iRIC Software

Changing River Science

Nays2DH
Solver Manual

Contents

I. Overview

I.1 What is Nays2DH?

Nays2DH is a computational model for simulating horizontal two-dimensional (2D) flow, sediment transport, morphological changes of bed and banks in rivers. Although iRIC has provided several 2D solvers, such as Nays2D, Morpho2D, FaSTMECH, etc., we believe that the users sometimes may confuse which solver is preferable for their own case. Therefore, we decided to combine Nays2D and Morpho2D to provide a more powerful and user friendly tool for iRIC users, we called it Nays2DH.

Nays2D, developed by Dr. Yasuyuki Shimizu in Hokkaido University in Japan, is a plane 2D solver for calculating flow, sediment transport, bed evolution and bank erosion in rivers. By joining many developers to Nays2D project, several functions, for instance, river confluence model, mixture grain size model and Hot start function, have been added. Nays2D is attached to iRIC and RIC-Nays which is a predecessor project of iRIC. Nays2D includes several options for simulating river flows such as an unsteady vortex generation in open channel flows and river morphodynamics. River morphodynamics includes the initiation and development of free bars in rivers and the interaction between free bars and forced bars in meandering channels. In addition, Nays2D has been applied to several practical applications: bed evolution process in rivers affected by trees and vegetation, calculation and prediction of inundation on floodplains, sedimentation in river confluences, analysis of bank erosion and flood disasters.

Morpho2D developed by Dr. Hiroshi Takebayashi is a solver to simulate the two-dimensional morphodynamical changes in rivers. Initially, it was attached in the RIC-Nays as ‘Mixture grain size model’. Since iRIC released version 1, this solver has been reformed as Morpho2D. Morpho2D includes several possibilities for simulating the morphological changes of river bed with uniform and mixture sediment, and simulating the development of free bars with sorting of sediment particles on the river bed. Morpho2D has also been applied to several real world applications in river engineering, such as analysis the bed evolution under vegetation effects, sediment transport and bed evolution with the non-erosional bed (e.g. bedrocks and fixed beds).

Both solvers have their own advantages, but they include a common part as they are both 2D models. We developed a new and more powerful solver by combining the functions of these models. In this version, the user can choose a sediment transport model based on the functions which were implemented in the both solvers. The user also can combine a river confluence model, bank erosion model, bedload-suspended load simulation in mixture sediment, bedload layer model and fixed bed model, and also able to change the sediment supply rate from the upstream end. However, there are some uncoupled parts in the current version. The combination of the functions explained above may not work in some points, and the seepage flow model used in Morpho2D is not implemented. These points will be improved in the near future.

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I.2 Features of the hydrodynamic model

(1) As a coordinate system, a general curvilinear coordinate system is adopted, allowing direct consideration of complex boundaries and riverbed shapes.

(2) Calculations involving confluences of a main channel and tributaries can be performed.

(3) For the finite differencing applied to the advection terms in the momentum equations, the user can select from among [Upwind difference method (first order)] and [CIP method][*1].

(4) For the turbulent field calculation method, the user can select from among [Constant eddy viscosity], [Zero-equation model] and [k-ε model].

(5) Various settings are possible for boundary conditions of the upstream and downstream ends, including periodic boundary conditions, downstream end water surface elevation setting and upstream end velocity setting. This makes it convenient to set boundary conditions from limited observation data.

(6) For setting the initial water surface profile, the user can select from among [Constant slope], [Line], [Uniform flow calculation] and [Non-uniform flow calculation].

(7) The bottom friction evaluation method is set by using Manning's roughness parameter. This parameter can be set to each computational cell.

(8) Any obstacle within the calculation target area can be taken into account on a calculation-cell basis. For each calculation cell, a flag can be determined to define an obstacle. By this means, river structures such as bridge piers can be easily incorporated in calculation.

(9) The effect of vegetation on the flow can be introduced as a drag force. The user can set the density of vegetation and the height of vegetation in each computational cell.

I.3 Features of the sediment transport model and riverbed deformation model

(1) The model allows the user to select between performing flow regime calculation only and performing riverbed deformation calculation along with flow regime calculation.

(2) For sediment transport, the user can select between bed load only and bed load + suspended load.

(3) For grain size distribution, the user can select between uniform and non-uniform. When non-uniform grain size is selected, changes in the grain size distribution in the depth direction during calculation can be stored on a multilayer basis.

(4) The sediment supply rate at the upstream boundary condition can be adjusted. The user can set the ratio of “sediment transport supplying rate from upstream end” to “equilibrium sediment transport rate”.

(5) The sediment transport and mass balance of it on the fixed bed can be calculated by introducing the bed-load layer model. In addition, the user can set either the elevation of river bed or fixed bed.

[*1] This is a high-order finite-difference method. By using a cubic polynomial as an interpolation function, numeric diffusion is reduced, thus enabling high-precision local interpolation.

(6) The user can select a total bedload transport formula from among [Meyer-Peter and Muller formula] and [Ashida and Michiue formula]. For calculating the bedload transport vector, the user can select [Watanabe formula] or [Ashida, Egashira and Liu formula].

(7) The effect of secondary flows on the bedload transport can be calculated from either an equilibrium model or a non-equilibrium model (use a depth-averaged vorticity equation).

(8) The user can select an upward flux of suspended sediment from river beds from among [Lane-Kalinske formula] and [Itakura and Kishi formula].

(9) Morphological factor which is an acceleration parameter of bed evolution can be set.

(10) The solver incorporates a slope collapse model. An unrealistic steep slope can sometimes occur in the bed evolution, if it is simulated by using only Exner equation. In this model, if the angle of riverbed exceeds the critical angle, which is a user parameter, the bed is instantaneously corrected to adjust the bed angle (to be not larger than the critical angle). By applying this model, the user can treat the morphologic evolution processes such as bank erosion, which cannot be well captured by using only continuity equation of riverbed.

(11) Bank erosion takes into account of the angle of repose. More specifically, when the riverbed slope exceeds the angle of repose as riverbed deformation proceeds, adjustment is made by exchanging sediment with surrounding cells such that the bed angle slope decreases and fall lower than the angle of repose. In addition, if the channel widens from bank erosion, calculation grids are automatically moved accordingly.

I.4 Other features

(1) HotStart is provided (starts the calculation from where the previous calculation left off, based on the previous calculation results).

(2) The user can use parallel computing by OpenMP for reducing the computational time. The user can specify the number of the CPU's used for the computation.

II. Basic equations

II.1 Basic flow equations

II.1.1 Basic equations in orthogonal coordinates

The basic equations in an orthogonal coordinate system (x, y) are as follows:
[Continuity equation]

$$\frac{\partial h}{\partial t}+\frac{\partial (hu)}{\partial x}+\frac{\partial (hv)}{\partial y}=0 \tag{1}$$

[Momentum equations in x and y-directions]

$$\frac{\partial (uh)}{\partial t}+\frac{\partial (hu^2)}{\partial x}+\frac{\partial (huv)}{\partial y}=-gh\frac{\partial H}{\partial x}-\frac{\tau_x}{\rho}+D^x+\frac{F_x}{\rho} \tag{2}$$

$$\frac{\partial (vh)}{\partial t}+\frac{\partial (huv)}{\partial x}+\frac{\partial (hv^2)}{\partial y}=-gh\frac{\partial H}{\partial y}-\frac{\tau_y}{\rho}+D^y+\frac{F_y}{\rho} \tag{3}$$

in which

$$\frac{\tau_x}{\rho}=C_f u\sqrt{u^2+v^2} \qquad \frac{\tau_y}{\rho}=C_f v\sqrt{u^2+v^2} \tag{4}$$

$$D^x=\frac{\partial}{\partial x}\left[\nu_t h\frac{\partial u}{\partial x}\right]+\frac{\partial}{\partial y}\left[\nu_t h\frac{\partial u}{\partial y}\right] \tag{5}$$

$$D^y=\frac{\partial}{\partial x}\left[\nu_t h\frac{\partial v}{\partial x}\right]+\frac{\partial}{\partial y}\left[\nu_t h\frac{\partial v}{\partial y}\right] \tag{6}$$

$$\frac{F_x}{\rho}=\frac{1}{2}C_D a_s h_v u\sqrt{u^2+v^2} \qquad \frac{F_y}{\rho}=\frac{1}{2}C_D a_s h_v v\sqrt{u^2+v^2} \tag{7}$$

where h is water depth, t is time, u and v are depth-averaged velocities in x- and y-directions, g is gravitational acceleration, H is water depth, τ_x and τ_y are the components of the shear stress of river bed in x- and y-directions, F_x and F_y are components of drag force by vegetation in the x- and y-direction, C_f is the drag coefficient of the bed shear stress, ν_t is eddy viscosity coefficient, C_D is drag coefficient of vegetation, a_s is the area of interception by vegetation per unit volume, and h_v is minimum value of water depth and height of vegetation.

II.1.2 Transformation into general curvilinear coordinates

Next, basic equations of two-dimensional plane flow at orthogonal coordinates are transformed into general coordinates (ξ, η). By transforming the equations into general coordinates, it becomes possible to set a calculation mesh of any shape (fitted with the boundary conditions). The following describes how the equations can be transformed from orthogonal coordinates into general curvilinear coordinates:

$$\frac{\partial}{\partial x} = \frac{\partial \xi}{\partial x}\frac{\partial}{\partial \xi} + \frac{\partial \eta}{\partial x}\frac{\partial}{\partial \eta} \tag{8}$$

$$\frac{\partial}{\partial y} = \frac{\partial \xi}{\partial y}\frac{\partial}{\partial \xi} + \frac{\partial \eta}{\partial y}\frac{\partial}{\partial \eta} \tag{9}$$

or,

$$\begin{pmatrix} \dfrac{\partial}{\partial x} \\ \dfrac{\partial}{\partial y} \end{pmatrix} = \begin{pmatrix} \xi_x & \eta_x \\ \xi_y & \eta_y \end{pmatrix} \begin{pmatrix} \dfrac{\partial}{\partial \xi} \\ \dfrac{\partial}{\partial \eta} \end{pmatrix} \tag{10}$$

where,

$$\xi_x = \frac{\partial \xi}{\partial x}, \quad \xi_y = \frac{\partial \xi}{\partial y}, \quad \eta_x = \frac{\partial \eta}{\partial x}, \quad \eta_y = \frac{\partial \eta}{\partial y} \tag{11}$$

likewise,

$$\frac{\partial}{\partial \xi} = \frac{\partial x}{\partial \xi}\frac{\partial}{\partial x} + \frac{\partial y}{\partial \xi}\frac{\partial}{\partial y} \tag{12}$$

$$\frac{\partial}{\partial \eta} = \frac{\partial x}{\partial \eta}\frac{\partial}{\partial x} + \frac{\partial y}{\partial \eta}\frac{\partial}{\partial y} \tag{13}$$

or,

$$\begin{pmatrix} \dfrac{\partial}{\partial \xi} \\ \dfrac{\partial}{\partial \eta} \end{pmatrix} = \begin{pmatrix} x_\xi & y_\xi \\ x_\eta & y_\eta \end{pmatrix} \begin{pmatrix} \dfrac{\partial}{\partial x} \\ \dfrac{\partial}{\partial y} \end{pmatrix} \tag{14}$$

where,

$$x_\xi = \frac{\partial x}{\partial \xi}, \quad x_\eta = \frac{\partial x}{\partial \eta}, \quad y_\xi = \frac{\partial y}{\partial \xi}, \quad y_\eta = \frac{\partial y}{\partial \eta} \tag{15}$$

Hence

$$\begin{pmatrix} \dfrac{\partial}{\partial \xi} \\ \dfrac{\partial}{\partial \eta} \end{pmatrix} = \frac{1}{\xi_x \eta_y - \xi_y \eta_x} \begin{pmatrix} \eta_y & -\eta_x \\ -\xi_y & \xi_x \end{pmatrix} \begin{pmatrix} \dfrac{\partial}{\partial x} \\ \dfrac{\partial}{\partial y} \end{pmatrix} = \begin{pmatrix} x_\xi & y_\xi \\ x_\eta & y_\eta \end{pmatrix} \begin{pmatrix} \dfrac{\partial}{\partial x} \\ \dfrac{\partial}{\partial y} \end{pmatrix} \tag{16}$$

Here, assuming $J = \xi_x \eta_y - \xi_y \eta_x$ we obtain

$$\frac{1}{J} \begin{pmatrix} \eta_y & -\eta_x \\ -\xi_y & \xi_x \end{pmatrix} = \begin{pmatrix} x_\xi & y_\xi \\ x_\eta & y_\eta \end{pmatrix} \tag{17}$$

Therefore

$$x_\xi = \frac{1}{J}\eta_y, \quad y_\xi = -\frac{1}{J}\eta_x, \quad x_\eta = -\frac{1}{J}\xi_y, \quad y_\eta = \frac{1}{J}\xi_x \tag{18}$$

or,

$$\eta_y = Jx_\xi, \quad \eta_x = -Jy_\xi, \quad \xi_y = -Jx_\eta, \quad \xi_x = Jy_\eta \tag{19}$$

$$J = \xi_x \eta_y - \xi_y \eta_x = J^2 \left(x_\xi y_\eta - x_\eta y_\xi \right) \tag{20}$$

Thus we obtain

$$J = \frac{1}{x_\xi y_\eta - x_\eta y_\xi} \tag{21}$$

Letting the (ξ, η) components of velocity be (u^ξ, u^η), then

$$u^\xi = \xi_x u + \xi_y v \tag{22}$$

$$u^\eta = \eta_x u + \eta_y v \tag{23}$$

or,

$$\begin{pmatrix} u^\xi \\ u^\eta \end{pmatrix} = \begin{pmatrix} \xi_x & \xi_y \\ \eta_x & \eta_y \end{pmatrix} \begin{pmatrix} u \\ v \end{pmatrix} \tag{24}$$

$$\begin{pmatrix} u \\ v \end{pmatrix} = \frac{1}{J} \begin{pmatrix} \eta_y & -\xi_y \\ -\eta_x & \xi_x \end{pmatrix} \begin{pmatrix} u^\xi \\ u^\eta \end{pmatrix} \tag{25}$$

II.1.3 Basic equations in a general curvilinear coordinate system

Basic equations as obtained by transforming basic equations in an orthogonal coordinates system (x, y) into a general curvilinear coordinate system are shown below:

[Continuity equation]

$$\frac{\partial}{\partial t}\left(\frac{h}{J}\right)+\frac{\partial}{\partial \xi}\left(\frac{hu^{\xi}}{J}\right)+\frac{\partial}{\partial \eta}\left(\frac{hu^{\eta}}{J}\right)=0 \tag{26}$$

[Momentum equations]

$$\begin{aligned}\frac{\partial u^{\xi}}{\partial t}+u^{\xi}\frac{\partial u^{\xi}}{\partial \xi}+u^{\eta}\frac{\partial u^{\xi}}{\partial \eta}+\alpha_1 u^{\xi}u^{\xi}+\alpha_2 u^{\xi}u^{\eta}+\alpha_3 u^{\eta}u^{\eta}= \\ -g\left[\left(\xi_x^2+\xi_y^2\right)\frac{\partial H}{\partial \xi}+\left(\xi_x\eta_x+\xi_y\eta_y\right)\frac{\partial H}{\partial \eta}\right] \\ -\left(C_f+\frac{1}{2}C_D a_s h_v\right)\frac{u^{\xi}}{hJ}\sqrt{\left(\eta_y u^{\xi}-\xi_y u^{\eta}\right)^2+\left(-\eta_x u^{\xi}+\xi_x u^{\eta}\right)^2}+D^{\xi}\end{aligned} \tag{27}$$

$$\begin{aligned}\frac{\partial u^{\eta}}{\partial t}+u^{\xi}\frac{\partial u^{\eta}}{\partial \xi}+u^{\eta}\frac{\partial u^{\eta}}{\partial \eta}+\alpha_4 u^{\xi}u^{\xi}+\alpha_5 u^{\xi}u^{\eta}+\alpha_6 u^{\eta}u^{\eta}= \\ -g\left[\left(\eta_x\xi_x+\eta_y\xi_y\right)\frac{\partial H}{\partial \xi}+\left(\eta_x^2+\eta_y^2\right)\frac{\partial H}{\partial \eta}\right] \\ -\left(C_f+\frac{1}{2}C_D a_s h_v\right)\frac{u^{\eta}}{hJ}\sqrt{\left(\eta_y u^{\xi}-\xi_y u^{\eta}\right)^2+\left(-\eta_x u^{\xi}+\xi_x u^{\eta}\right)^2}+D^{\eta}\end{aligned} \tag{28}$$

where,

$$\alpha_1=\xi_x\frac{\partial^2 x}{\partial \xi^2}+\xi_y\frac{\partial^2 y}{\partial \xi^2},\quad \alpha_2=2\left(\xi_x\frac{\partial^2 x}{\partial \xi\partial \eta}+\xi_y\frac{\partial^2 y}{\partial \xi\partial \eta}\right),\quad \alpha_3=\xi_x\frac{\partial^2 x}{\partial \eta^2}+\xi_y\frac{\partial^2 y}{\partial \eta^2} \tag{29}$$

$$\alpha_4=\eta_x\frac{\partial^2 x}{\partial \xi^2}+\eta_y\frac{\partial^2 y}{\partial \xi^2},\quad \alpha_5=2\left(\eta_x\frac{\partial^2 x}{\partial \xi\partial \eta}+\eta_y\frac{\partial^2 y}{\partial \xi\partial \eta}\right),\quad \alpha_6=\eta_x\frac{\partial^2 x}{\partial \eta^2}+\eta_y\frac{\partial^2 y}{\partial \eta^2} \tag{30}$$

$$D^{\xi}=\left(\xi_x\frac{\partial}{\partial \xi}+\eta_x\frac{\partial}{\partial \eta}\right)\left[\nu_t\left(\xi_x\frac{\partial u^{\xi}}{\partial \xi}+\eta_x\frac{\partial u^{\xi}}{\partial \eta}\right)\right]+\left(\xi_y\frac{\partial}{\partial \xi}+\eta_y\frac{\partial}{\partial \eta}\right)\left[\nu_t\left(\xi_y\frac{\partial u^{\xi}}{\partial \xi}+\eta_y\frac{\partial u^{\xi}}{\partial \eta}\right)\right] \tag{31}$$

$$D^{\eta} = \left(\xi_x \frac{\partial}{\partial \xi} + \eta_x \frac{\partial}{\partial \eta} \right) \left[\nu_t \left(\xi_x \frac{\partial u^{\eta}}{\partial \xi} + \eta_x \frac{\partial u^{\eta}}{\partial \eta} \right) \right] + \left(\xi_y \frac{\partial}{\partial \xi} + \eta_y \frac{\partial}{\partial \eta} \right) \left[\nu_t \left(\xi_y \frac{\partial u^{\eta}}{\partial \xi} + \eta_y \frac{\partial u^{\eta}}{\partial \eta} \right) \right] \tag{32}$$

$$\xi_x = \frac{\partial \xi}{\partial x}, \quad \xi_y = \frac{\partial \xi}{\partial y}, \quad \eta_x = \frac{\partial \eta}{\partial x}, \quad \eta_y = \frac{\partial \eta}{\partial y} \tag{33}$$

$$u^{\xi} = \xi_x u + \xi_y v \, , \; u^{\eta} = \eta_x u + \eta_y v \tag{34}$$

$$J = \frac{1}{x_{\xi} y_{\eta} - x_{\eta} y_{\xi}} \tag{35}$$

As for diffusion terms D^{ξ} and D^{η} in the momentum equations in general coordinates, since developing those terms will make the number of terms huge, they are simplified by assuming the following conditions:

1. The second-order derivative for the metric coefficient is assumed to be locally zero.
2. Those terms are locally treated as pseudo-orthogonal coordinates.

