

JICE助成金成果報告会2023

実測データを考慮した
山地と河川・沿岸域の広域土砂動態・
地形変化モデルの構築に関する研究

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1. 緒論：研究背景・目的

地球温暖化→河川氾濫の甚大化,
海岸侵食の加速化

ひとつの迅速な対策＝ダム・堰の「排砂」

課題点：①中流域の河床上昇抑制

②海岸線回復の質・量的予測

吉井・佐藤（2010）：天竜川

山野井・藤田（2014）：山地の土砂生産

JICE・総合土砂管理計画策定の手引き（2019）

cf. 流域流出：重枝ら（2010）

RRIモデル（土研）

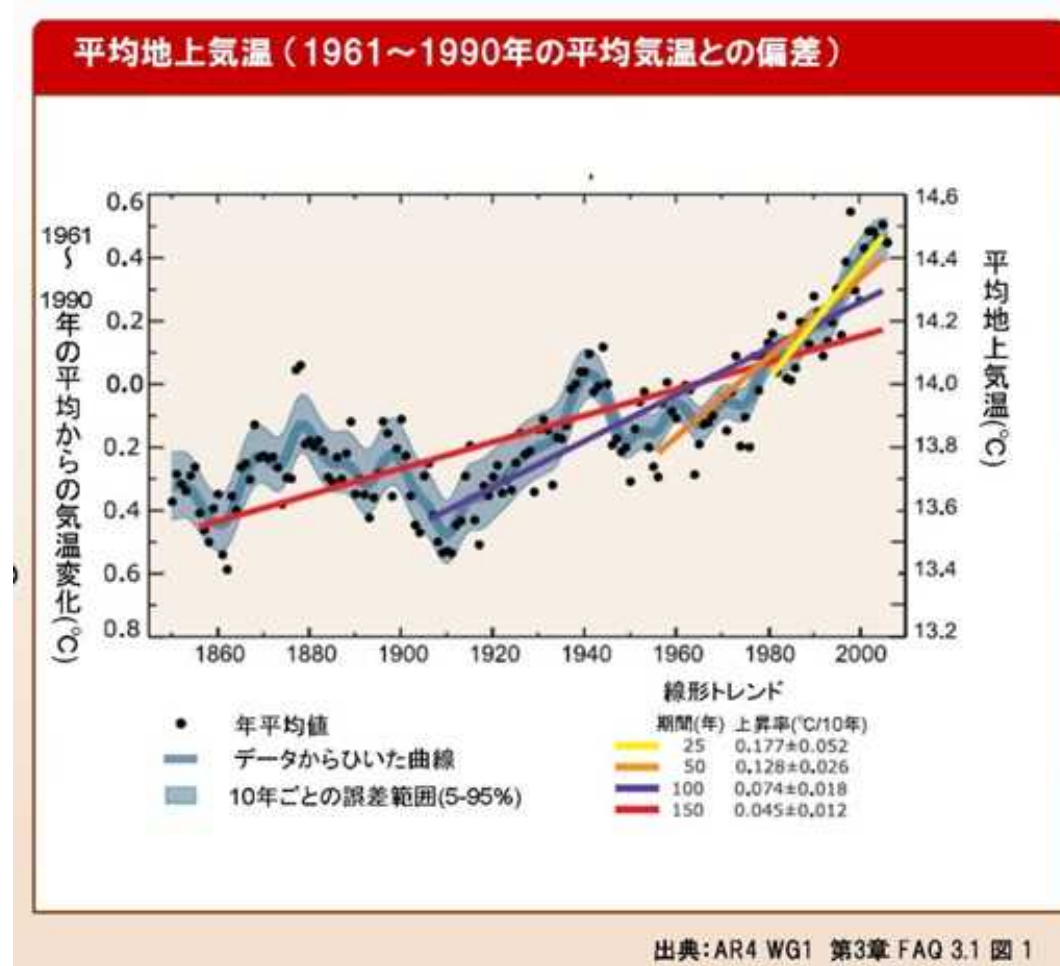
→画期的/計算土砂量の定量的精度・低

→河川・海岸の土砂量計測の難しさ

萬矢ら（2011），武川・二瓶（2013），

鵜崎ら（2021）：「実測データと簡易モデルによる河川の土砂量算定モデル：gRSM」

鵜崎ら（2009）：「潮汐と風波を考慮した泥と砂による広域土砂動態・地形変化モデルWDM-POM」



