfc1950> (zlib format), [rfc1951](http://tools.ietf.org/html/rfc1951) (deflate format) and [rfc1952](http://tools.ietf.org/html/rfc1952) (gzip format).

All functions of the compression library are documented in the file `zlib.h` (volunteer to write man pages welcome, contact [zlib@gzip.org](mailto:zlib@gzip.org)). A usage example of the library is given in the file `test/example.c` which also tests that the library is working correctly. Another example is given in the file `test/minigzip.c`. The compression library itself is composed of all source files in the root directory.

To compile all files and run the test program, follow the instructions given at the top of `Makefile.in`. In short `./configure; make test`, and if that goes well, `make install` should work for most flavors of Unix. For Windows, use



one of the special makefiles in win32/ or contrib/vstudio/ . For VMS, use make\_vms.com.

Questions about zlib should be sent to <zlib@gzip.org>, or to Gilles Vollant <info@winimage.com> for the Windows DLL version. The zlib home page is <http://zlib.net/> . Before reporting a problem, please check this site to verify that you have the latest version of zlib; otherwise get the latest version and check whether the problem still exists or not.

PLEASE read the zlib FAQ [http://zlib.net/zlib\\_faq.html](http://zlib.net/zlib_faq.html) before asking for help.

Mark Nelson <markn@ieee.org> wrote an article about zlib for the Jan. 1997 issue of Dr. Dobbs's Journal; a copy of the article is available at <http://marknelson.us/1997/01/01/zlib-engine/> .

The changes made in version 1.2.12 are documented in the file ChangeLog.

Unsupported third party contributions are provided in directory contrib/ .

zlib is available in Java using the java.util.zip package, documented at <http://java.sun.com/developer/technicalArticles/Programming/compression/> .

A Perl interface to zlib written by Paul Marquess <pmqs@cpan.org> is available at CPAN (Comprehensive Perl Archive Network) sites, including <http://search.cpan.org/~pmqs/IO-Compress-Zlib/> .

A Python interface to zlib written by A.M. Kuchling <amk@amk.ca> is available in Python 1.5 and later versions, see <http://docs.python.org/library/zlib.html> .

zlib is built into tcl: <http://wiki.tcl.tk/4610> .

An experimental package to read and write files in .zip format, written on top of zlib by Gilles Vollant <info@winimage.com>, is available in the contrib/minizip directory of zlib.

Notes for some targets:

- For Windows DLL versions, please see win32/DLL\_FAQ.txt
- For 64-bit Irix, deflate.c must be compiled without any optimization. With -O, one libpng test fails. The test works in 32 bit mode (with the -n32 compiler flag). The compiler bug has been reported to SGI.
- zlib doesn't work with gcc 2.6.3 on a DEC 3000/300LX under OSF/1 2.1 it works when compiled with cc.
- On Digital Unix 4.0D (formerly OSF/1) on AlphaServer, the cc option -std1 is necessary to get gzprintf working correctly. This is done by configure.
- zlib doesn't work on HP-UX 9.05 with some versions of /bin/cc. It works with other compilers. Use "make test" to check your compiler.
- gzdopen is not supported on RISCOS or BEOS.
- For PalmOs, see <http://palmzlib.sourceforge.net/>

Acknowledgments:

The deflate format used by zlib was defined by Phil Katz. The deflate and zlib specifications were written by L. Peter Deutsch. Thanks to all the people who reported problems and suggested various improvements in zlib; they are too numerous to cite here.

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