As a result, the diffusion terms can be approximated as follows:

$$D^{\xi} \simeq \frac{\partial}{\partial \xi} \left(\nu_t \xi_r^2 \frac{\partial u^{\xi}}{\partial \xi} \right) + \frac{\partial}{\partial \eta} \left(\nu_t \eta_r^2 \frac{\partial u^{\xi}}{\partial \eta} \right) \tag{36}$$

$$D^{\eta} \simeq \frac{\partial}{\partial \xi} \left(\nu_t \xi_r^2 \frac{\partial u^{\eta}}{\partial \xi} \right) + \frac{\partial}{\partial \eta} \left(\nu_t \eta_r^2 \frac{\partial u^{\eta}}{\partial \eta} \right) \tag{37}$$

where ξ_r and η_r are parameters each representing the ratio of the local grid size in general coordinates to the full-scale length of the grid. They are defined as follows:

$$\frac{\Delta \xi}{\Delta \tilde{\xi}} = \xi_r , \quad \frac{\Delta \eta}{\Delta \tilde{\eta}} = \eta_r \tag{38}$$

Note that to derive the approximate equations of D^{ξ} and D^{η} above, the following relations are used, based on the assumption of a relationship of local orthogonality.

$$\xi_x^2 + \xi_y^2 = \xi_r^2 \left(\tilde{\xi}_x^{\,2} + \tilde{\xi}_y^{\,2} \right) = \xi_r^2 \left(\sin^2 \theta + \cos^2 \theta \right) = \xi_r^2 \tag{39}$$

$$\xi_x \eta_x + \xi_y \eta_y = \xi_r \eta_r \left(\tilde{\xi}_x \tilde{\eta}_x + \tilde{\xi}_y \tilde{\eta}_y \right) = \xi_r \eta_r \left(-\cos\theta \sin\theta + \cos\theta \sin\theta \right) = 0 \tag{40}$$

$$\eta_x^2 + \eta_y^2 = \eta_r^2 \left(\tilde{\eta}_x^{\,2} + \tilde{\eta}_y^{\,2} \right) = \eta_r^2 \left(\sin^2 \theta + \cos^2 \theta \right) = \eta_r^2 \tag{41}$$

$$J = \xi_x \eta_y - \xi_y \eta_x = \xi_r \eta_r \left(\tilde{\xi}_x \tilde{\eta}_y - \tilde{\xi}_y \tilde{\eta}_x \right) = \xi_r \eta_r \left(\sin^2 \theta + \cos^2 \sin \theta \right) = \xi_r \eta_r \tag{42}$$

where θ represents the angle formed by x-axis and ξ-axis (or y-axis and η-axis).

II.2 Turbulence model

Turbulence means disordered flow that contains eddies of various sizes and structures. Nays2DH allows the user to select the turbulence closure from among [Constant eddy viscosity], [Zero-equation model] and [k-ε model].

II.2.1 Constant eddy viscosity

Eddy viscosity coefficient ν_t is the apparent kinetic viscosity coefficient of a flow in turbulent state. When [Constant eddy viscosity] is selected, calculation is performed with ν_t in Eqs. (5) and (6) being taken as 10^{-6} (m^2/s).

II.2.2 Zero-equation model

Eddy viscosity coefficient ν_t is generally represented as the product of the turbulence representative velocity ν_t and the representative length l.

$$\nu_t = v_t l \tag{43}$$

For a flow field where depth and roughness change only slowly in the transverse direction, assuming that the eddy viscosity coefficient in the horizontal direction and the eddy viscosity coefficient in the vertical direction are on the same order, and considering that the bottom friction velocity and the depth dominantly dictate momentum transport, the eddy viscosity coefficient ν_t is expressed by the following equation.

$$\nu_t = a u_* h \tag{44}$$

where a is a proportional constant.

Since the value of a related to the momentum transport in the vertical direction, according to experiments by Fisher[1) and Webel and Schatzmann[2), is around 0.07. The eddy viscosity coefficient ν_t is expressed using the von Kàrmàn coefficient κ (0.4) as per the following formula:

$$\nu_t = \frac{\kappa}{6} A u_* h + B \tag{45}$$

Since this modeling does not require any transport equation for turbulence statistics, it is called "zero-equation model." A and B are user defined parameters for the eddy viscosity coefficient. Default values of A and B are 1 and 0, respectively. In horizontal two dimensional flow calculations, since three dimensional flow structures are completely neglected, unrealistic horizontal large vortices tend to be generated. The user can adjust A and B for controlling such vortices.

II.2.3 k-ε model

The eddy viscosity coefficient ν_t in the standard k-ε model is expressed by the following equation:

$$\nu_t = C_\mu \frac{k^2}{\varepsilon} \tag{46}$$

where C_μ is a model constant. k and ε are obtained by the following equations:

$$\frac{\partial k}{\partial t}+u\frac{\partial k}{\partial x}+v\frac{\partial k}{\partial y}=\frac{\partial}{\partial x}\left(\frac{\nu_t}{\sigma_k}\frac{\partial k}{\partial x}\right)+\frac{\partial}{\partial y}\left(\frac{\nu_t}{\sigma_k}\frac{\partial k}{\partial y}\right)+P_h+P_{kv}-\varepsilon \tag{47}$$

$$\frac{\partial \varepsilon}{\partial t}+u\frac{\partial \varepsilon}{\partial x}+v\frac{\partial \varepsilon}{\partial y}=\frac{\partial}{\partial x}\left(\frac{\nu_t}{\sigma_\varepsilon}\frac{\partial \varepsilon}{\partial x}\right)+\frac{\partial}{\partial y}\left(\frac{\nu_t}{\sigma_\varepsilon}\frac{\partial \varepsilon}{\partial y}\right)+C_{1\varepsilon}\frac{\varepsilon}{k}P_h+P_{\varepsilon v}-C_{2\varepsilon}\frac{\varepsilon^2}{k} \tag{48}$$

Table II-1 Model constants

C_μ	$C_{1\varepsilon}$	$C_{2\varepsilon}$	σ_k	σ_ε
0.09	1.44	1.92	1.0	1.3

where $C_{1\varepsilon}$, $C_{2\varepsilon}$, σ_k and σ_ε are model constants. Their respective values are shown in Table II-I.

Note that P_h , P_{kv} and $P_{\varepsilon v}$ are calculated by the following equations:

$$P_h=\nu_t\left[2\left(\frac{\partial u}{\partial x}\right)^2+2\left(\frac{\partial v}{\partial y}\right)^2+\left(\frac{\partial u}{\partial y}\right)^2+\left(\frac{\partial v}{\partial x}\right)^2+2\frac{\partial u}{\partial y}\frac{\partial v}{\partial x}\right] \tag{49}$$

$$P_{kv}=C_k\frac{u_*^3}{h} \tag{50}$$

$$P_{\varepsilon v}=C_\varepsilon\frac{u_*^4}{h^2} \tag{51}$$

II.3 Model of river bed friction

In Nays2DH, the friction of river bed is set using Manning's roughness parameter. For Manning's roughness parameter, the user can define this parameter locally in each computational cell.

In Eq. (4), bed shear forces τ_x and τ_y are expressed by using coefficient of bed shear force C_f. The coefficient of bed shear force C_f is estimated by Manning's roughness parameter n_m as follows:

$$C_f=\frac{g n_m^2}{h^{1/3}} \tag{52}$$

This Manning’s roughness parameter can be estimated from the relative roughness height, k_s, by using the Manning-Strickler equation[3] as follows.

$$n_m = \frac{k_s^{1/6}}{7.66\sqrt{g}} \tag{53}$$

where k_s is the relative roughness height which is defines as αd, d is sediment diameter, α is an empirical constant in the range of 1 to 3, and g is the gravitational acceleration.

II.4 Method for calculating resistance by vegetation

In Nays2DH, resistance exerted by vegetation is set with the drag coefficient of vegetation C_D, the area of interception by vegetation per unit volume a_s and the height of vegetation. The area of interception by vegetation per unit volume a_s can be set at each computational cell.

The area of interception by vegetation per unit volume a_s is calculated by the following equation proposed by Shimizu et al.[4]:

$$a_s = \frac{n_s D_s}{S_s^{\,2}} \tag{54}$$

where n_s is the number of vegetation, D_s is the average diameter of trunks and S_s is the sampling grid width.

II.5 Basic equations of sediment transport

When performing riverbed deformation calculations with Nays2DH, the user can select among two sediment types: [Bed load only] and [Bed load and suspended load]. The user can also select between two types of bed material: [Uniform grain size] and [Non-uniform grain size]. The followings are the description of morphodynamic model with uniform grain size. The description of morphodynamic model with non-uniform size is given in Section II.6.

II.5.1 Dimensionless river bed shearing force

Composite velocity V is defined by the following equation:

$$V = \sqrt{u^2 + v^2} \tag{55}$$

The Shields number τ_* exerted on the riverbed is as follows:

$$\tau_* = \frac{hI_e}{s_g d} \tag{56}$$

where h is depth, I_e is energy slope, s_g is specific weight of bed material in fluid and d is grain size of bed material. By applying Manning's formula to I_e, τ_* is expressed as follows:

$$\tau_* = \frac{C_f V^2}{s_g g d} = \frac{n_m^2 V^2}{s_g d h^{1/3}} \tag{57}$$

II.5.2 Bed load transport

The user can estimate the total bedload transport q_b in the depth-averaged velocity direction (in the direction of V) by using Meyer-Peter and Müller formula[5)] or Ashida and Michiue formula[6-8)] as follows.

- Meyer-Peter and Müller formula

$$q_b = 8(\tau_* - \tau_{*c})^{1.5} \sqrt{s_g g d^3}\, r_b \tag{58}$$

- Ashida and Michiue formula

$$q_b = 17 \tau_{*e}^{1.5} \left(1 - K_c \frac{\tau_{*c}}{\tau_*}\right)\left(1 - \sqrt{K_c \frac{\tau_{*c}}{\tau_*}}\right)\sqrt{s_g g d^3}\, r_b \tag{59}$$

where, effective Shields number is calculated as follows:

$$u_{*e}^2 = \frac{V^2}{\left(6 + 2.5\ln\frac{h}{d(1+2\tau_*)}\right)^2}, \quad \tau_{*e} = \frac{u_{*e}^2}{sgd} \tag{60}$$

where τ_{*c} is critical Shields number which is calculated by Iwagaki formula[9].

K_c is the modification function of the effect of the local bed slope on the sediment transport as follows:

$$K_c = 1 + \frac{1}{\mu_s}\left[\left(\frac{\rho}{\rho_s - \rho} + 1\right)\cos\alpha\tan\theta_x + \sin\alpha\tan\theta_y\right] \tag{61}$$

where α is the angle of deviation of near-bed flow from x-direction and defined as follows:

$$\alpha = \arctan\left(\frac{v_b}{u_b}\right) \tag{62}$$

μ_s is the static friction coefficient of bed material. θ_x and θ_y are bed inclinations in the x and y-directions, respectively. These inclinations are evaluated as follows,

$$\theta_x = \arctan\left(\frac{\partial\xi}{\partial x}\frac{\partial z_b}{\partial\xi} + \frac{\partial\eta}{\partial x}\frac{\partial z_b}{\partial\eta}\right), \quad \theta_y = \arctan\left(\frac{\partial\xi}{\partial y}\frac{\partial z_b}{\partial\xi} + \frac{\partial\eta}{\partial y}\frac{\partial z_b}{\partial\eta}\right) \tag{63}$$

r_b is the function of the exchange layer thickness as follows:

$$r_b = 1 \qquad E_{sd} \geq E_{be} \tag{64}$$

$$r_b = \frac{E_b}{E_{be}} \qquad E_{sd} \leq E_{be} \tag{65}$$

II.5.3 Calculation of bedload transport vector

The total bedload transport is divided to the contravariant form of bedload fluxes in ξ and η directions by considering the effect of secondary flows to the velocity field near river bed and the local bed slope effect to the sediment transport direction. For the calculation of bedload flux in ξ and η directions, the user can select among [Watanabe formula[10]] and [Ashida, Egashira and Liu formula[6-8]].

-<u>Watanabe formula</u>

Bed load transport in $\xi-$ and η-directions is given by the following equations:

$$\tilde{q}_b^{\ \xi} = q_b\left[\frac{\tilde{u}_b^{\xi}}{V_b} - \gamma\left(\frac{\partial z_b}{\partial\tilde{\xi}} + \cos\theta\frac{\partial z_b}{\partial\tilde{\eta}}\right)\right] \tag{66}$$

$$\tilde{q}_b^{\eta} = q_b\left[\frac{\tilde{u}_b^{\eta}}{V_b} - \gamma\left(\frac{\partial z_b}{\partial \tilde{\eta}} + \cos\theta\frac{\partial z_b}{\partial \tilde{\xi}}\right)\right] \tag{67}$$

where $\tilde{u}_b^{\xi}$ and $\tilde{u}_b^{\eta}$ are flow velocities near river bed in ξ- and η-directions, respectively, V_b is composite velocity near the riverbed, and θ is the angle formed the ξ and η axes.

γ is a correction coefficient for local bed slope effect, which is given by the following equation according to Hasegawa[11]:

$$\gamma = \sqrt{\frac{\tau_{*c}}{\mu_s \mu_k \tau_*}} \tag{68}$$

where μ_s and μ_k are static friction coefficient and dynamic friction coefficient of bed material, respectively.

-Ashida, Egashira and Liu formula

The contravariant form of bedload transport flux in ξ and η direction is calculated as follows:

$$\tilde{q}_b^{\xi} = \frac{\partial \xi}{\partial x} q_{bx} + \frac{\partial \xi}{\partial y} q_{by}, \quad \tilde{q}_b^{\eta} = \frac{\partial \eta}{\partial x} q_{bx} + \frac{\partial \eta}{\partial y} q_{by} \tag{69}$$

q_{bx} and q_{by} are the bed load in x- and y-directions, respectively as follows[6-8].

$$q_{bx} = q_b \cos\beta, \quad q_{by} = q_b \sin\beta \tag{70}$$

The local bed slope along direction of bed load of sediment mean diameter (θ) is obtained as follows.

$$\sin\theta = \cos\beta \sin\theta_x + \sin\beta \sin\theta_y \tag{71}$$

The angle of deviation of vector of bedload from x-direction is given as follows:

$$\tan\beta = \frac{\sin\alpha - \Pi\Theta_y\left(\frac{u_{*c}^2}{u_*^2}\right)\tan\theta_y}{\cos\alpha - \Pi\Theta_x\left(\frac{u_{*c}^2}{u_*^2}\right)\tan\theta_x} \tag{72}$$

$$\Pi = K_{ld} + 1/\mu_s \tag{73}$$

$$\Theta_y = \frac{1}{1+\tan^2\theta_x + \tan^2\theta_y}, \quad \Theta_x = \Theta_y + \frac{\rho}{\rho_s - \rho}\cos^2\theta_x \tag{74}$$

where, K_{ld} (=0.85) is the ratio of lift force to drag force.

II.5.4 Calculation of velocity near the riverbed

The following simple relational equation is given for the relationship between depth-averaged velocity and near-bed velocity along the depth-averaged flow:

$$\tilde{u}_b^s = \beta V \tag{75}$$

where $\tilde{u}_b^s$ is near-bed velocity along the streamline of the depth-averaged flow (hereafter simply called streamline). According to Engelund[12], if we apply a parabolic distribution to the velocity distribution in the depth direction, then β is given by the following equation:

$$\beta = 3\frac{1-\sigma}{3-\sigma}, \quad \sigma = \frac{3}{\phi_0 \kappa + 1} \tag{76}$$

where ϕ_0 is velocity coefficient (=V/u_*).

Generally, in cases where the streamline curves, secondary flow (helical flow) occurs. In Nays2DH, the user can select how to evaluate secondary flows from by (1) setting the strength of secondary flow (e.g., Engelund[12]) or by (2) solving an equation of depth-averaged vorticity in streamwise direction[13, 14].

-<u>setting the strength of secondary flow</u>

We use the following equation for calculating the near-bed velocity under the effect of such secondary flow:

$$\tilde{u}_b^n = \tilde{u}_b^s N_* \frac{h}{r_s} \tag{77}$$

where $\tilde{u}_b^n$ is near-bed velocity in transverse direction, r_s is curvature radius of the streamline and N_* is a constant (=7; Engelund[12]).

-Solve an equation of depth-averaged vorticity in streamwise direction

The transport of depth-averaged vorticity in streamwrise direction is calculated by solving the equation of depth-averaged vorticity, and the streamwise vorticity is related to secondary flows. By using this model, the effect of development of secondary flows in time and space can be included into the bedload transport direction. The equation of depth-averaged vorticity in streamwise direction is given as follows:

$$\frac{\partial}{\partial t}\left(\lambda_v \frac{A_n}{J}\right)+\frac{\partial}{\partial \xi}\left(u^{\xi}\,|_s\,\frac{\tilde{u}_s^n}{J}-u^{\xi}\,|_b\,\frac{\tilde{u}_b^n}{J}\right)+\frac{\partial}{\partial \eta}\left(u^{\eta}\,|_s\,\frac{\tilde{u}_s^n}{J}-u^{\eta}\,|_b\,\frac{\tilde{u}_b^n}{J}\right)$$
$$=\frac{1}{r_s}\left(\tilde{u}^{s2}\,|_s-\tilde{u}^{s2}\,|_b\right)=\frac{A_n}{J}\frac{V}{\chi_1^3 h}\left(\chi^2+\frac{7}{12}\chi+\frac{1}{12}\right) \tag{78}$$

where, A_n is intensity of secondary flows, the lower suffixes *s* and *b* represent the value in water surface and river bed, respectively, and the upper suffixes *s* and *n* represent the streamwise direction and the orthogonal direction in the streamwise direction, respectively. The other coefficients and velocity in Eq. (78) can be defined as follows:

$$\lambda_v=-\frac{1}{\alpha\chi_1^3}\frac{V}{u_*}\left(\frac{1}{12}\chi^2+\frac{11}{360}\chi+\frac{1}{504}\right) \tag{79}$$

$$\tilde{u}_b^s=V\frac{\chi}{\chi_1},\quad \tilde{u}_s^s=V\frac{\chi+1/2}{\chi_1} \tag{80}$$

$$\tilde{u}_b^n=A_n\frac{\chi}{\alpha^2\chi_1^2}\left(\frac{2}{45}\chi+\frac{4}{315}\right) \tag{81}$$

$$\tilde{u}_s^n=-A_n\frac{1}{\alpha^2\chi_1^2}\left(\frac{7}{180}\chi^2+\frac{1}{56}\chi+\frac{1}{504}\right) \tag{82}$$

$$\chi_1=\alpha\frac{V}{u_*},\quad \chi=\chi_1-\frac{1}{3},\quad \alpha=\frac{\kappa}{6} \tag{83}$$

The flow velocity in *n* direction near river bed which includes the development of secondary flow intensity is given by Eq. (81). Eq. (81) corresponds to Eq. (77) which is flow velocity in *n* direction by Engelund model.