環境省，IPCC_AR4より引用

1. 研究背景・目的

「定量的精度を有する広域土砂動態モデル：SPR-WDMPOMを構築」

※SPR-WDMPOM = Sand Production and River sediment transport with WDM-POM

「山国川・中津干潟，利根川中上流域に現地適用」

➡全国展開！

2. 現地観測－山国川・中津干潟

2.1 観測概要

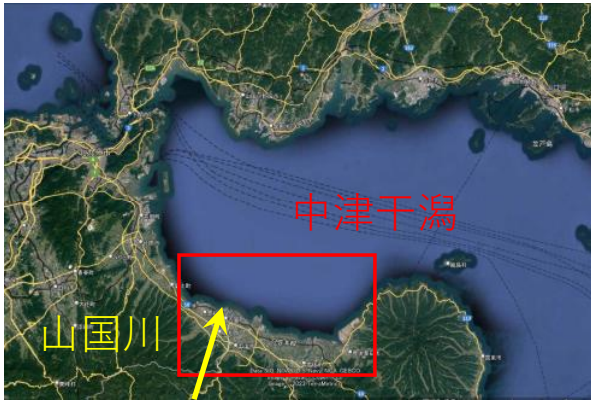


図-1 周防灘と中津干潟



図-2 観測領域と観測線



図-4 ADCPとRTK-GPS



図-3 採泥地点

■山国川と中津干潟の問題： 防災と環境の土砂問題の典型例

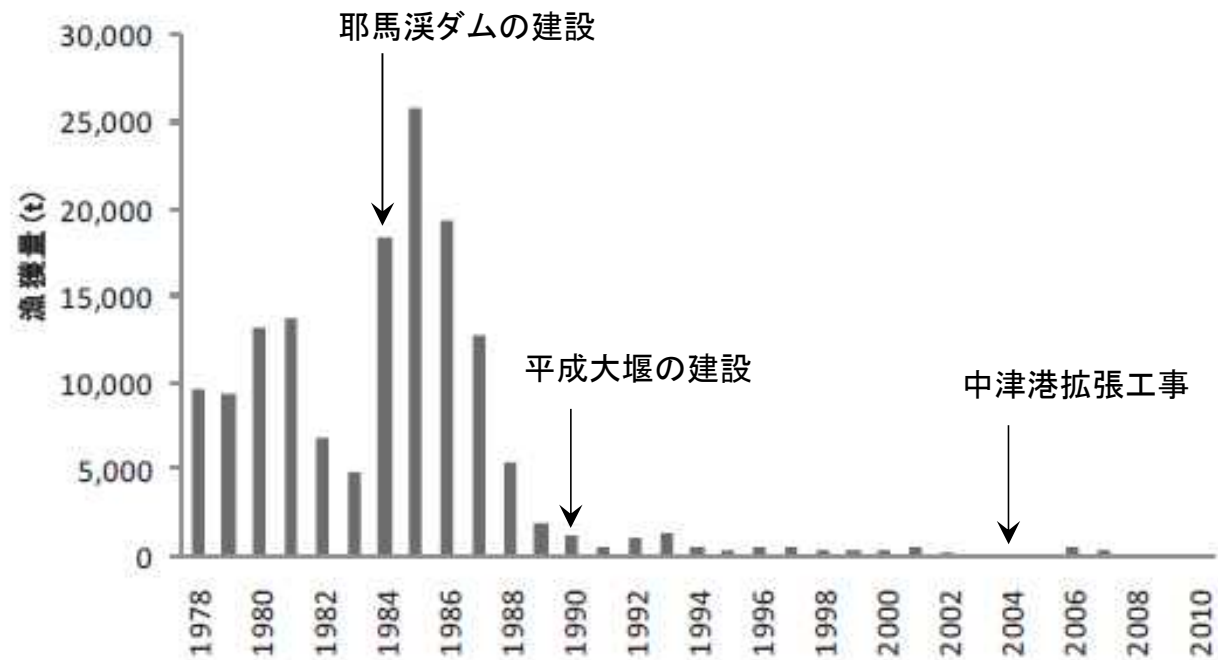


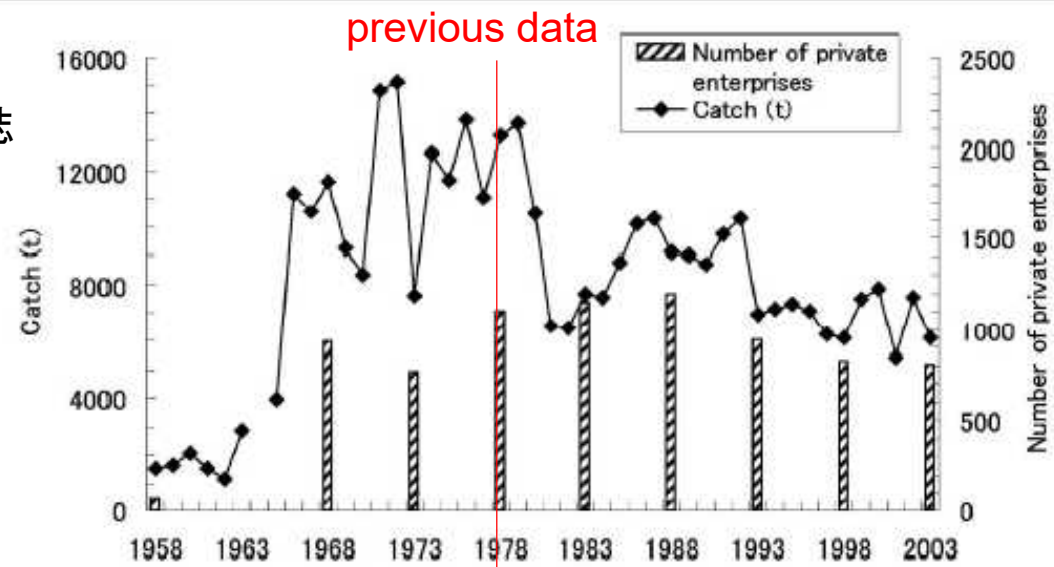
図1-1 中津市のアサリ漁獲量の推移

(1978～2009；大分県農林水産統計年報、2010；浅海チームによる推定)

アサリの減少原因＝餌(懸濁態・溶存態のN,P)＋水温・塩分濃度＋**底質：粗砂減少・堆泥**
＋乱獲＋食害

松川ら(2008)
日本水産学会誌

盤洲干潟・千葉
Banzu / Chiba



中津干潟・大分
Nakatsu / Oita

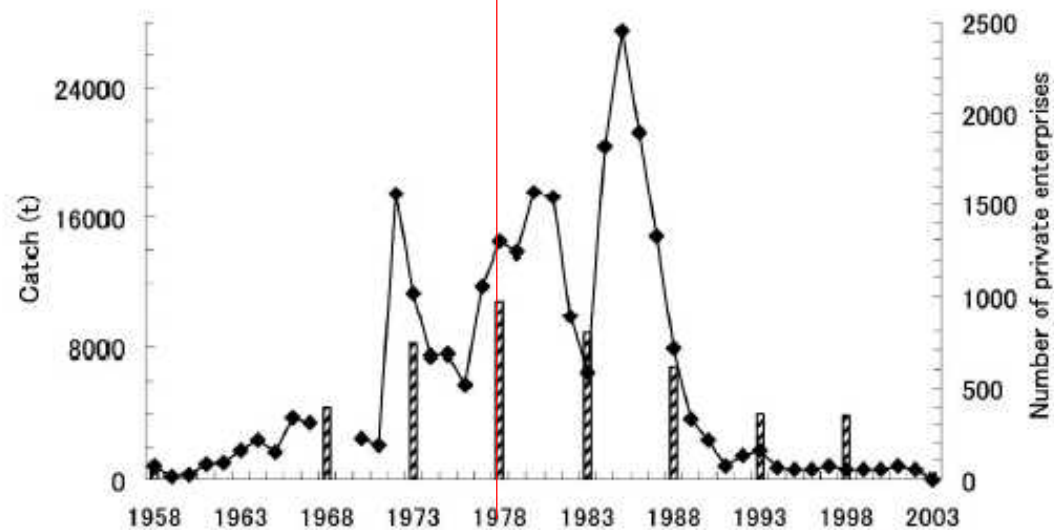
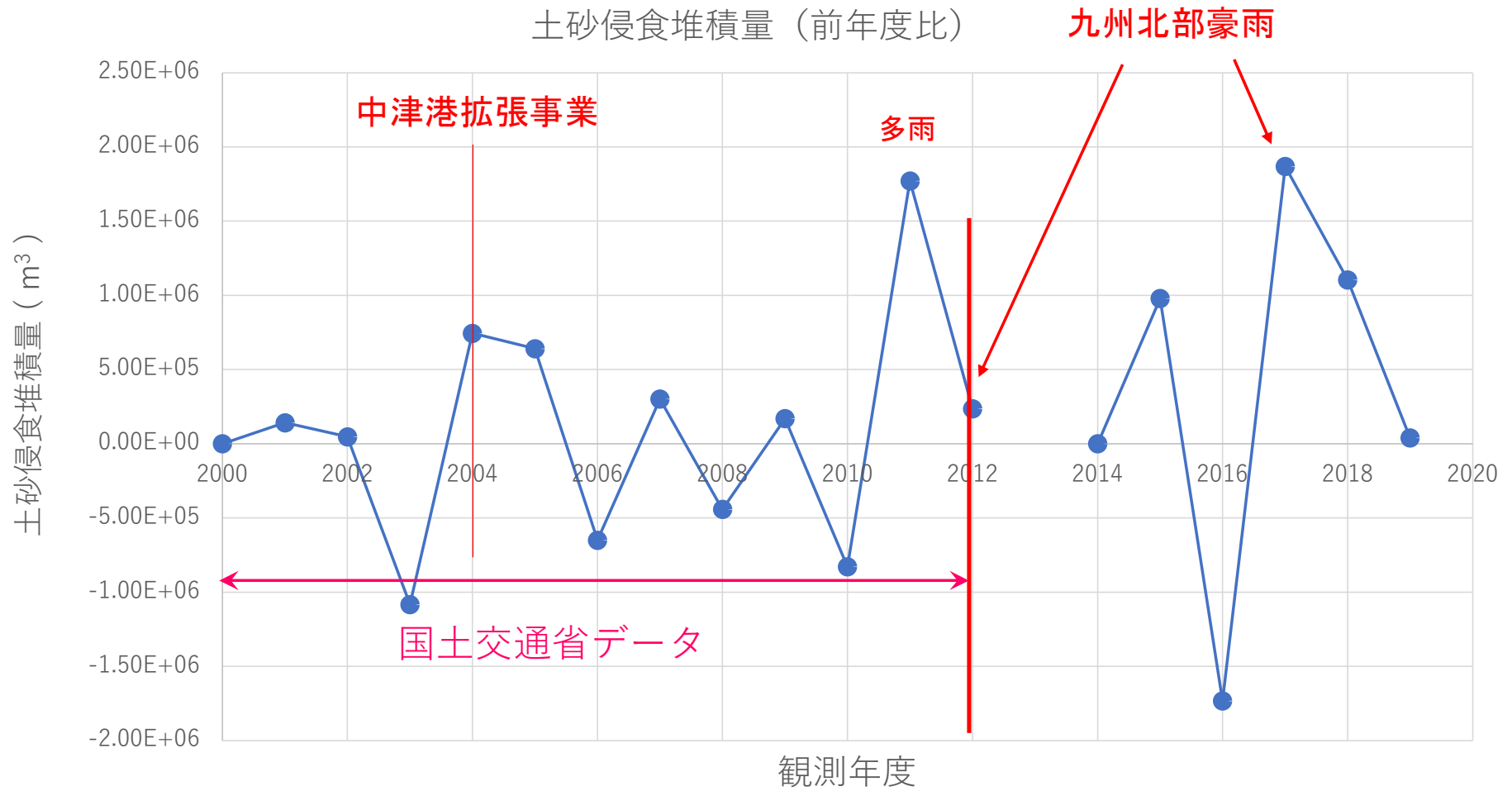