By using the velocity of secondary flows near river bed, the direction of bedload transport is correlated. From Eqs. (75) and (77), V_b in Eqs. (66) and (67) are given as follows:

$$V_b=\sqrt{\tilde{u}_b^{s2}+\tilde{u}_b^{n2}}\approx\tilde{u}_b^s \tag{84}$$

Note that the approximation in Eq. (84) is due to the fact that $\tilde{u}_b^n$ is generally one order of magnitude smaller than $\tilde{u}_b^s$.

$\tilde{u}_b^{\xi}$ and $\tilde{u}_b^{\eta}$ are obtained with the following transformation:

$$\tilde{u}_b^{\xi}=\frac{\partial\tilde{\xi}}{\partial s}\tilde{u}_b^s+\frac{\partial\tilde{\xi}}{\partial n}\tilde{u}_b^n=\left(\frac{\partial x}{\partial s}\frac{\partial\tilde{\xi}}{\partial x}+\frac{\partial y}{\partial s}\frac{\partial\tilde{\xi}}{\partial y}\right)\tilde{u}_b^s+\left(\frac{\partial x}{\partial n}\frac{\partial\tilde{\xi}}{\partial x}+\frac{\partial y}{\partial n}\frac{\partial\tilde{\xi}}{\partial y}\right)\tilde{u}_b^n$$
$$=\left(\cos\theta_s\tilde{\xi}_x+\sin\theta_s\tilde{\xi}_y\right)\tilde{u}_b^s+\left(-\sin\theta_s\tilde{\xi}_x+\cos\theta_s\tilde{\xi}_y\right)\tilde{u}_b^n$$

$$=\frac{1}{\xi_r}\left\{\left(\cos\theta_s\xi_x+\sin\theta_s\xi_y\right)\tilde{u}_b^s+\left(-\sin\theta_s\xi_x+\cos\theta_s\xi_y\right)\tilde{u}_b^n\right\} \tag{85}$$

$$\tilde{u}_b^\eta=\frac{\partial\tilde{\eta}}{\partial s}\tilde{u}_b^s+\frac{\partial\tilde{\eta}}{\partial n}\tilde{u}_b^n=\left(\frac{\partial x}{\partial s}\frac{\partial\tilde{\eta}}{\partial x}+\frac{\partial y}{\partial s}\frac{\partial\tilde{\eta}}{\partial y}\right)\tilde{u}_b^s+\left(\frac{\partial x}{\partial n}\frac{\partial\tilde{\eta}}{\partial x}+\frac{\partial y}{\partial n}\frac{\partial\tilde{\eta}}{\partial y}\right)\tilde{u}_b^n$$

$$=\left(\cos\theta_s\tilde{\eta}_x+\sin\theta_s\tilde{\eta}_y\right)\tilde{u}_b^s+\left(-\sin\theta_s\tilde{\eta}_x+\cos\theta_s\tilde{\eta}_y\right)\tilde{u}_b^n$$

$$=\frac{1}{\eta_r}\left\{\left(\cos\theta_s\eta_x+\sin\theta_s\eta_y\right)\tilde{u}_b^s+\left(-\sin\theta_s\eta_x+\cos\theta_s\eta_y\right)\tilde{u}_b^n\right\} \tag{86}$$

where θ_s represent the angle of the streamline with x-axis, relational equations including the following are used:

$$\frac{\partial x}{\partial n}=-\frac{v}{V}=-\sin\theta_s,\quad \frac{\partial y}{\partial n}=\frac{u}{V}=\cos\theta_s \tag{87}$$

$$\frac{\partial x}{\partial s}=-\frac{u}{V}=\cos\theta_s,\quad \frac{\partial y}{\partial s}=\frac{v}{V}=\sin\theta_s \tag{88}$$

In addition, although β is expressed as in Eq. (76), β turns out to be included both in the numerator and denominator of the first terms of the right side of Eqs. (66) and (67), so β can be an arbitrary value.

II.5.5 Streamline curvature

The streamline curvature applied in Eqs. (77) and (78) is obtained by the following equation:

$$\frac{1}{r_s}=\frac{\partial\theta_s}{\partial s} \tag{89}$$

where, θ_s is the angle between the x-axis and the s-direction defined as follows;

$$\theta_s=\tan^{-1}\left(\frac{v}{u}\right) \tag{90}$$

hence

$$\frac{1}{r_s}=\frac{\partial}{\partial s}\left[\tan^{-1}(T)\right]=\frac{\partial}{\partial T}\left[\tan^{-1}(T)\right]\frac{\partial T}{\partial s}=\frac{1}{1+T^2}\frac{\partial T}{\partial s} \tag{91}$$

where $T=v/u$. Here,

$$\frac{1}{1+T^2}=\frac{1}{1+\left(\frac{v}{u}\right)^2}=\frac{u^2}{u^2+v^2}=\frac{u^2}{V^2} \tag{92}$$

$$\frac{\partial T}{\partial s}=\frac{\partial}{\partial s}\left(\frac{v}{u}\right)=\frac{u\frac{\partial v}{\partial s}-v\frac{\partial u}{\partial s}}{u^2} \tag{93}$$

$$\frac{\partial}{\partial s}=\frac{\partial x}{\partial s}\frac{\partial}{\partial x}+\frac{\partial y}{\partial s}\frac{\partial}{\partial y}=\frac{u}{V}\frac{\partial}{\partial x}+\frac{\upsilon}{V}\frac{\partial}{\partial y}=\frac{u}{V}\left(\xi_x\frac{\partial}{\partial \xi}+\eta_x\frac{\partial}{\partial \eta}\right)+\frac{v}{V}\left(\xi_y\frac{\partial}{\partial \xi}+\eta_y\frac{\partial}{\partial \eta}\right) \tag{94}$$

Therefore, curvature $1/r_s$ is expressed by the following equation:

$$\frac{1}{r_s}=\frac{1}{V^3}\left[u^2\left(\xi_x\frac{\partial v}{\partial \xi}+\eta_x\frac{\partial v}{\partial \eta}\right)+uv\left(\xi_y\frac{\partial v}{\partial \xi}+\eta_y\frac{\partial v}{\partial \eta}\right)-uv\left(\xi_x\frac{\partial u}{\partial \xi}+\eta_x\frac{\partial u}{\partial \eta}\right)-v^2\left(\xi_y\frac{\partial u}{\partial \xi}+\eta_y\frac{\partial u}{\partial \eta}\right)\right] \tag{95}$$

II.5.6 Upward flux of suspended load from river bed

For upward flux of suspended load from river bed, the user can select from among [Itakura and Kishi formula[15)]] and [Lane-Kalinske formula[16)]].

-<u>Itakura and Kishi formula</u>

$$q_{su}=K\left[a_*\frac{\rho_s-\rho}{\rho_s}\cdot\frac{gd}{u_*}\Omega-w_f\right]r_b \tag{96}$$

$$\Omega=\frac{\tau_*}{B_*}\cdot\frac{\int_{a'}^{\infty}\xi\frac{1}{\sqrt{\pi}}\exp\left[-\xi^2\right]d\xi}{\int_{a'}^{\infty}\frac{1}{\sqrt{\pi}}\exp\left[-\xi^2\right]d\xi}+\frac{\tau_*}{B_*\eta_0}-1 \tag{97}$$

$$a'=\frac{B_*}{\tau_*}-\frac{1}{\eta_0},\quad \eta_0=0.5,\quad a_*=0.14,\quad K=0.008 \tag{98}$$

where q_{su} is the upward flux of suspended load from river bed and w_f is the settling velocity of suspended sediment, which is obtained with Rubey's equation[17)]. B_* is a conversion factor for applying friction velocity to the velocity in lift force calculation. The constant value of $B_*=0.143$ is used in the case of uniform grain size.

-<u>Lane and Kalinske formula (unit: ppm)</u>

$$q_{su}=5.55\left[\frac{1}{2}\frac{u_*}{w_f}\exp\left(-\frac{w_f}{u_*}\right)\right]^{1.61}w_f r_b \tag{99}$$

II.5.7 Continuity equation of suspended load concentration

The continuity equation of suspended load concentration in the general curvilinear coordinate system is represented as follows:

$$\frac{\partial}{\partial t}\left(\frac{ch}{J}\right)+\frac{\partial}{\partial \xi}\left(\frac{u^{\xi}ch}{J}\right)+\frac{\partial}{\partial \eta}\left(\frac{u^{\eta}ch}{J}\right)=\frac{q_{su}-w_f c_b}{J} \tag{100}$$

where c is depth-averaged suspended load concentration and c_b is reference concentration of suspended load. Diffusion terms have been omitted from the equation, for simplicity. The reference concentration of suspended load is calculated by assuming an exponential distribution of suspended sediment in vertical direction as follows[18]:

$$c_b=\frac{\beta_c c}{1-\exp(-\beta_c)} \tag{101}$$

with,

$$\beta_c=\frac{6w_f}{\kappa u_*} \tag{102}$$

II.5.8 Continuity equation of sediment transport

First, a continuity equation of sediment transport in two-dimensional orthogonal coordinates is given by:

$$\frac{\partial}{\partial t}(c_b E_b)+(1-\lambda)\frac{\partial z}{\partial t}+\left[\frac{\partial q_{bx}}{\partial x}+\frac{\partial q_{by}}{\partial y}+q_{su}-w_f c_b\right]=0 \tag{103}$$

where t is time, z is riverbed elevation, and λ is a void ratio of bed material.

When [Bed load only] is selected, suspended load supplied per unit area, settling velocity of suspended sediment and concentration at the control point area assumed to be zero.

Next, just as for the continuity equation of flow, we transform the above equation into general coordinates.

$$\frac{\partial}{\partial t}\left(\frac{c_b E_b}{J}\right)+(1-\lambda)\frac{\partial}{\partial t}\left(\frac{z}{J}\right)+\left[\frac{\partial}{\partial \xi}\left(\frac{q_b^{\xi}}{J}\right)+\frac{\partial}{\partial \eta}\left(\frac{q_b^{\eta}}{J}\right)+\frac{q_{su}-w_f c_b}{J}\right]=0 \quad \cdots \quad E_{sd}\geq E_{be}\frac{c_b}{1-\lambda} \tag{104}$$

$$\frac{\partial}{\partial t}\left(\frac{z}{J}\right)=0 \quad \cdots \quad E_{sd}\leq E_{be}\frac{c_b}{1-\lambda} \tag{105}$$

II.6 Non-uniform grain size model

When [Non-uniform grain size] is selected for the riverbed material, the basic equations to be used for riverbed deformation are described below.

A non-uniform grain size riverbed is made up of a certain grading of riverbed material. To handle this grading mathematically, we divide the cumulative grading curve into n layers as shown in Figure II-1, and each layer is expressed by its representative grain size d_k and the probability of that grain size existing p_k. Median diameter d_m is defined by the following equation:

$$d_m = \sum_{k=1}^{n} p_k d_k \tag{106}$$

where d_k is representative grain size of layer k, and p_k is layer k's grain size as a proportion of the entire riverbed.

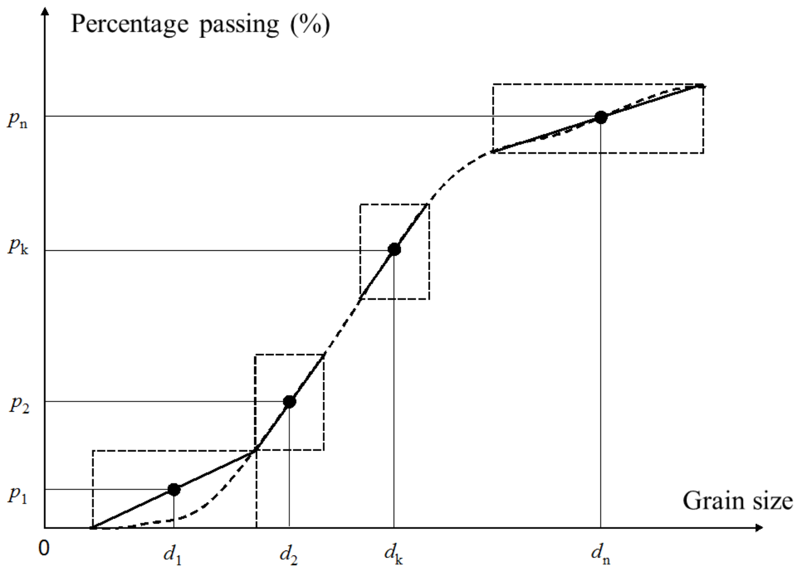

Figure II-1. Handling of grain size distribution

II.6.1 Continuity equation of sediment transport

The riverbed continuity equation in generalized curvelinear coordinate system is given as follows:

$$\frac{\partial}{\partial t}\left(\frac{z}{J}\right)+\frac{1}{1-\lambda}\left[\frac{\partial}{\partial \xi}\left(\frac{\sum q_{bk}{}^{\xi}}{J}\right)+\frac{\partial}{\partial \eta}\left(\frac{\sum q_{bk}{}^{\eta}}{J}\right)+\frac{\sum\left(q_{suk}-c_{bk}w_{fk}\right)}{J}\right]=0 \tag{107}$$

where, suffix k represents the value of size class k, $q_{bk}{}^{\xi}$ and $q_{bk}{}^{\eta}$ are bed load transport in the ξ and η directions, q_{suk} is upward flux of suspended load from river bed, c_{bk} is the reference concentration of suspended load and w_{fk} is the settling velocity of bed material in fluid.

II.6.2 Bedload transport

Total bed load transport in each sediment size class is calculated with the Ashida-Michiue formula that has been expanded for each grain size.

$$q_b = 17 p_{mk} \tau_{*ek}^{1.5} \left(1 - K_c \frac{\tau_{*ck}}{\tau_{*k}}\right)\left(1 - \sqrt{K_c \frac{\tau_{*ck}}{\tau_{*k}}}\right)\sqrt{s_g g d_k^3}\, r_b \tag{108}$$

where q_{bk} is total sediment transport in the streamline direction, s_g is submerged specific gravity of bed material, τ_{*k} is Shields number on grain of the size in layer k, τ_{*ck} is critical Shields number for grain of the size in layer k, p_{mk} is the fraction of sediment size class in bedload layer, τ_{*ek} is the effective Shields number which obtained by following equation:

$$u_{*em}^2 = \frac{V^2}{\left(6 + 2.5 \ln \dfrac{h}{d_m (1 + 2\tau_{*m})}\right)^2}, \quad \tau_{*ek} = \frac{u_{*em}^2}{s g d_k} \tag{109}$$

where, suffix m represents the physical value related to the mean diameter.

Furthermore, in calculating the critical Shields number, the hiding and exposure effect must be taken into consideration. Here, we adopt Asada's formula[19].

$$\frac{\tau_{*ck}}{\tau_{*cm}} = \left[\frac{\log_{10} 23}{\log_{10}\left(21 \dfrac{d_k}{d_m} + 2\right)}\right]^2 \tag{110}$$

where τ_{*cm} is critical Shields number for grains of median diameter, which is calculated using Iwagaki's formula[9].

The user can select the calculation method of the contravariant form of bedload flux in ξ and η direction from [Watanabe formula[10]] and [Ashida, Egashira and Liu formula[6-8]].

-<u>Watanabe formula</u>

$$\tilde{q}_{bk}{}^{\xi} = q_{bk}\left[\frac{\tilde{u}_b^{\xi}}{V_b} - \gamma\left(\frac{\partial z}{\partial \tilde{\xi}} + \cos\theta \frac{\partial z}{\partial \tilde{\eta}}\right)\right] \tag{111}$$

$$\tilde{q}_{bk}{}^{\eta} = q_{bk}\left[\frac{\tilde{u}_b^{\eta}}{V_b} - \gamma\left(\frac{\partial z}{\partial \tilde{\eta}} + \cos\theta \frac{\partial z}{\partial \tilde{\xi}}\right)\right] \tag{112}$$

For γ, we expand Hasegawa's formula[11] to each grain size in an abbreviated manner.

$$\gamma = \sqrt{\frac{\tau_{*ck}}{\mu_s \mu_k \tau_{*k}}} \tag{113}$$

-Ashida, Egashira and Liu formula

The contravariant form of bedload transport flux in ξ and η direction is calculated as follows:

$$\tilde{q}_{bk}^{\xi} = \frac{\partial \xi}{\partial x} q_{bxk} + \frac{\partial \xi}{\partial y} q_{byk}, \quad \tilde{q}_{bk}^{\eta} = \frac{\partial \eta}{\partial x} q_{bxk} + \frac{\partial \eta}{\partial y} q_{byk} \tag{114}$$

q_{bxk} and q_{byk} are the bed load transport of size class k in x- and y-directions, respectively as follows[6-8].

$$q_{bxk} = q_{bk} \cos \beta_k, \quad q_{byk} = q_{bk} \sin \beta_k \tag{115}$$

The local bed slope along direction of bed load of sediment mean diameter (θ) is obtained as follows.

$$\sin \theta = \cos \beta_k \sin \theta_x + \sin \beta_k \sin \theta_y \tag{116}$$

where β_m is the deviation angle of bed load of mean diameter to the x-direction. The deviation angle of bed load of size class k to the x-direction (β_k), which depends on the flow near bed and inclination of the bed, is calculated by the following relation.

$$\tan \beta_k = \frac{\sin \alpha - \Pi \Theta_y \left(\frac{u_{*ck}^2}{u_{*k}^2} \right) \tan \theta_y}{\cos \alpha - \Pi \Theta_x \left(\frac{u_{*ck}^2}{u_{*k}^2} \right) \tan \theta_x} \tag{117}$$

The other coefficients in Equations related to Ashida, Egashira and Liu formula are same as the coefficients in case of uniform sediment.