Fig. 1 Annual catches of Manila clam and the number of private enterprises harvesting the clam in Futtsu and Kisarazu areas, Chiba prefecture (top) and in Buzen area, Oita prefecture (bottom) from 1958 to 2003.

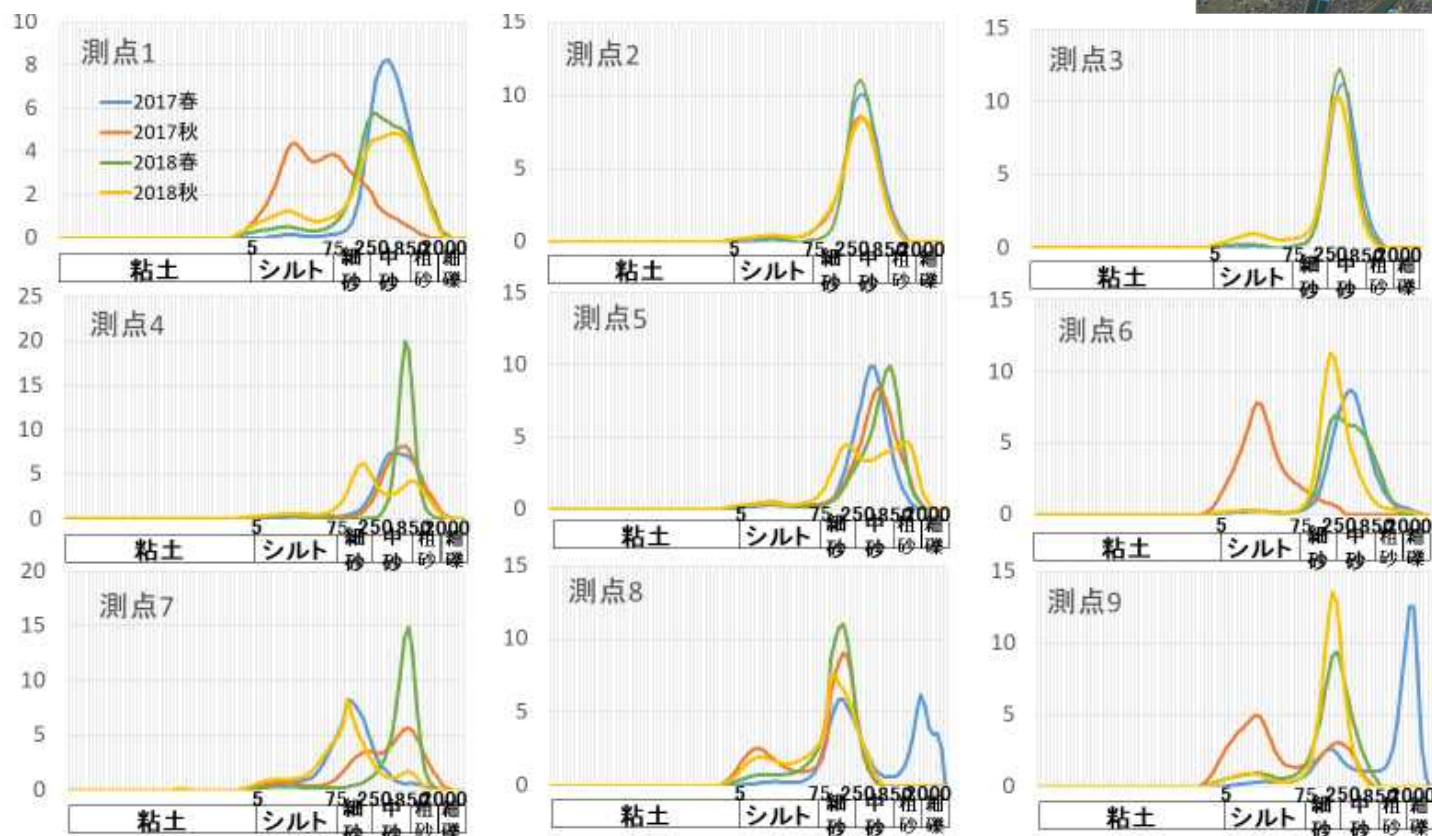
2.2 干潟の侵食・堆積



2.3 出水による河口—大新田地区沖の堆泥

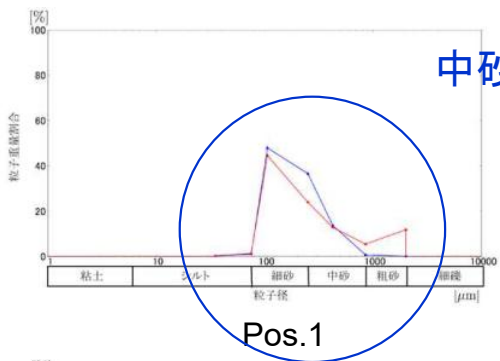


出現率 (%)



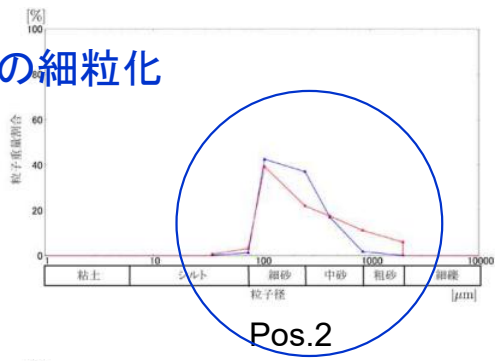
各測点の粒径分布 (表層採泥, 2017・18)

▲ 2022年春 ▲ 2022年秋

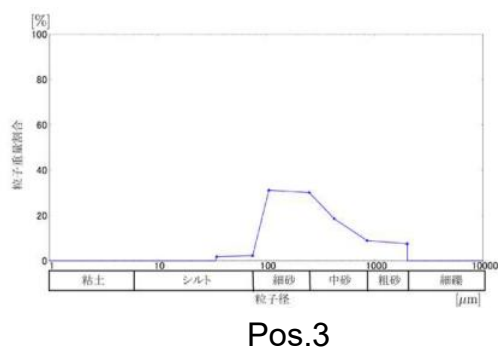


中砂への細粒化

Pos.1



Pos.2

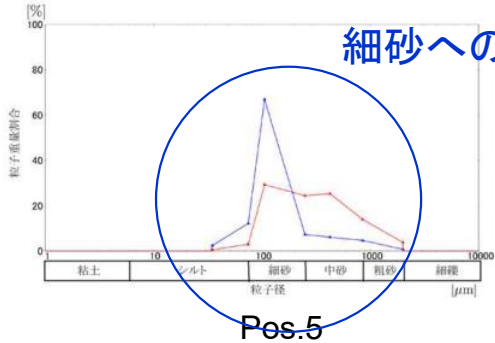


Pos.3



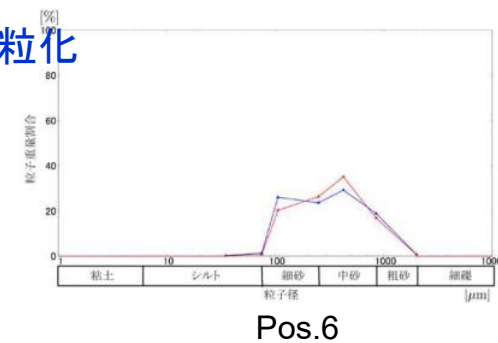
中砂への粗粒化

Pos.4

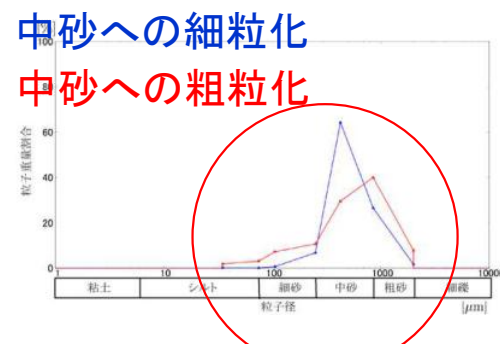


細砂への細粒化

Pos.5



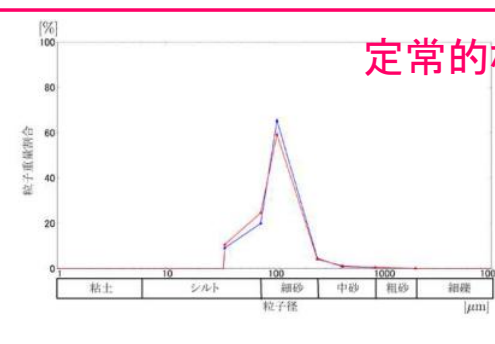
Pos.6



中砂への細粒化

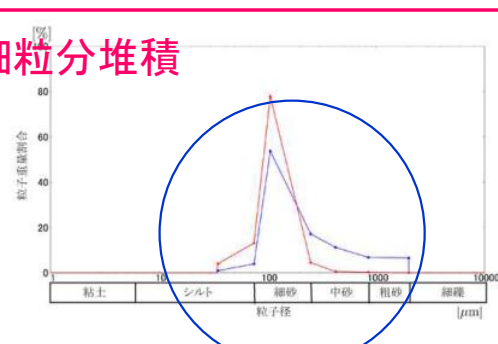
中砂への粗粒化

Pos.7



定常的な細粒分堆積

Pos.8



Pos.9

粒度分布変化(2022)

降水量と流量データ

		降水量の合計(mm)	最大流量(m ³ /s)	総流量(m ³)
2017	7月	484.5	943.5	2259.02
	8月	174.5		
	9月	284.5		
2018	7月	422.5	845	2301.85
	8月	9.5		
	9月	413		

		20mm/h以上の降水回数	20mm/h以上の降水量の合計(mm)	降水量の合計(mm)	20mm/h以上の降水量割合(%)	任意の1時間における降水量の最大(mm)
2017	7月	6	184	484.5	38.0	48.0
	8月	2	57	174.5	32.7	40.0