In addition, also in case of non-uniform sediment, the user can use same framework for evaluating the secondary flows to the bedload transport direction that is explained in the uniform sediment case.

II.6.3 Upward flux of suspended load

Similar to uniform sediment case, the user also can select [Itakura and Kishi formula[15]] or [Lane-Kalinske formula[16]] for upward flux of suspended load in non-uniform sediment.

-Itakura and Kishi formula

$$q_{suk} = p_{mk} K \left[a_* \frac{\rho_s - \rho}{\rho_s} \cdot \frac{g d_k}{u_*} \Omega_k - w_{fk} \right] r_b \tag{118}$$

$$\Omega_k = \frac{\tau_{*k}}{B_{*k}} \cdot \frac{\int_{a'}^{\infty} \xi \frac{1}{\sqrt{\pi}} \exp\left[-\xi^2\right] d\xi}{\int_{a'}^{\infty} \frac{1}{\sqrt{\pi}} \exp\left[-\xi^2\right] d\xi} + \frac{\tau_{*k}}{B_{*k}\eta_0} - 1 \tag{119}$$

$$a' = \frac{B_{*k}}{\tau_{*k}} - \frac{1}{\eta_0}, \quad \eta_0 = 0.5, \quad a_* = 0.14, \quad K = 0.008 \tag{120}$$

For evaluating B_{*k} in non-uniform sediment case, the formula which included the hiding and exposure effect is adopted as follows:

$$B_{*k} = \xi_k B_{*0}, \quad \xi_k = \frac{\tau_{*ck}}{\tau_{*ck0}}, \quad B_{*0} = 0.143 \tag{121}$$

where, τ_{*ck0} is the critical Shields number of d_k which is not included the hiding and exposure effect, just like in the uniform sediment case.

・ Lane-Kalinske formula (unit: ppm)

$$q_{suk} = 5.55 p_{mk} \left[\frac{1}{2} \frac{u_*}{w_{fk}} \exp\left(-\frac{w_{fk}}{u_*} \right) \right]^{1.61} w_{fk} r_b \tag{122}$$

II.6.4 Continuity equation of suspended load concentration

The continuity equation of suspended load concentration with non-uniform sediment case is given as follows:

$$\frac{\partial}{\partial t}\left(\frac{c_k h}{J}\right) + \frac{\partial}{\partial \xi}\left(\frac{u^{\xi} c_k h}{J}\right) + \frac{\partial}{\partial \eta}\left(\frac{u^{\eta} c_k h}{J}\right) = \frac{q_{suk} - w_{fk} c_{bk}}{J} \tag{123}$$

II.6.5 Calculation of sorting

In order to reproduce the sorting phenomena of bed surface in non-uniform sediment case, we introduce the multilayer model proposed by Ashida, Egashira and Liu et al.[20] In this concept, the riverbed is divided into a bedload layer, a transition layer and a deposited layer (Figure II-2). The grain size distribution time series in the bed-load layer can then be calculated by the following equation:

$$\frac{\partial}{\partial t}\left(\frac{c_b E_b p_{mk}}{J}\right)+(1-\lambda)p_{bk}\frac{\partial}{\partial t}\left(\frac{z}{J}\right)+\left[\frac{\partial}{\partial \xi}\left(\frac{q_{bk}^{\xi}}{J}\right)+\frac{\partial}{\partial \eta}\left(\frac{q_{bk}^{\eta}}{J}\right)+\frac{q_{suk}-c_{bk}w_{fk}}{J}\right]=0$$

$$\begin{cases} p_{bk}=p_{tk}, \quad \dfrac{\partial z}{\partial t}\le 0, \quad E_{sd}\ge E_{be}\dfrac{c_b}{1-\lambda} \\ p_{bk}=0, \quad \dfrac{\partial z}{\partial t}\le 0, \quad E_{sd}< E_{be}\dfrac{c_b}{1-\lambda} \\ p_{bk}=p_{mk}, \quad \dfrac{\partial z}{\partial t}>0 \end{cases} \tag{124}$$

where, p_{dmk} is fraction of sediment in size class k in the m^{th} bed layer. E_{be} is the equilibrium bed load layer thickness; it is estimated by the following equation[21)].

$$\frac{E_{be}}{d_m}=\frac{1}{c_b\cos\theta(\tan\phi-\tan\theta)}\tau_{*m} \tag{125}$$

where d_m is the mean diameter of bed load, ϕ is the angle of repose, and τ_{*m} is the Shields number of mean diameter. E_{sd} is the sediment layer thickness on fixed bed. E_b is the bed load layer thickness as follows.

$$E_b=E_{be} \qquad E_{sd}\ge E_{be}\frac{c_b}{1-\lambda} \tag{126}$$

$$E_b=E_{sd}\frac{1-\lambda}{c_b} \qquad E_{sd}\le E_{be}\frac{c_b}{1-\lambda} \tag{127}$$

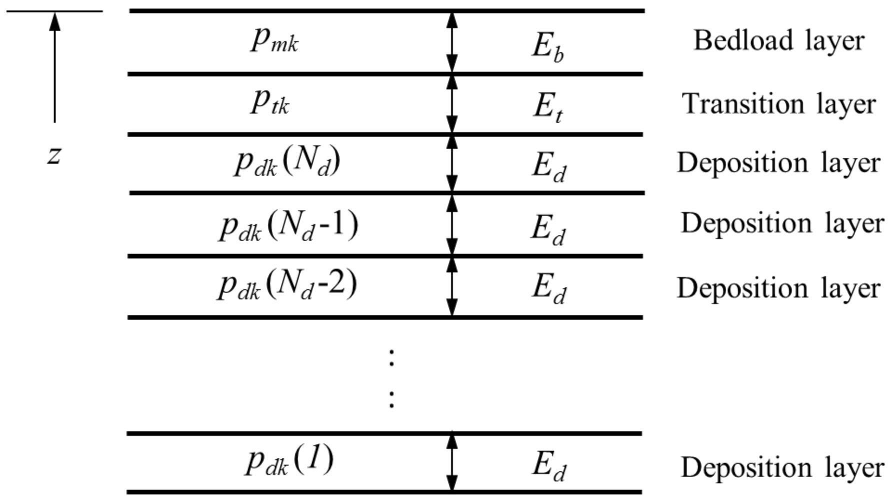

Figure II-2. Conceptual diagram of the multilayer model

In numerical calculation, we must not only consider the aggregation and degradation but also we need to take the bed deformation volume into account. For example, in the case of sedimentation, if a single step of bed deformation grows the transition layer to be higher than the deposited layer, then the thickness portion of the deposited layer is treated as a new deposited layer and the rest of the thickness is treated as a transition layer. However, in the case of erosion, if the bed deformation washes out the transition layer, then the deposited layer directly below the transition layer is treated as a new accumulation layer.

II.7 Bank erosion model

Nays2DH can deal with the bank erosion. The amount of bank erosion, which is caused by the sediment transport from the bank region and bed evolution near the bank, is estimated by using following equation[22].

$$der = \pm \frac{1}{1-\lambda} \frac{\tilde{q}_b^{\eta}}{\eta_r B_h} \Delta t + \frac{\Delta z}{\tan \theta_c} \tag{128}$$

where, *der* is bank erosion amount which defines the erosion as a positive value, Δz is the bed evolution near the bank, B_h is the bank height and Δt is the computational time step.

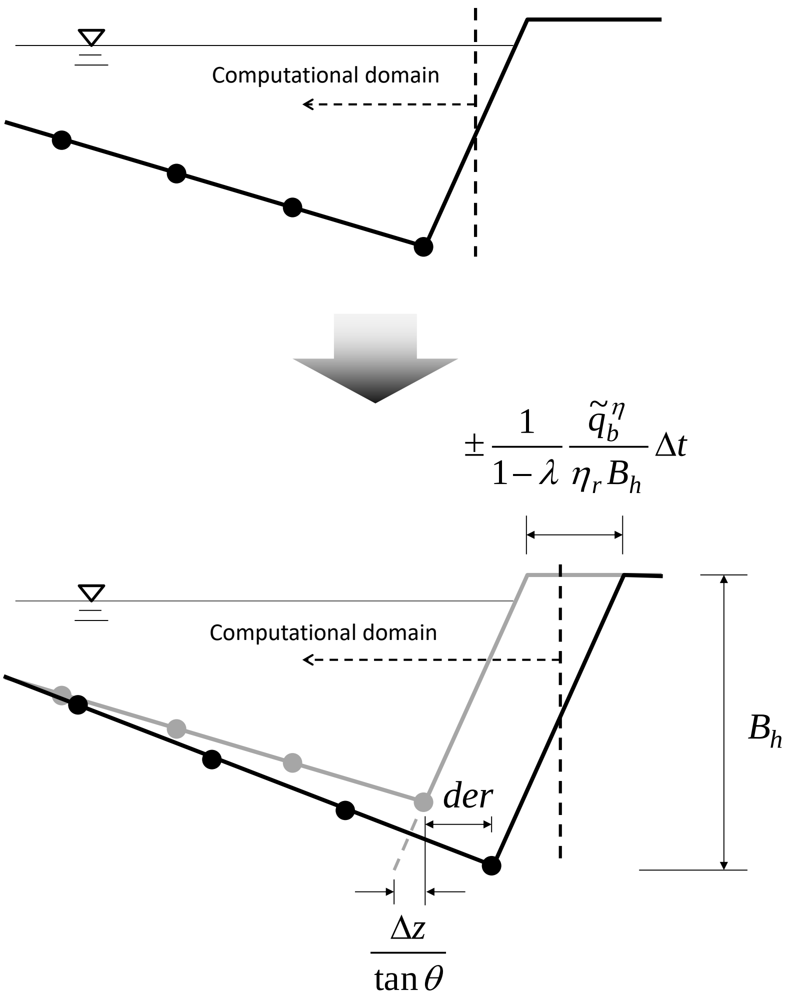

Figure II-3. Conceptual diagram of bank erosion calculation

II.8 Slope collapse model

The model calculates the bed evolution by using Exner equation [Eqs. (104) or (105)] which expresses the continuity of bed due to the sediment transport. However, the computed bed may sometimes include unrealistic steep slope over the angle of repose. This slope usually occurs between the low water channel and the floodplain, or near obstacles. Moreover, the collapse of riverbed is an important morphological evolution in the bank erosion processes[23)]. Nays2DH adopts a simple slope collapse model. This model assumes that if the computed bed slope exceeds the angle of repose (defined by the user), the bed is instantaneously corrected to satisfy the critical angle considering the mass balance.

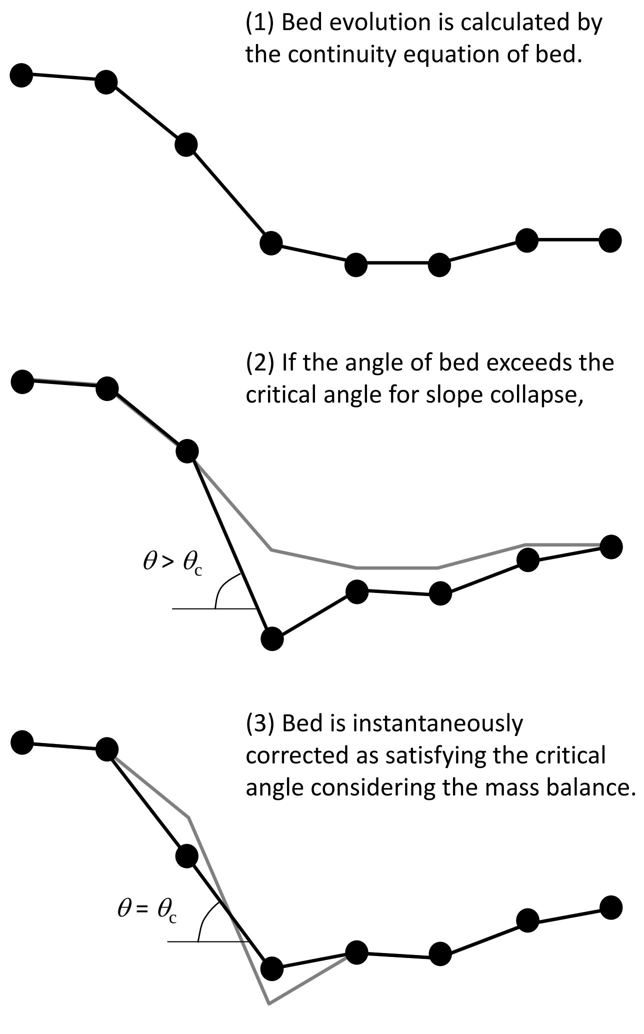

Figure II-4. A schematic diagram on the description of a slope collapse model

II.9 About the confluence model

(1) The number of incoming channels is 2 (main channel and tributary channel), and such channels must join to become a single channel at downstream.

(2) For tributary confluence, the user can select whether the tributary channel merges from the left bank or from the right bank of the main channel.

(3) For confluence calculation mesh type, the user can select between the two types shown in Figure II-5 and Figure II-6. (Type A: Branch junction, Type B: T-junction)

(4) When performing confluence calculation, [Bank erosion] cannot be selected.

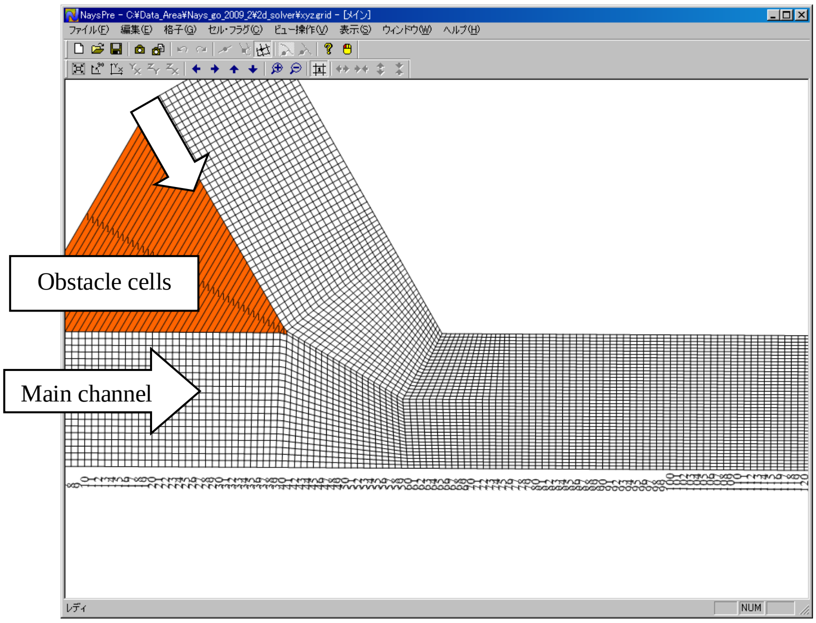

Figure II-5. Conceptual schematic of grid for mesh Type A

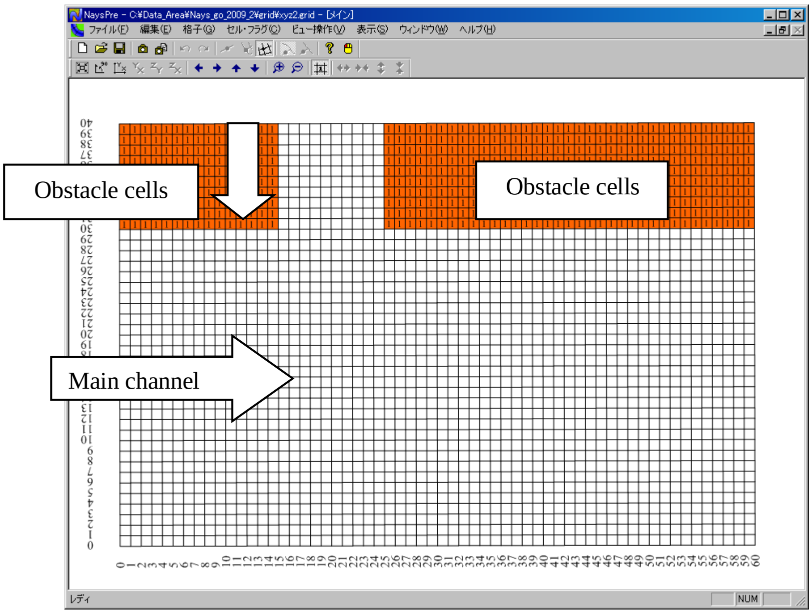

Figure II-6. Conceptual schematic of grid for mesh Type B

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III. Calculation conditions

This chapter describes calculation conditions of the Nays2DH solver with the setting dialogs of the calculation.

III.1 Setting the solver type

Nays2DH has two types of solvers: the standard edition and the advanced edition.

The standard edition is an entry-level version that allows you to perform flow regime analysis of general river segments and analysis of bed deformation. The advanced edition allows you to deal with more complex boundary conditions and initial conditions by selecting from among items such as the following: non-uniform grain size multilayer model, river confluence model and HotStart.

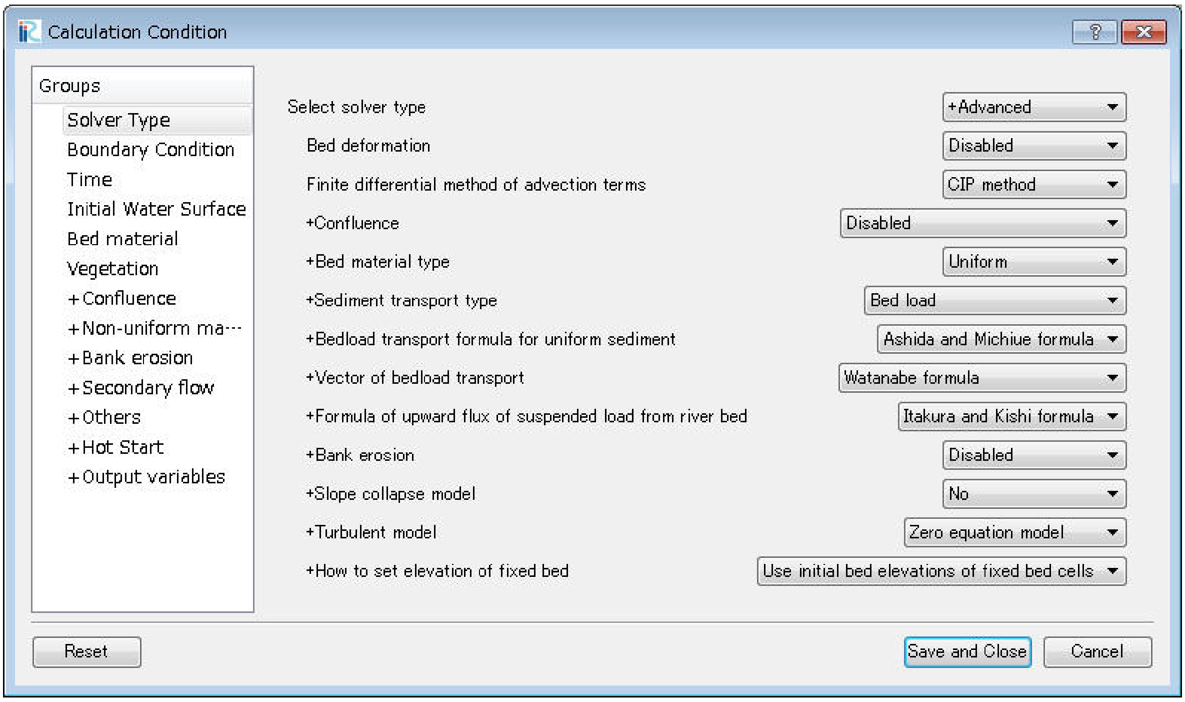


Figure III-1. Solver Type setting dialog

Table III-1. Description of Solver Type settings

#	Item	Description	Remarks
1	Solver type	Sets the solver type. Choose [Standard] or [Advanced].	When [Standard] is selected, items marked [+] cannot be set.
2	Bed deformation	Sets whether or not bed deformation is calculated.	
3	Finite-difference method of advection terms	Select the finite-difference method of the advection terms between [Upwind difference method] or [CIP method[24)]].	
4	+Confluence	Sets whether to use the confluence model. Select from among [Disabled], [Type A, from left bank], [Type B, from left bank] and [Type B, from right bank].	When using the confluence model, [Bank erosion] cannot be used.
5	+Bed material type	Sets the bed material type for bed deformation calculation. [Uniform] or [Non-uniform] can be selected.	When set to [Non-uniform], [M.P.M formula] and [Bank erosion] cannot be used.
6	+Sediment transport type	Sets the sediment transport type for bed deformation calculation. Select [Bed load only] or [Bed load and suspended load].	
7	+Bedload transport formula for uniform sediment	The user can select the bedload transport formula from [M.P.M formula] and [Ashida and Michiue formula].	When set to [M.P.M formula], [Non-uniform] cannot be used.
8	+Vector of bedload transport	The user can select how to calculate vector of bedload transport from [Watanabe formula] and [Ashida, Egashira and Liu formula].	When you select [Ashida, Egashira and Liu formula], [bank erosion] cannot be used.
9	+Formula of upward flux of suspended load from river bed	The user can select the formula of upward flux of suspended load from river bed from [Itakura and Kishi formula] and [Lane and Kalinske formula].	
10	+Bank erosion	Enables/disables bank erosion based on the sediment transport from the river bank region.	When set to [Enabled], [Non-uniform] and [Confluence] cannot be used.
11	+Slope collapse model	Enables/disables slope collapse by critical angle. The user can set the critical angle in [Bank Erosion].	
12	+Turbulent model	Sets the turbulent model type. Select from among [Constant eddy viscosity], [Zero-equation model] or [k-ε model].	

13	+How to set the elevation of fixed bed	The user can select from [use initial bed elevations of fixed bed cells] and [use elevation data of fixed bed].	When you select [use initial bed elevations of fixed bed cells], the elevation of fixed bed is set to be initial elevation of fixed bed cell if you specify fixed bed. When you select [use elevation data of fixed bed], the mapped elevation on the grid is used.

III.2 Setting the boundary conditions

Set various items regarding the boundary conditions of the upstream and downstream ends.

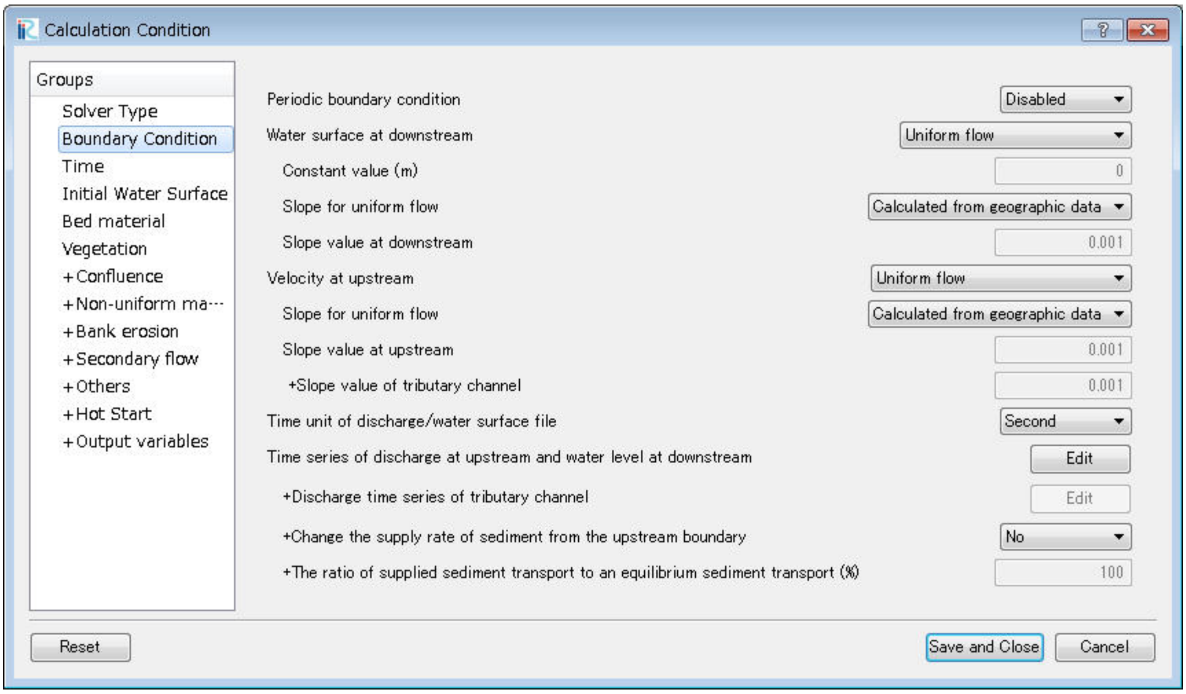


Figure III-2. Boundary conditions setting dialog

Table III-2. Descriptions of boundary condition settings

#	Item	Description	Remarks
1	Periodic boundary condition	Enables/disables the periodic boundary condition*[2].	
2	Water surface at downstream end	Sets water surface at downstream end. [Constant value], [Uniform flow], [Given from time series data] or [Free Outflow[3]] can be selected.	When set to [Given from time series data], the user should set the water level changes in [Time series of discharge at upstream and water level at downstream]
3	Constant value (m)	Enter a value if you have selected [Constant value] for [Water surface elevation at downstream end].	
4	Slope for uniform flow	Enter a value if you have selected [Uniform flow] for [Water surface elevation at downstream end]. Select [Calculated from geographic data] or [Constant value].	If [Calculated from geographic data] is selected, uniform flow calculation is done using the slope of the downstream end calculation grid.
5	Slope value at downstream end	Enter a value if you have selected [Constant value] for [Slope gradient for uniform flow simulation].	
6	Flow velocity distribution at the upstream end	Sets the velocity distribution at the upstream end. Select [Calculate from water depth at upstream end] or [Uniform flow].	
7	Slope for uniform flow	Set this if you select [Uniform flow] for [Flow velocity distribution at the upstream end]. Select [Calculated from geographical data] or [Constant value].	If [Calculated from geographic data] is selected, uniform flow calculation is done using the upstream toe slope of the calculation grid.
8	Slope value at upstream end	Enter a value if you have selected [Constant value] for [Slope gradient for uniform flow simulation].	
9	+Slope value of tributary channel	Enter a value when using the confluence model when you select [Constant value] for [Slope gradient for uniform flow simulation].	

*[2] [Periodic boundary condition] gives the hydraulic and sediment transport conditions at the downstream end to the upstream end. It is used when hydraulic conditions, channel conditions and sediment feeding conditions are of periodic nature, such as in the case of experiments. Before using it for an actual river, sufficient verification of periodicity is necessary.

[3] In [Free Outflow] condition, the water level at downstream edge is given using the depth which is calculated at next to the downstream boundary. Water level is estimated by using it and the bed elevation at the boundary edge.

10	Time unit of discharge/water surface elevation file	Sets the unit of time for the time columns of Discharge time series, Stage at downstream time series, and +Discharge time series of tributary channel. Select [Second] or [Hour].	
11	Time series of discharge at upstream and water level at downstream	Click [Edit] and set the discharge hydrograph and the water level variation	Select [Second] or [Hour]. The unit of discharge is "m^3/s."
13	+Discharge time series of tributary channel	Sets the discharge hydrograph of tributary channel when the confluence model is used.	Select [Second] or [Hour]. The unit of discharge is "m^3/s." The same data number and start-end time is required between the main and tributary channel.
14	+Change the supply rate of sediment from the upstream boundary	The user can change the supply rate of sediment from the upstream boundary.	When you select [Periodic boundary condition], you cannot select this option.
15	+The ratio of supplied sediment transport to an equilibrium sediment transport (%)	The ratio of supplied sediment transport rate to an equilibrium sediment transport rate can be defined.	Unit for this parameter is %.

III.3 Setting the time-related conditions

Set various conditions that relate to time.

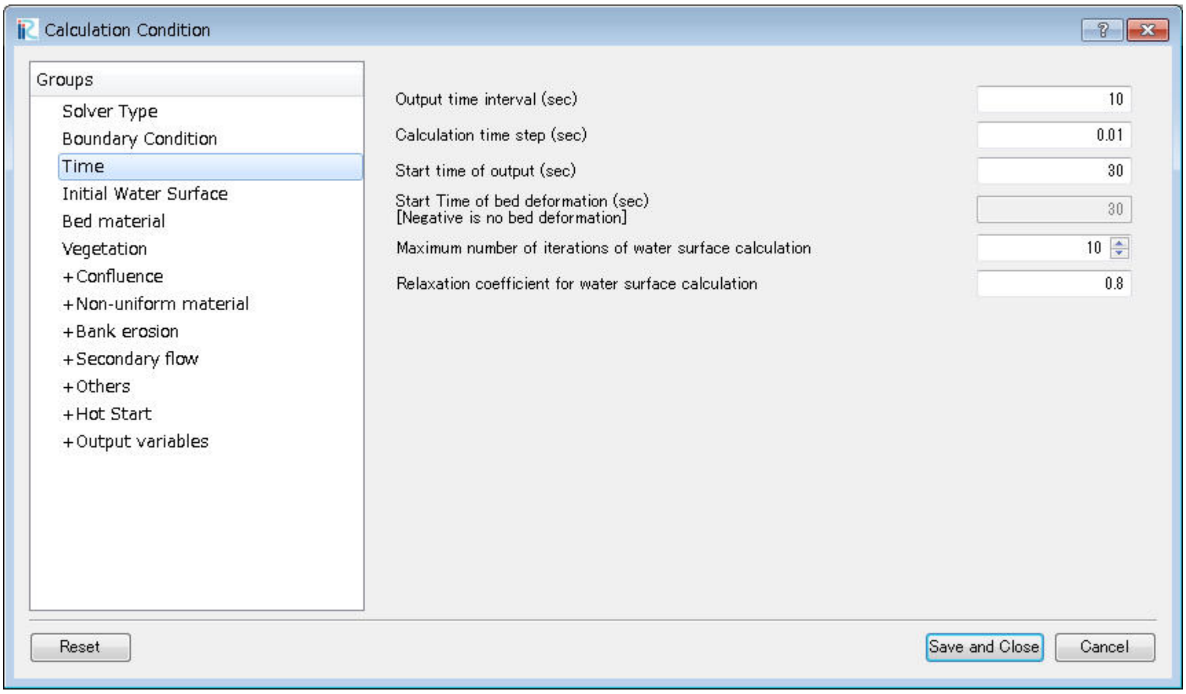


Figure III-3. Dialog for setting time-related conditions

Table III-3. Descriptions of time-related settings

#	Item	Description	Remarks
1	Output time interval (sec)	Sets the time interval at which calculation results are to be output.	
2	Calculation time step (sec)	Sets the time interval of calculation steps.	This is an important parameter that determines calculation efficiency and stability.
3	Start time of output (sec)	Sets the time to start outputting calculation results.	
4	Start time of bed deformation (sec)	Sets the time to start bed deformation calculation.	If a negative sign is specified, bed deformation calculation is not performed.
5	Maximum number of iterations of water surface elevation calculation	Sets the number of internal iterations when water surface elevation is calculated.	If water surface elevation calculation is unstable, adjust this setting.
6	Relaxation coefficient for water surface elevation calculation	This is the coefficient that is used when water surface level calculation is performed.	

III.4 Setting the initial water surface conditions

Set various conditions that relate to initial water surface profile.

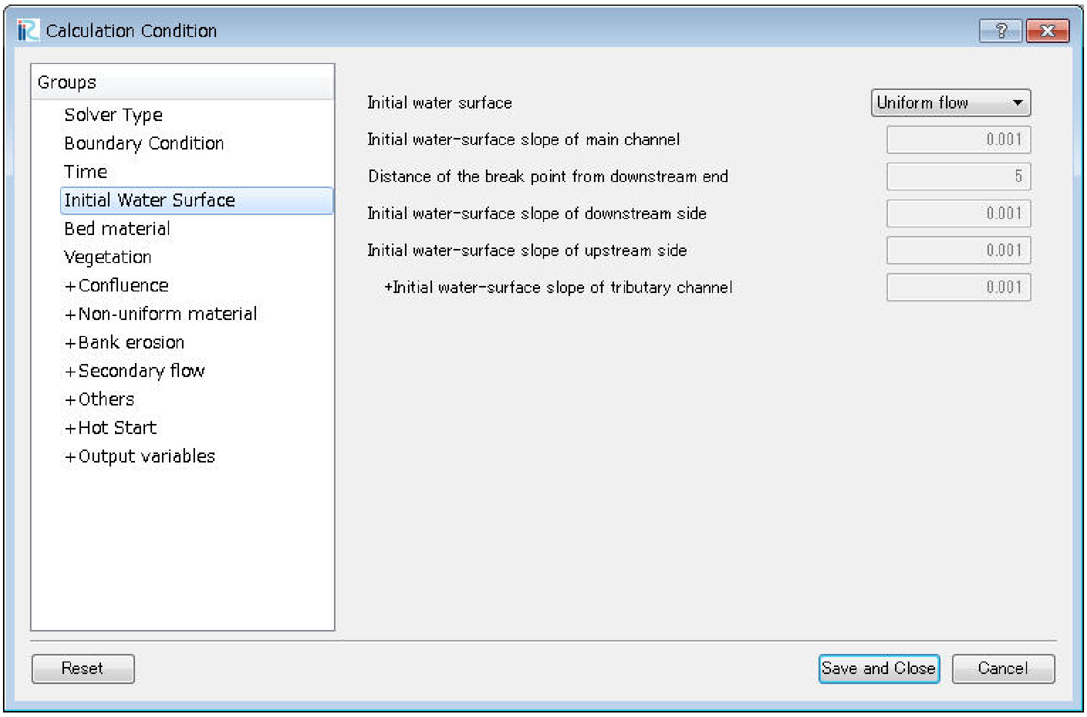


Figure III-4. Dialog for setting the initial water surface

Table III-4. Description of the initial water surface settings

#	Item	Description	Remarks
1	Initial water surface	Select the setting method for the initial water surface. Select from among [Constant slope (straight line)], [Line], [Uniform flow] and [Non-uniform].	
2	"Initial water surface slope of main channel"	This can be specified only when [Constant slope (straight line)] has been selected for [Initial Water Surface Profile].	
3	Distance of the slope break point from the downstream end	This can be specified only when [Line] has been selected for [Initial Water Surface Profile].	The unit is (m).
4	Initial water surface slope of the downstream end	This can be specified only when [Line] has been selected for [Initial Water Surface Profile].	
5	Initial water surface slope of the upstream end	This can be specified only when [Line] has been selected for [Initial Water Surface Profile].	

6	+Initial water surface slope of tributary channel	This can be specified when using the confluence model and when [Constant slope (straight line)] has been selected for [Initial Water Surface Profile].	

III.5 Setting Bed material

Set various conditions that relate to Bed material.

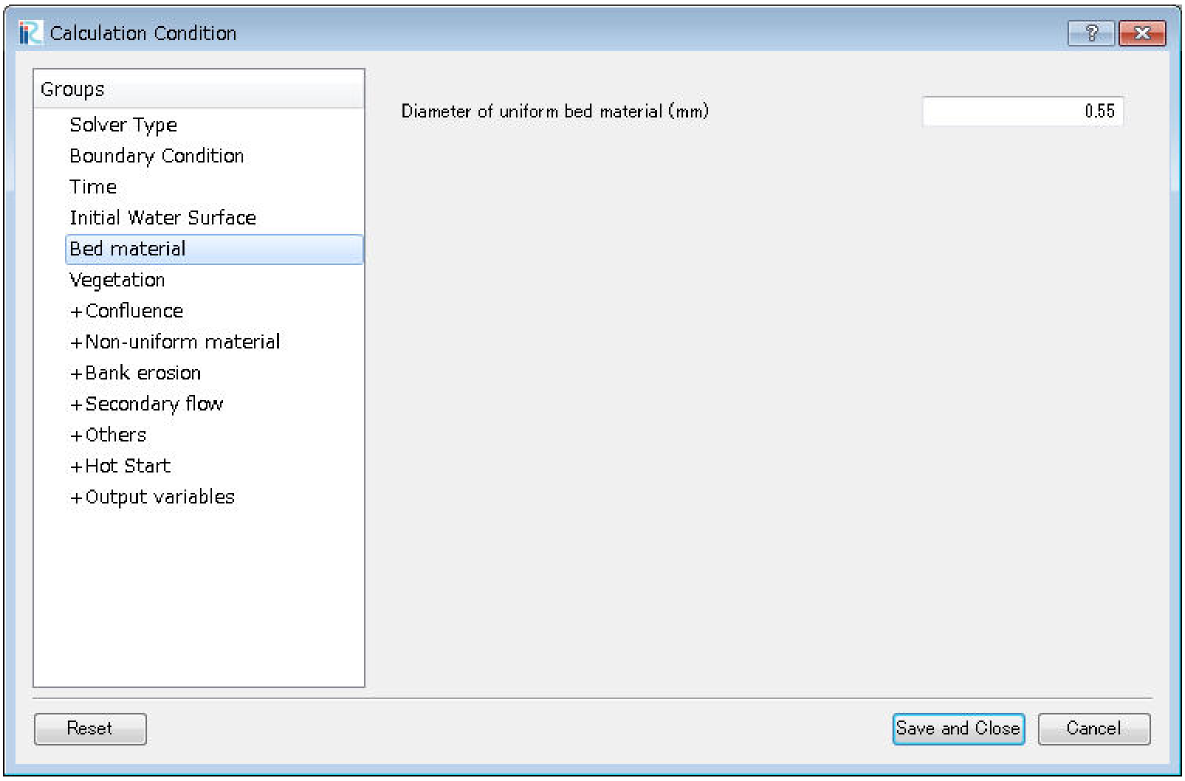


Figure III-4. Window for setting the roughness

Table III-4. Description of the roughness settings

#	Item	Description	Remarks
1	Diameter of uniform bed material (mm)	Enter a bed material grain size to be used for bed deformation calculation.	The unit is mm.