	9月	1	33	284.5	11.6	46.5
2018	7月	3	73	422.5	17.3	36.0
	8月	0	0	9.5	0	2.5
	9月	0	0	413	0	22.5

		降水量の合計		最大流量
2017年	6月	222.5	1166	2600.05
	7月	484.5		
	8月	174.5		
	9月	284.5		
2018年	6月	268	1113	2431.9
	7月	422.5		
	8月	9.5		
	9月	413		
2022年	6月	309.5	1218.5	2558.09
	7月	285.5		
	8月	159.5		
	9月	464		

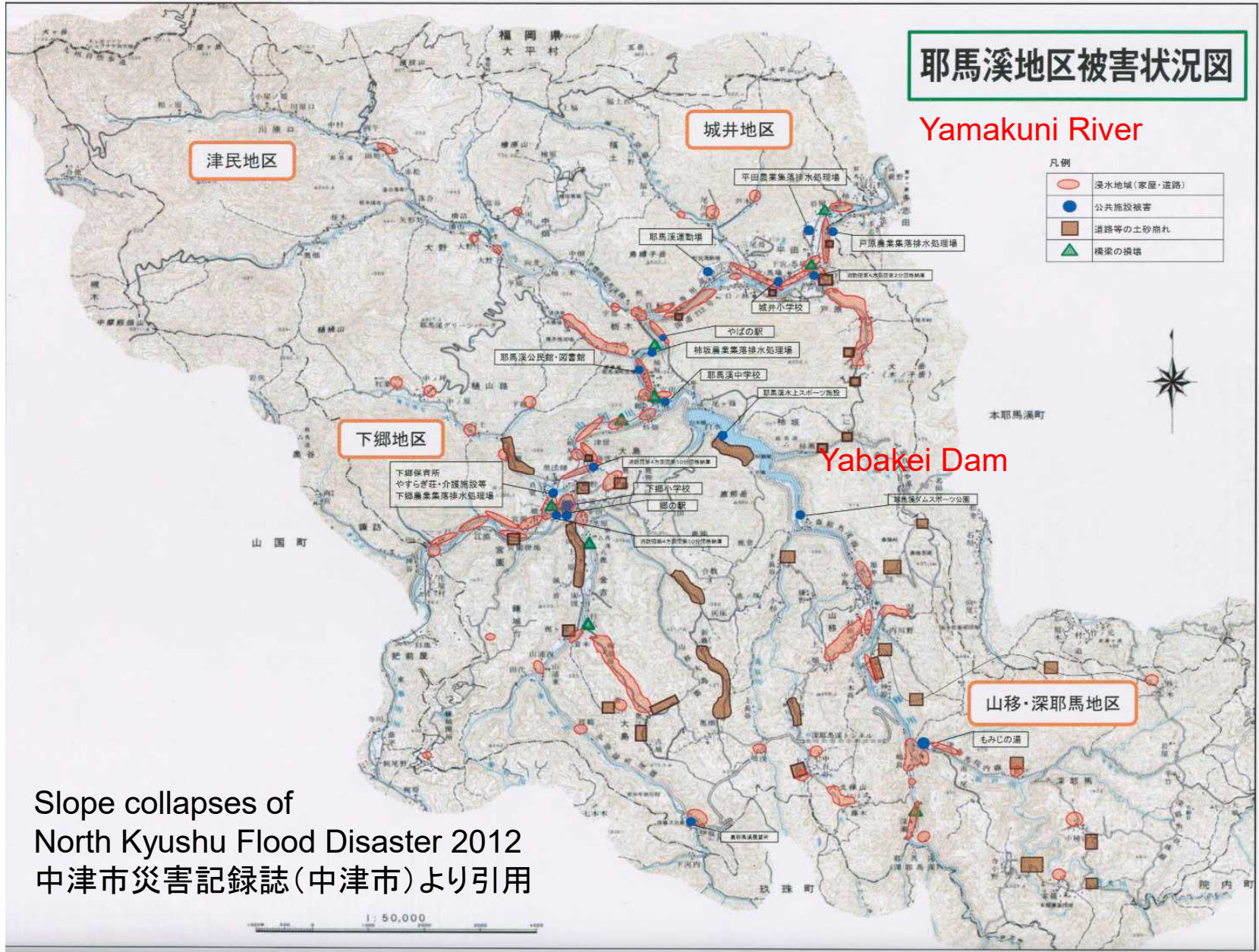
		20 mm/h 以上の 降水回数	20 mm/h 以上の 降水量の 合計(mm)	降水量の合計(mm)	20 mm/h 以上の 降水量割合(%)	任意の1時間 における降水量の 最大値(mm)
2017年	6月	2	51.5	222.5	23.1	30
	7月	6	184	484.5	38.0	48
	8月	2	57	174.5	32.7	40
	9月	1	33	284.5	11.6	46.5
2018年	6月	1	24	268	9.0	24
	7月	5	113	422.5	26.7	36
	8月	0	0	9.5	0	2.5
	9月	0	0	413	0	22.5
2022年	6月	0	0	309.5	0	19
	7月	4	138	285.5	48.3	70
	8月	1	36.5	159.5	22.9	36.5
	9月	9	293	464	63.1	58