III.6 Setting vegetation conditions

Set various conditions that relate to vegetation.

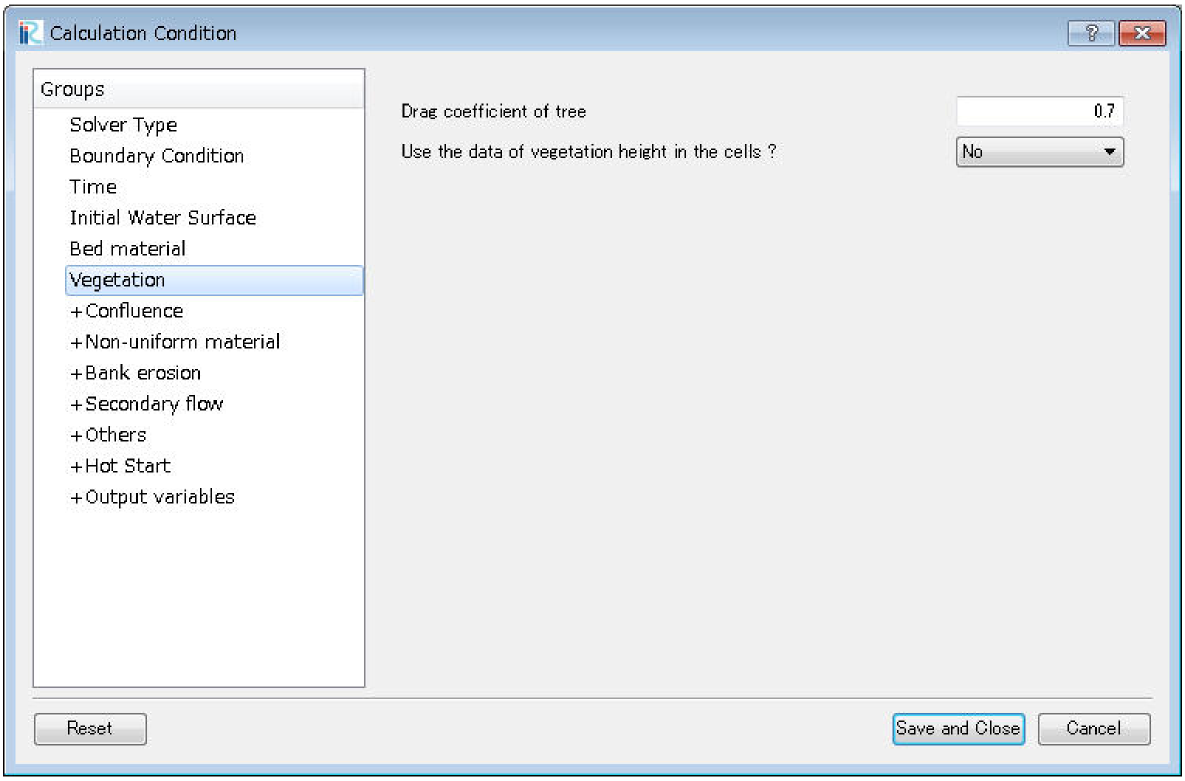


Figure III-6. Vegetation setting dialog

Table III-6. Description of the vegetation settings

#	Item	Description	Remarks
1	Drag coefficient of tree	Enter tree drag coefficient.	
2	Use the data of vegetation height in the cells?	The user can select whether the data of vegetation height is used or not.	When set to [Yes], the vegetation height mapped in the computational cells is effective. When set to [No], unsubmerged condition is set in the computation.

III.7 Setting the confluence information

Set various conditions related to confluence information. These parameters on this window depend on the confluence type in which you set shown in Fig. III-8.

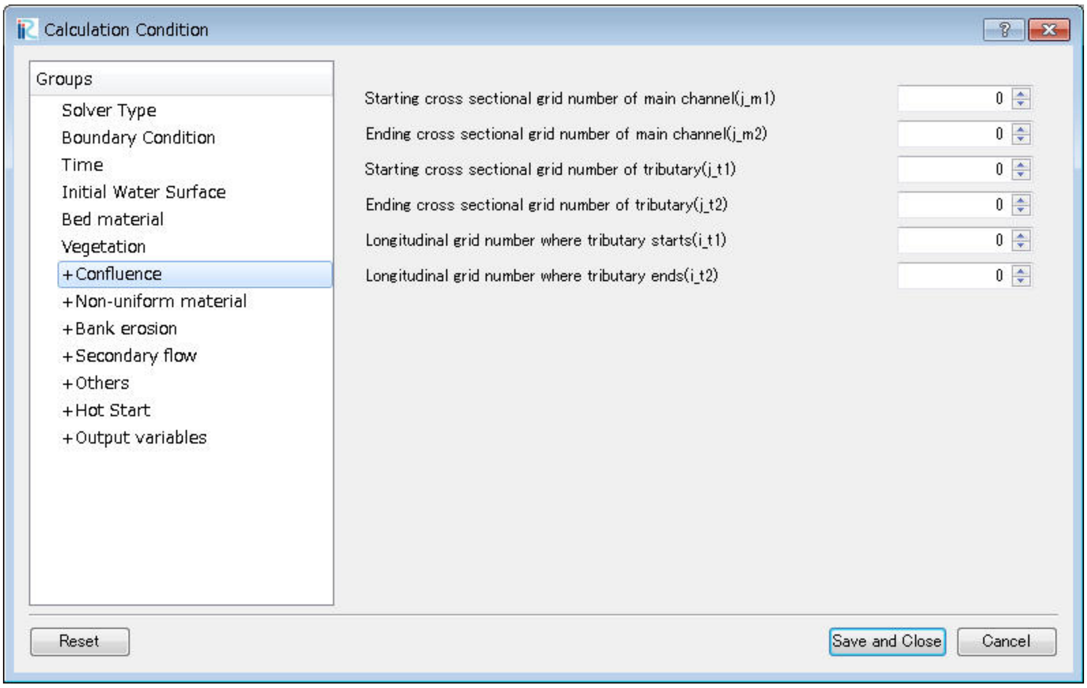


Figure III-7. Confluence information setting dialog

Table III-7. Description of the confluence information

#	Item	Description	Remarks
1	Main channel transverse grid number where simulation starts (j_m1)	See Figure III-8.	
2	Main channel transverse grid number where simulation finishes (j_m2)	See Figure III-8.	
3	Tributary transverse grid number where simulation starts (j_t1)	See Figure III-8.	
4	Tributary transverse grid number where simulation finishes (j_t2)	See Figure III-8.	
5	Upstream-end longitudinal grid number where the main channel and the tributary merge and calculation starts (i_t1)	See Figure III-8.	
6	Downstream-end longitudinal grid number where the main channel and the tributary merge and calculation finishes (i_t2)	See Figure III-8.	

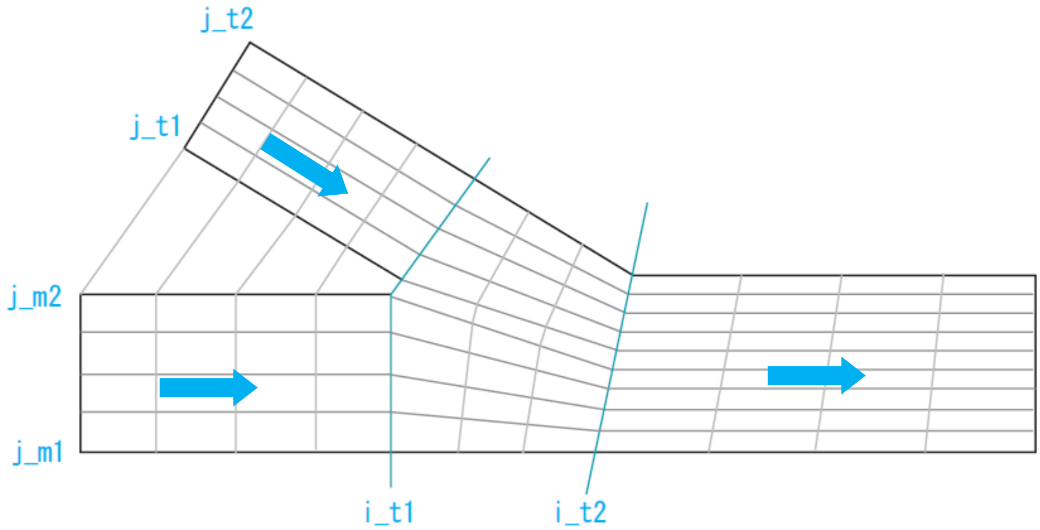


Figure III-8. (1) Conceptual diagram of confluence information (mesh Type A, merging from left bank)

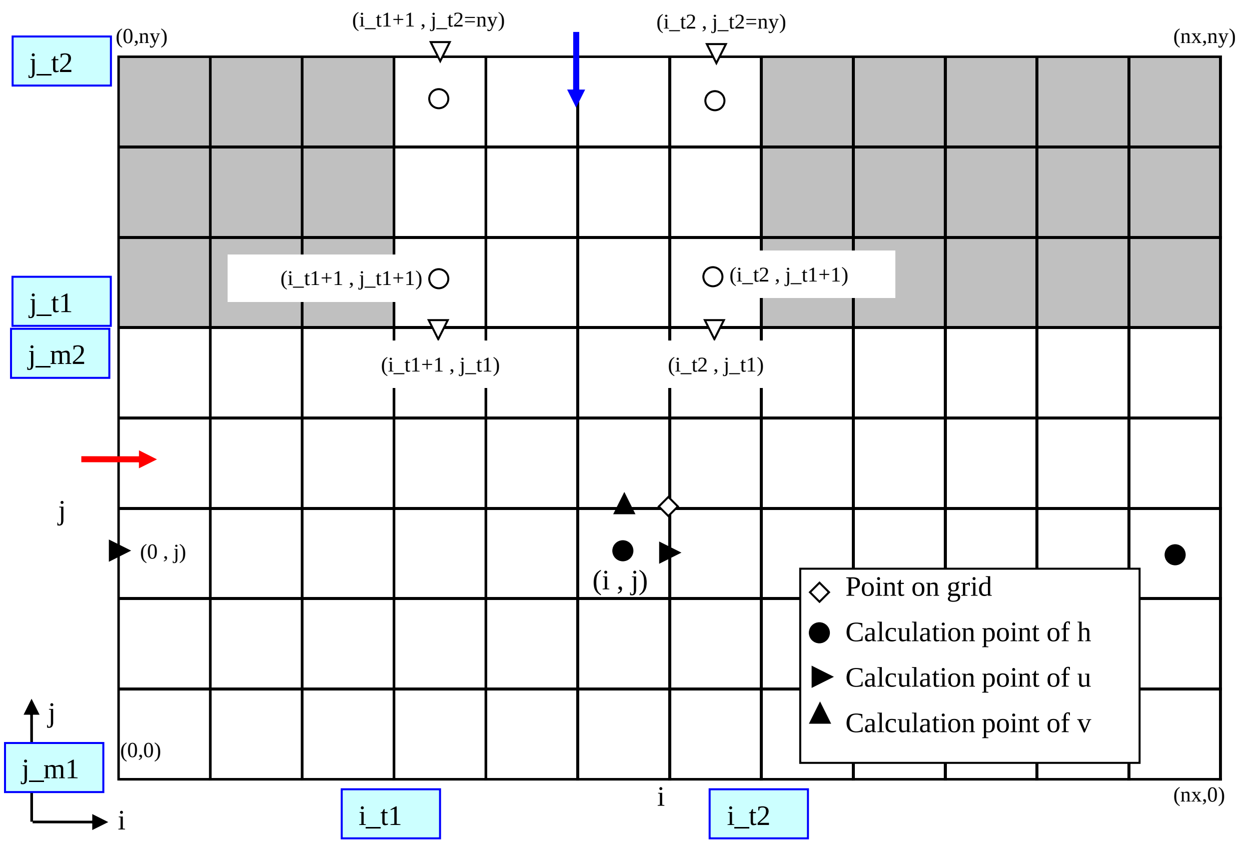


Figure III-8. (2) Conceptual diagram of confluence information (mesh Type B, merging from the left bank)

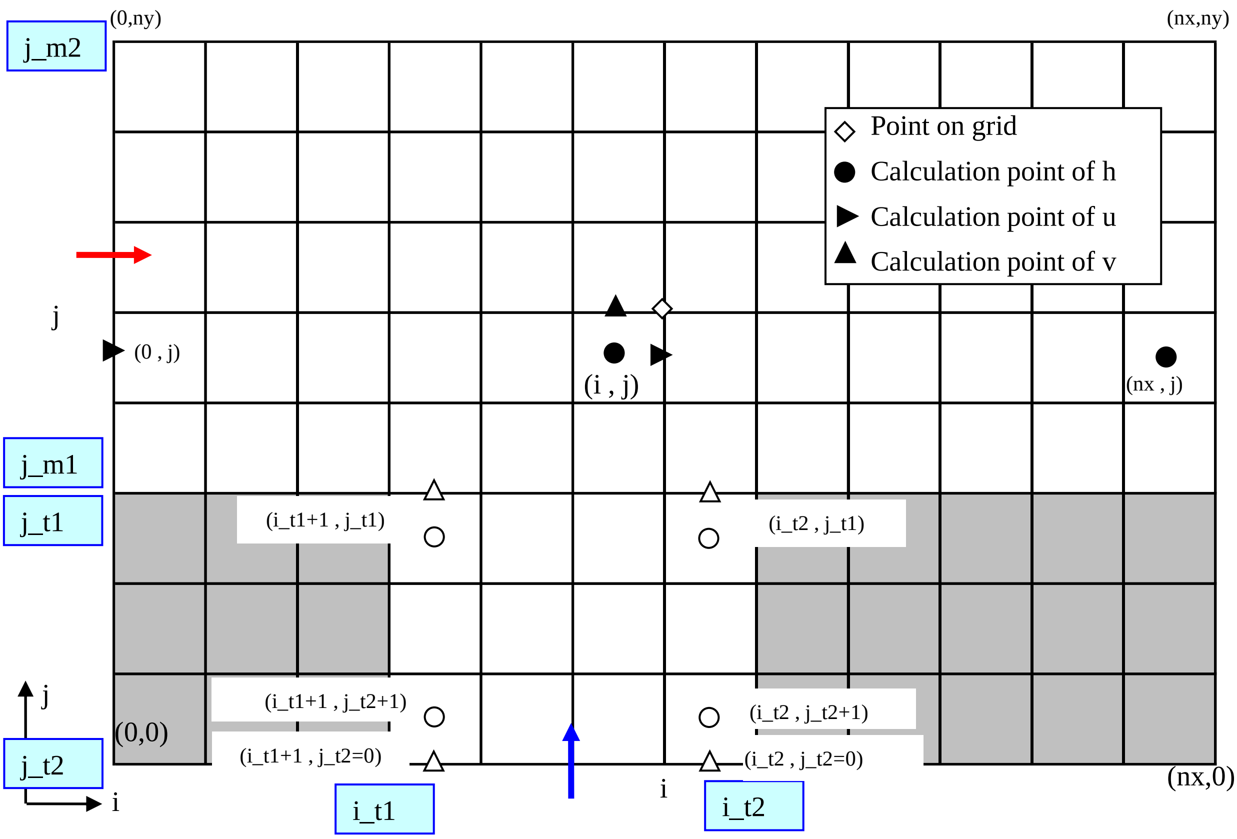


Figure III−8. (3) Conceptual diagram of confluence information (**mesh** Type B, merging from the right bank)

III.8 Setting non-uniform material information

Set various conditions related to non-uniform material.

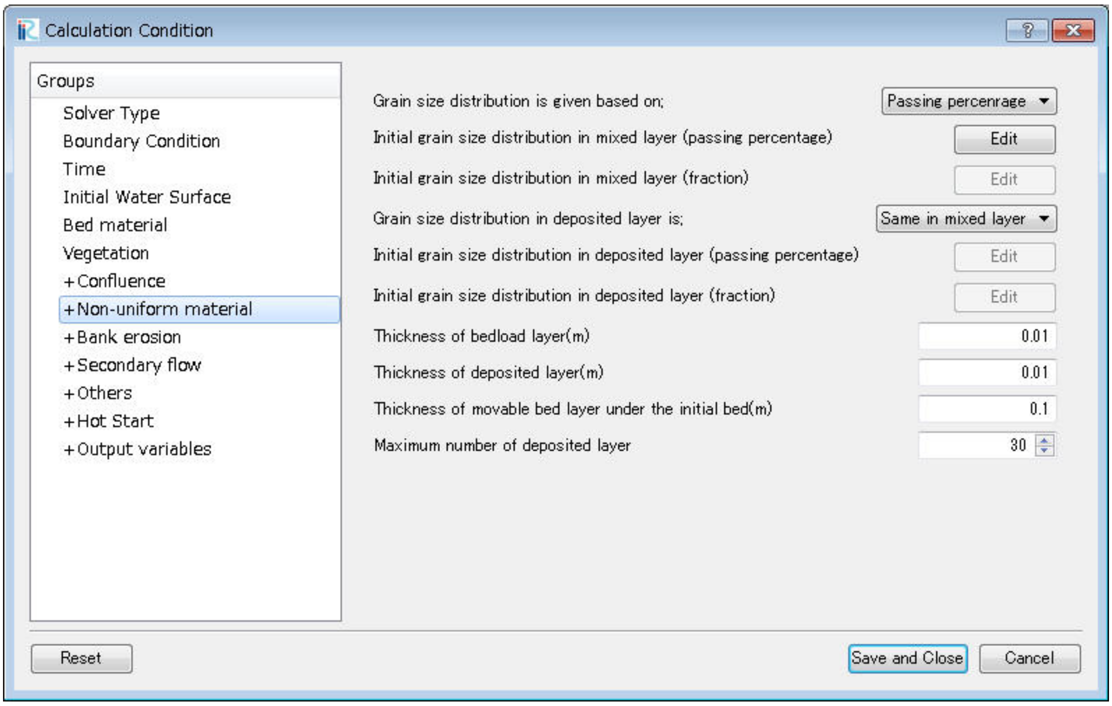


Figure III-9. Non-uniform material information setting dialog

Table III-8. Description of the non-uniform material information

#	Item	Description	Remarks
1	Grain size distribution is given base on;	You can select the initial grain size distribution from [Passing percentage] and [Fraction].	
2	Initial grain size distribution in mixed layer (percentage passing)	Initial grain size distribution in mixed layer is given based on passing percentage. In the first line, grain size (unit: mm) is given, and from the second line the passing percentage of each grain size (%) is given for the region defined in the cell condition.	See Figure III-10
3	Initial grain size distribution in mixed layer (fraction)	Initial grain size distribution in mixed layer is given based on fraction of each grain size. In the first line, grain size (unit: mm) is given, and from the second line the fraction of each grain size (%) is given for the region defined in the cell condition.	See Figure III-11

4	Grain size distribution in deposited layer is:	Initial grain size distribution in deposited layer is given from [same in deposited layer] or [Given]. When you select [Given], you can set the grain size distribution by same method in setting for the mixed layer.	
5	Initial grain size distribution in deposited layer (percentage passing)	Initial grain size distribution in deposited layer is given based on passing percentage. In the first line, grain size (unit: mm) is given, and from the second line the passing percentage of each grain size (%) is given for the region defined in the cell condition.	See Figure III-10
6	Initial grain size distribution in deposited layer (fraction)	Initial grain size distribution in deposited layer is given based on fraction of each grain size. In the first line, grain size (unit: mm) is given, and from the second line the fraction of each grain size (%) is given for the region defined in the cell condition.	See Figure III-11
7	Thickness of bedload layer (m)	Enter the thickness of bedload layer.	
8	Thickness of deposited layer (m)	Enter the deposited layer thickness into the multilayer model that is used to reproduce bed deformation.	
9	Thickness of movable bed layer (m)	Enter the thickness to be considered with regard to bed deformation.	
10	Maximum number of deposited layers	Enter a layer count to specify the maximum number of deposited layers that may be stored.	

Note 1) When a multilayer model is used for riverbed simulation, a large memory capacity is required, because the number of parameters is huge. The number depends on "grain of each size as a proportion of all the grains in each layer" x "the number of stored bed layers" x "the number of grids." In some cases, we may run out of memory. Therefore, we place some limitations on the number of layers to be considered.

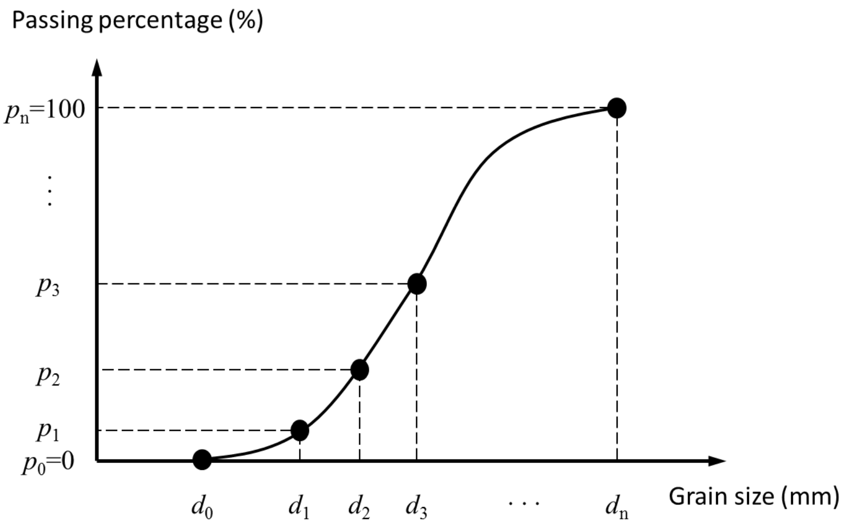


Figure III-10. Setting the initial grain size distribution by grain size distribution curve

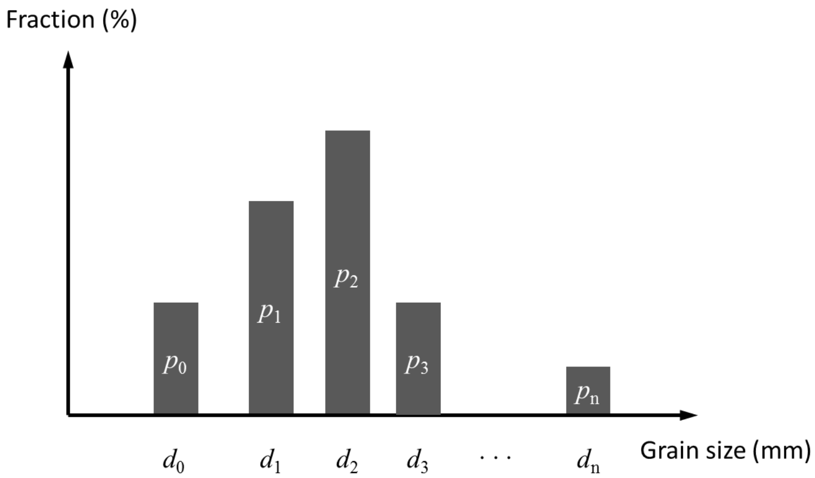


Figure III-11. Setting the initial grain size distribution by fraction

III.9 Setting the bank erosion information

Set various conditions regarding bank erosion information.

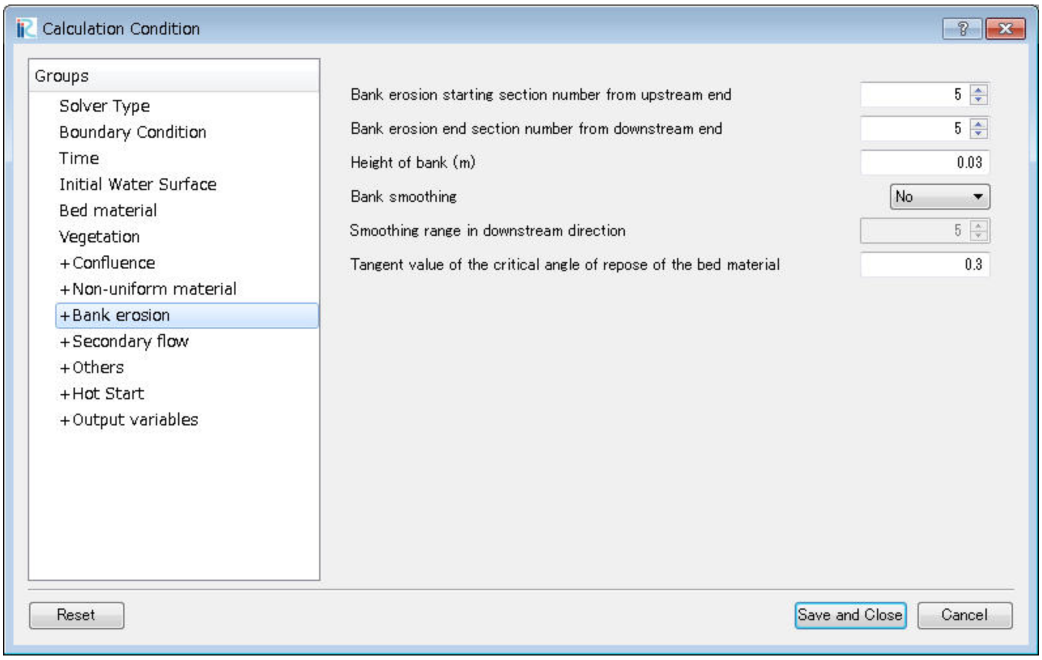


Figure III-12. Bank erosion information setting dialog

Table III-9. Description of the bank erosion information

#	Item	Description	Remarks
1	Bank erosion starting section number from upstream end	Enter the upstream-end cross section number of the bank erosion calculation section.	
2	Bank erosion end section number from downstream end	Enter the downstream-end cross section number of the bank erosion calculation section.	
3	Height of bank (m)	Enter the bank height on the outside of the grid (in the cross-sectional direction).	
4	Bank smoothing	If bank erosion occurs on the left and right ends of the calculation grid ($j=1$ and $j=ny$), the calculation area will be expanded and the calculation grid relocated. Select whether or not to smooth the calculation grid in such a case.	
5	Smoothing range in the longitudinal direction	Enter the number of calculation grids to be taken into account in the longitudinal direction when performing calculation grid smoothing at one point.	

6	Tangent of submerged angle of repose of the bed material	Enter the tangent value of the tangent of submerged angle of repose of the bed material. It is typically around tan 30 degrees.	It becomes the threshold of bank erosion and slope-collapse model.

III.10 Settings the calculation of secondary flows

Set the calculation of secondary flows which has an important role for determining the direction of bedload transport.

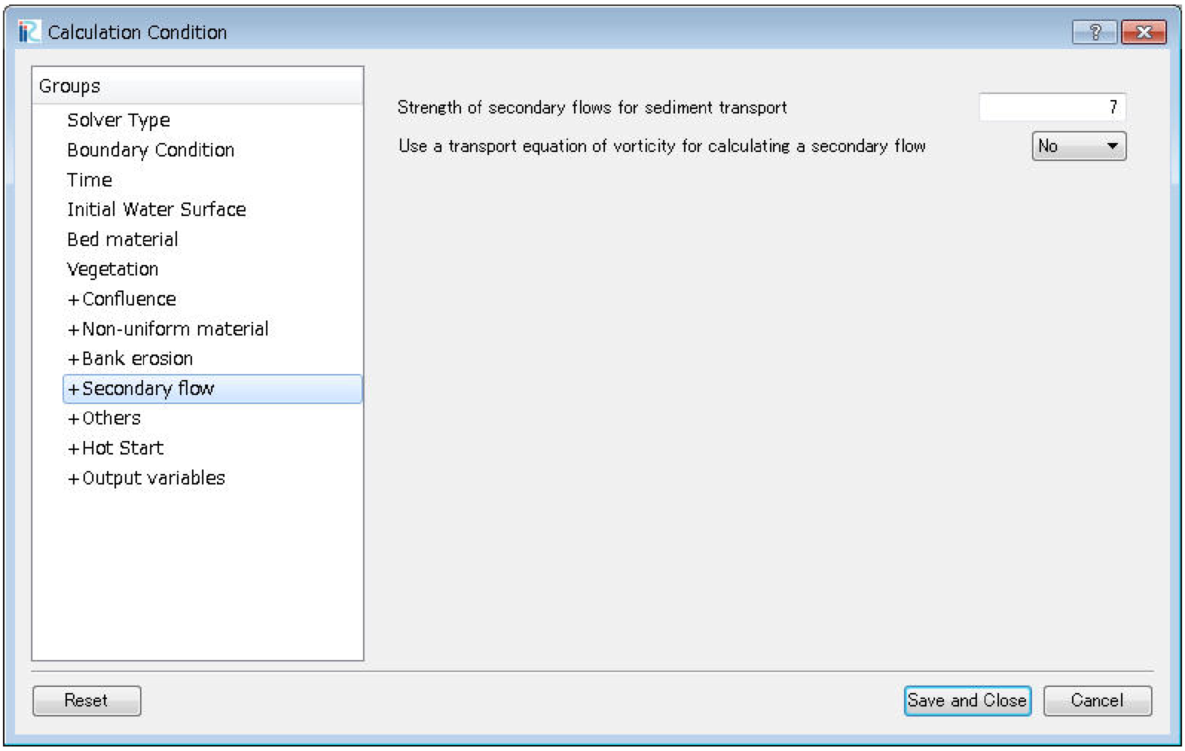


Figure III-13. Secondary flows setting dialog

Table III-10. Description of settings of secondary flows

#	Item	Description	Remarks
1	Strength of secondary flows for sediment transport	This is a coefficient for secondary flow calculation.	See Eq. (77) in Chapter 2.
2	Use a transport equation of vorticity for calculating a secondary flow	When set to [Yes], the strength of secondary flows is calculated by solving the transport equation of depth-averaged vorticity in streamwise direction. When the user choose [No], an equilibrium type model is applied.	See Chapter 2: 5.4

III.11 Other settings

Set various other conditions.

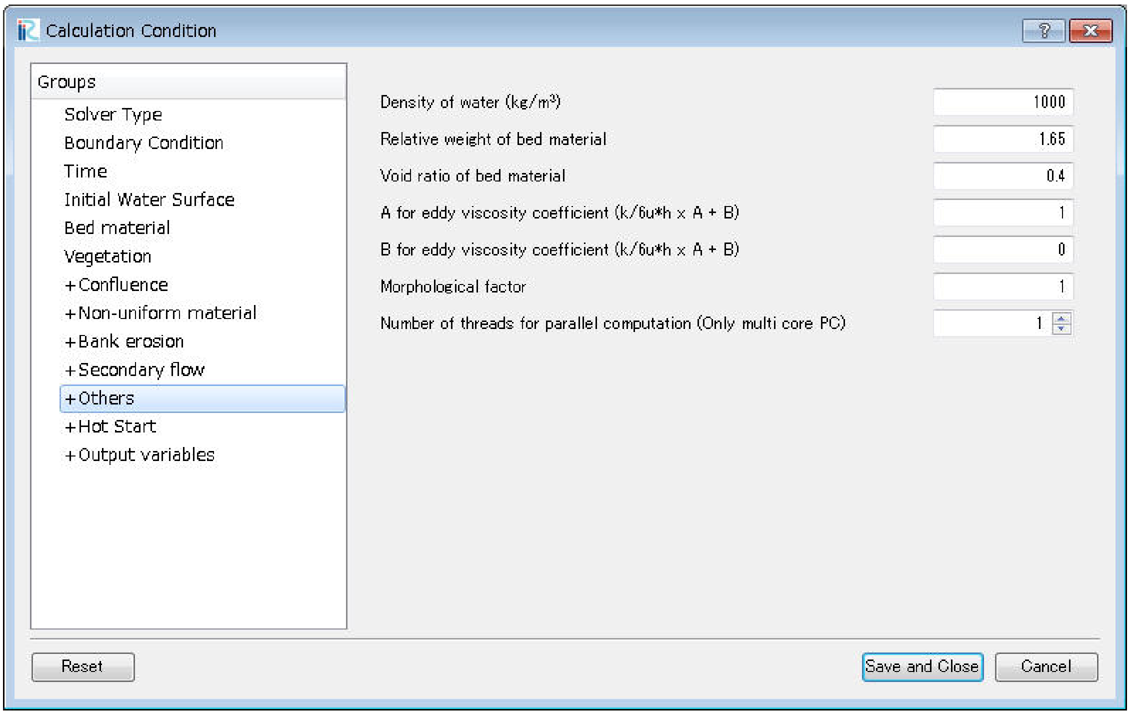


Figure III-14. Dialog for setting other information

Table III-11. Description of **the** other information

#	Item	Description	Remarks
1	Density of water (kg/m^3)	Enter the density of water in kg/m^3.	
2	Specific weight of bed material	Enter the specific weight of bed material.	
3	Void ratio of bed material	Enter the void ratio of bed material.	
4	*A* for eddy viscosity coefficient ($k/6\ u^*\ h\ x\ A + B$)	The user defined parameter [*A*] for an eddy viscosity coefficient can be set.	See Eq. (45) in chapter 2. Default value is 1.
5	*B* for eddy viscosity coefficient ($k/6\ u^*\ h\ x\ A + B$)	The user defined parameter [*B*] for an eddy viscosity coefficient can be set.	See Eq. (45) in chapter 2. Default value is 0.
6	Morphological factor	Ratio of Bed Deformation Δt to Flow Δt is set. Basically, users should use 1. In case, the unsteady characteristics of the flow are very week and users want to get faster, please set the value between 1 and 10.	

7	Number of threads for parallel computation (Only multi core PC)	The user can set number of CPU for the computation.	Be aware that number of CPU in the computation has to be less than the number of CPU in the PC.

III.12 Setting HotStart

Set various conditions of HotStart. Using this function, calculation results can be carried over or recalculated. For example, when performing calculations for a flood over a long period of time, this function is useful for observing the entire hydrograph with large output intervals, while visualizing the flow regime in detail with smaller output intervals in and around peak periods.

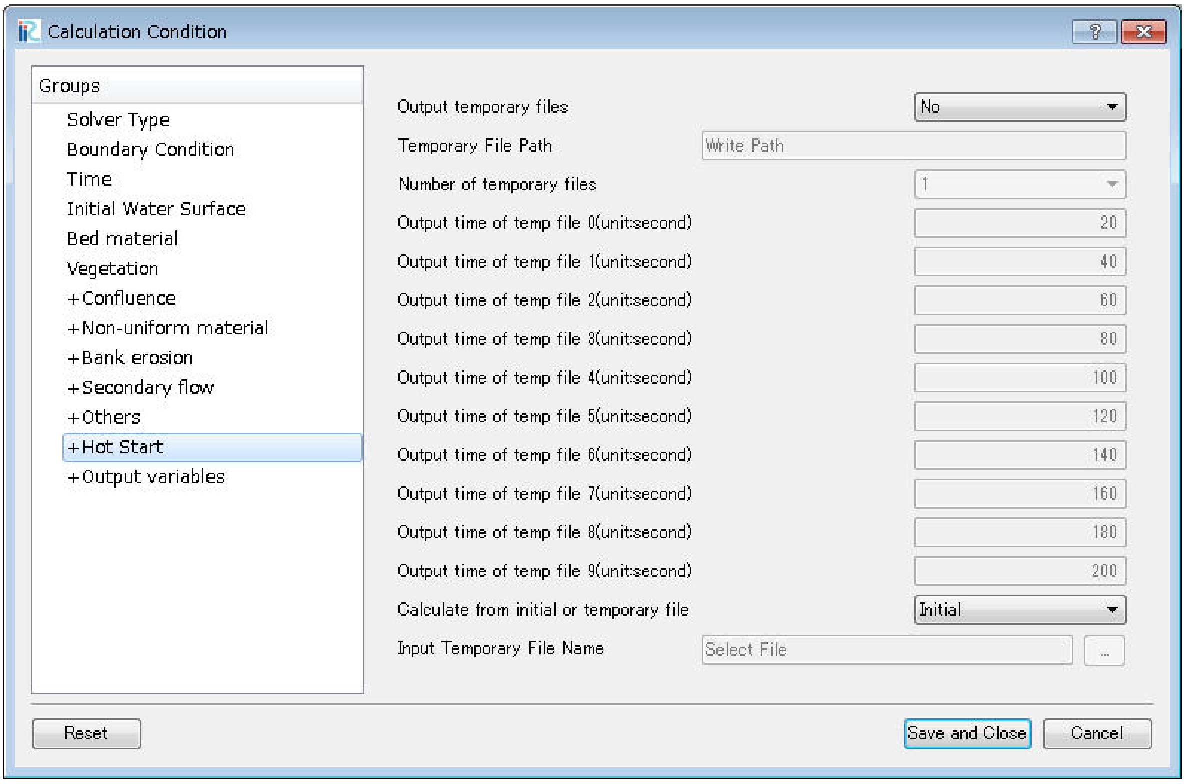


Figure III-15. HotStart setting window

Table III-12. Description of HotStart

#	Item	Description	Remarks
1	Output temporary files	Choose [Yes] if you are going to use HotStart.	
2	Number of temporary files	Up to ten temporary files can be created.	
3	Output time of temp files 0 to 9 (unit: second)	Enter calculation time at which to output each temporary file.	
4	Calculation from initial file or a temporary file	If you want to input from a temporary file and make recalculation, select [Read temp file and restart].	
5	Input Temporary File Name	Select a temporary file to input.	