耶馬溪地区被害状況図

Yamakuni River

凡例

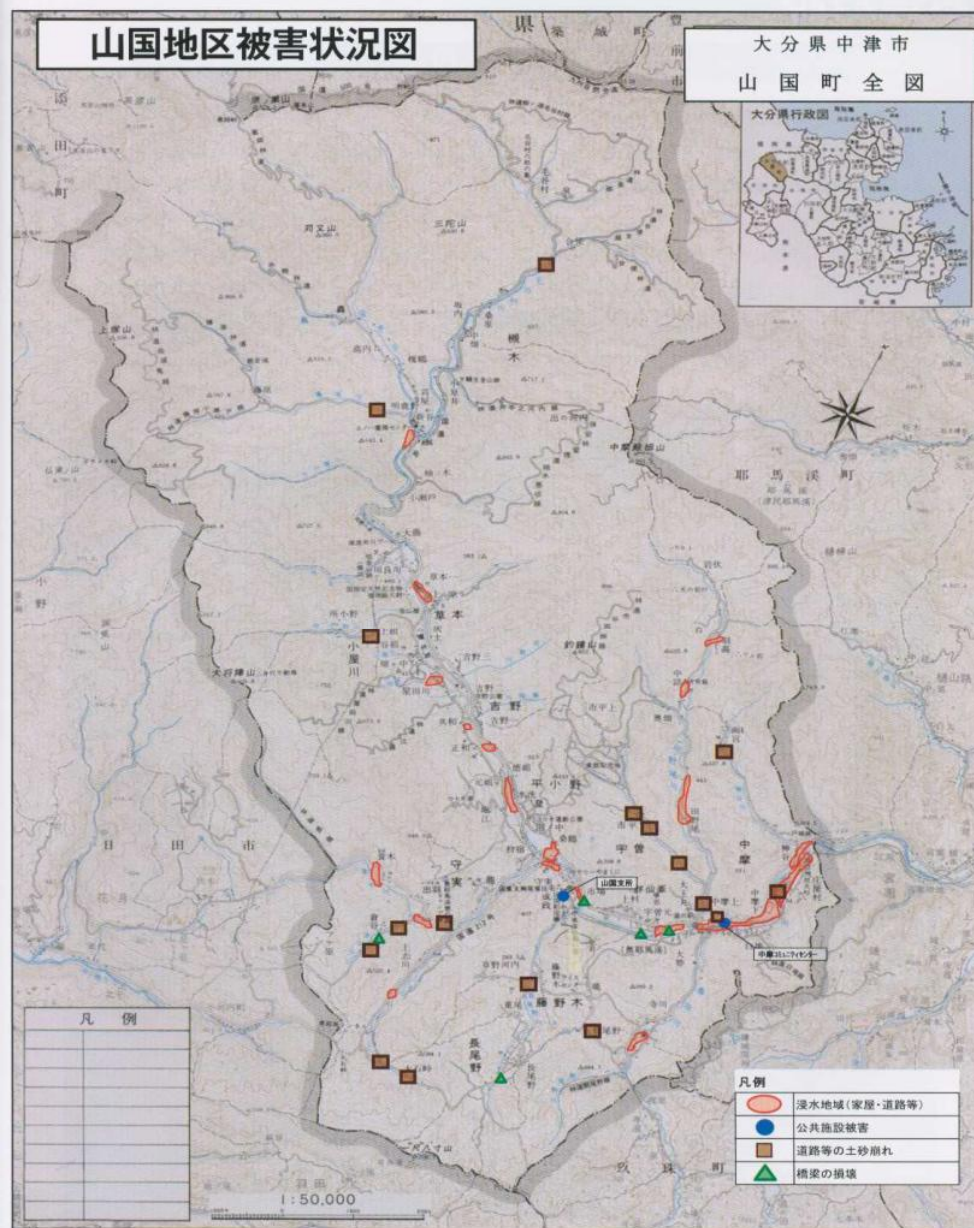
	浸水地域(家屋・道路)
	公共施設被害
	道路等の土砂崩れ
	橋梁の損壊



Slope collapses of