III.13 Setting output variables

You can select the output variables for visualization on the following dialog. This setting contributes to reduce the file size. Some fundamental variables such as water depth, elevation are always output to the file.

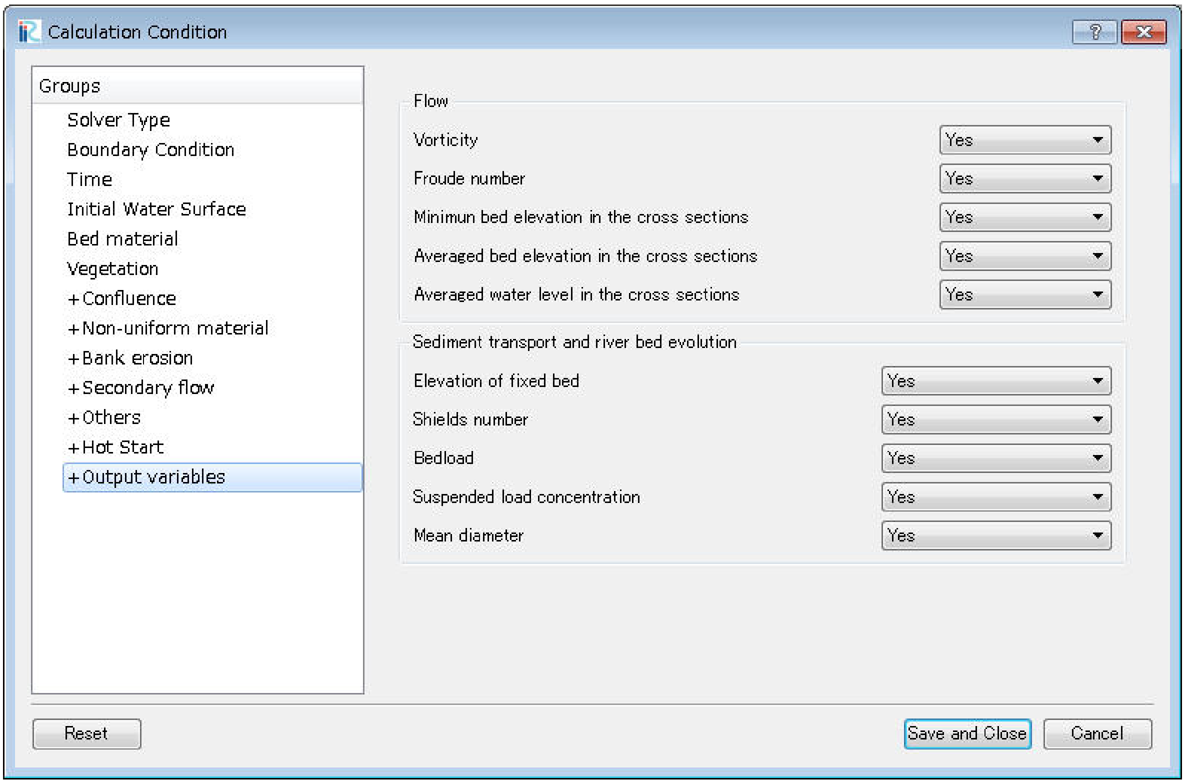


Figure III-16. Output variables setting window

IV. Setting the grid and cell conditions

The user can set the following grid conditions; 1. [Elevation] and 2. [Elevation of fixed bed], and also following cell conditions in the computations; 1. [Obstacles], 2. [Fixed or Movable cell], 3. [Density of Vegetation], 4. [Height of vegetation], 5. [Manning's roughness coefficient] and 6. [Grain size distribution]. The setting the grid and cell condition can be set in [Object Browser] on the pre-processing window.

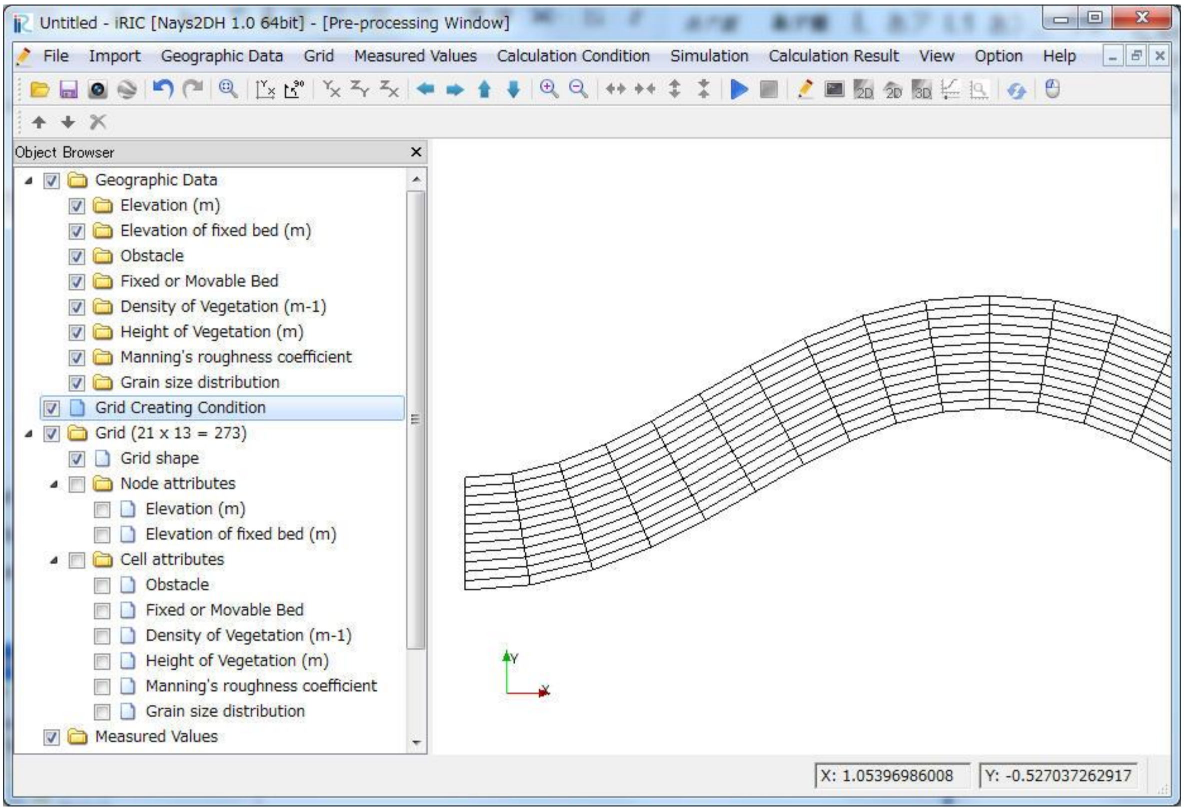


Figure IV－1　The setting the cell conditions in Pre-processing windows

There are two methods to set the grid and cell conditions; one is direct setting and the other is use the mapping function by using polygons. For the direct setting the grid and cell conditions, you should check [Cell attributes] and cell condition of which you want to set the condition (Figure shows an example for the setting the obstacles). Then, by right clicking on the computational grid and clicking [Edit value] shown in Fig. IV-2, you can set the cell conditions. When you use function of polygons for mapping the physical value to the grids or cells, please also see the iRIC user manual.

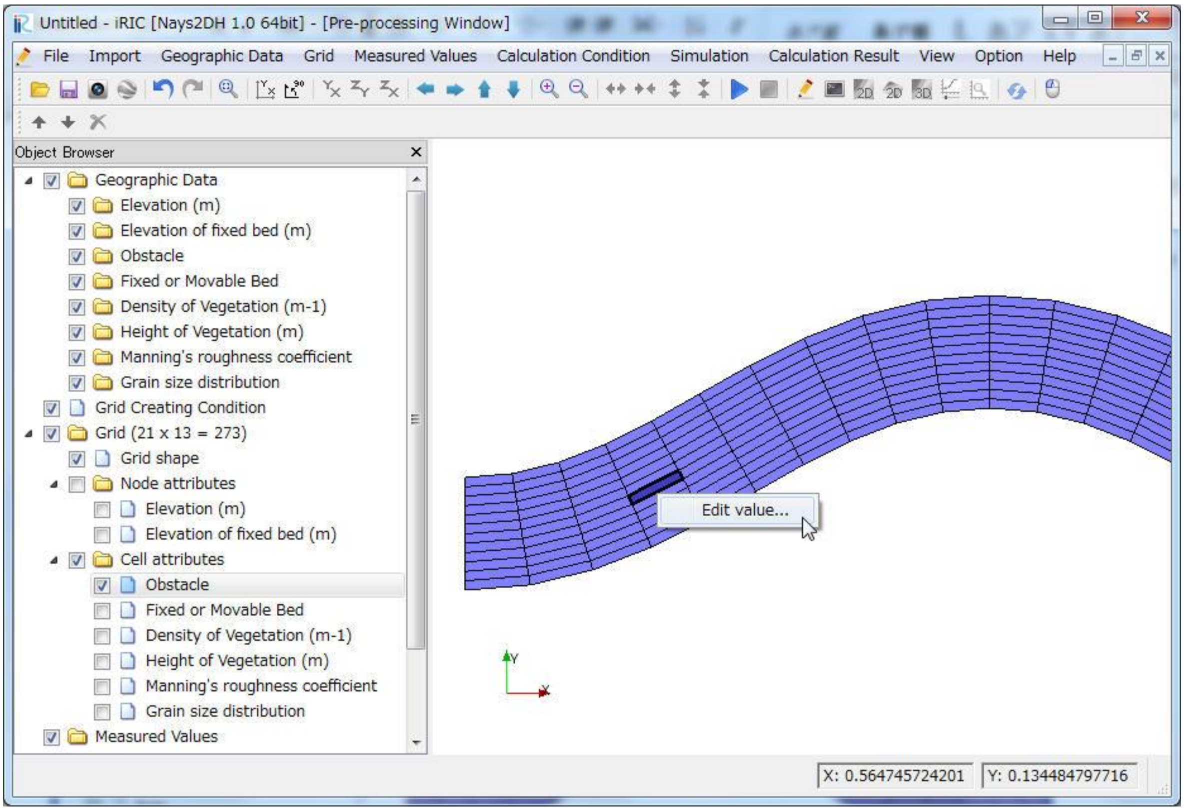


Figure IV－2 Setting cell conditions. Figure shows an example for the setting [Obstacles].

IV.1 Setting the grid conditions

In Nays2DH, you can set the bed elevation and the elevation of fixed bed as the grid condition. When you set the elevation data of fixed bed is effective in the computational settings (see the settings solver type, item [How to set the elevation of fixed bed]), the bed elevation have to be higher than the elevation of fixed bed. If you set the elevation of fixed bed higher than the bed elevation, the elevation of fixed bed is set to be the bed elevation.

When the fixed bed is set in the computational cells, the elevation of fixed bed is equal to the initial bed elevation. In the other cells, the elevation of fixed bed is set to be infinity low value.

IV.2 Setting the cell conditions

1. Edit Obstacle

The user can set the cell as [Normal cell] or [Obstacle] in [Edit Obstacle] window. In the obstacle cells, the flux for both flow and sediment transport will be set to zero.

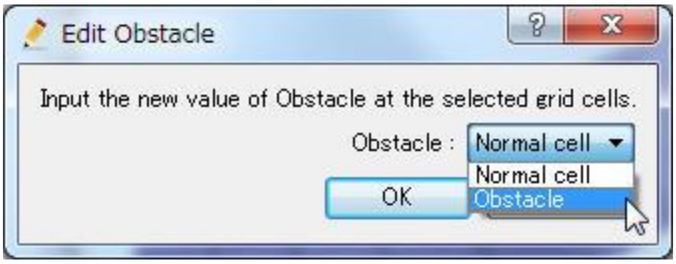


Figure IV－3 Edit Obstacle

2. Edit Fixed or Morabe Bed

The user can set the cell as [Movable bed] or [Fixed bed] in [Edit Fixed or Movable Bed] window. In the fixed bed cells, the bed elevation does not eroded under the initial bed elevation.

Figure IV－4 Edit Fixed or Morabe Bed

3. Edit Density of Vegetation

The user can set the density of vegetation in the following window. Zero means no vegetation in the computational cell. This value cannot be set as negative. In the case of negative inpout, the solver will give an error message.

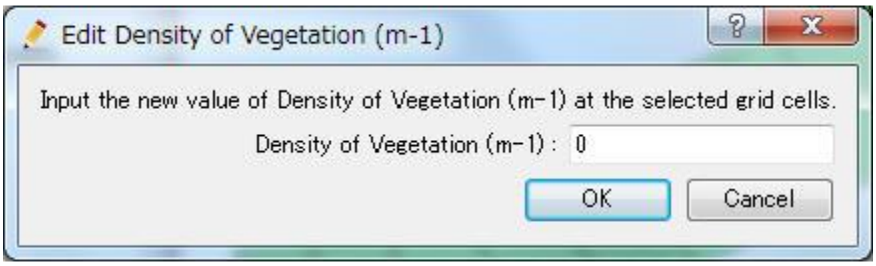


Figure IV－5 Edit Density of Vegetation

4. Edit Height of Vegetation

The user can set the height of vegetation in the following window. This value is not allowed to be negative. To use the vegetation height, the option [Yes] in [Use the data of vegetation height in the cells?], in the computational settings, has to be selected (see section 6 in Chapter 3). When the data of vegetation height is not effective, the unsubmerged vegetation is assumed in the computation.

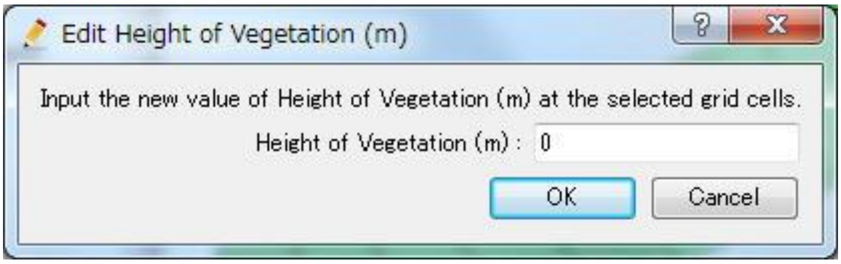


Figure IV－6 Edit Height of Vegetation

5. Edit Manning's roughness coefficient

The user can set the Manning's roughness coefficient in the following window. The user have to set this value over zero. Besides, if the user imports the computational grid from file with format [.csv] or [.grid], the cell condition will set to a default value, therefore, the user needs to adjust it.

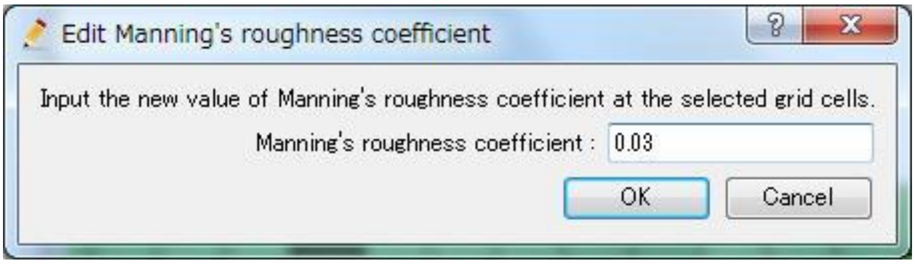


Figure IV－7 Edit Manning's roughness coefficient

6. Edit Grain size distribution

The user can set the regions of grain size distribution which was set in the computational settings. The user can set these regions up to ten. The grain size distribution which corresponds to these regions have to set in the [initial grain size distribution] (see section 8 in Chapter 3).

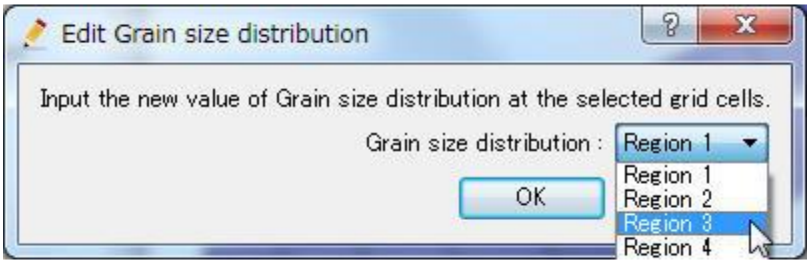
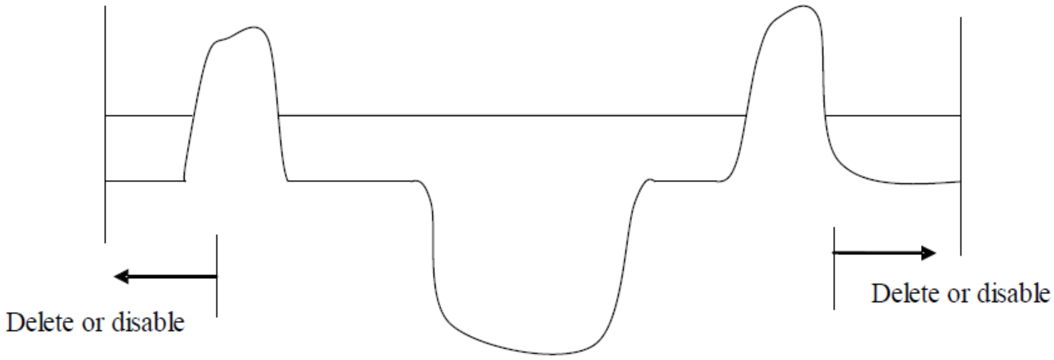


Figure IV－8 Edit Grain size distribution

V. Remarks

When river measurement data include points inside of levees, it is assumed that the water is also fill the levee-enclosed areas, as shown in the figure below. For this reason, it is necessary to remove or disable intra-levee measurement points at the pre-processing stage.



To Reader

- Please indicate the use of iRIC software, if you publish a paper with the results using iRIC software.
- The datasets provided at the Web site are sample data. Therefore you can use it for test computations.
- Let us know your suggestions, comments and concerns at **http://i-ric.org.**

iRIC Software
Nays2DH Solver Manual

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