North Kyushu Flood Disaster 2012
中津市災害記録誌(中津市)より引用

山国地区被害状況図

大分県中津市 山国町全図



国土交通省
「山国川水系河川整備基本方針」
より引用

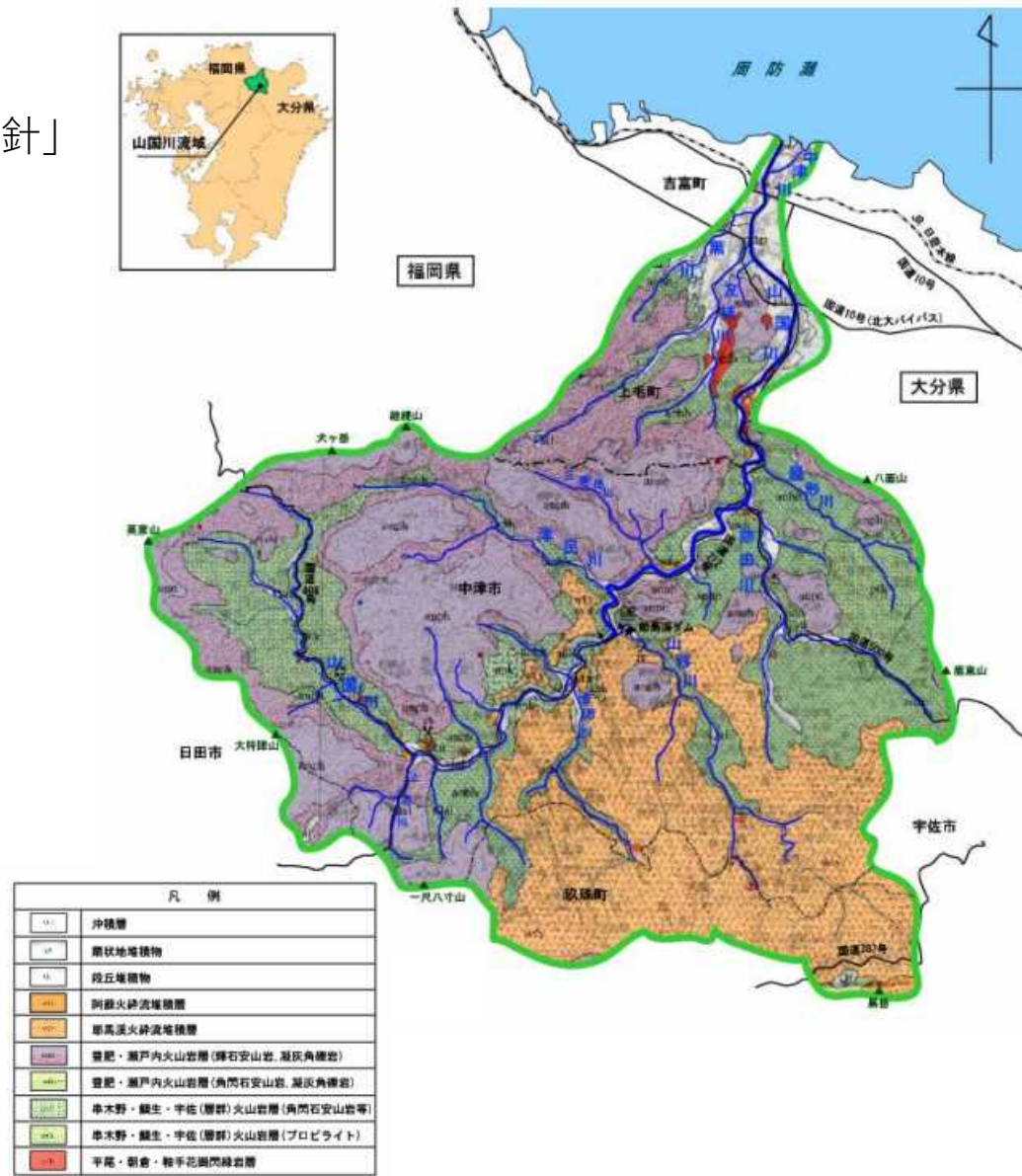
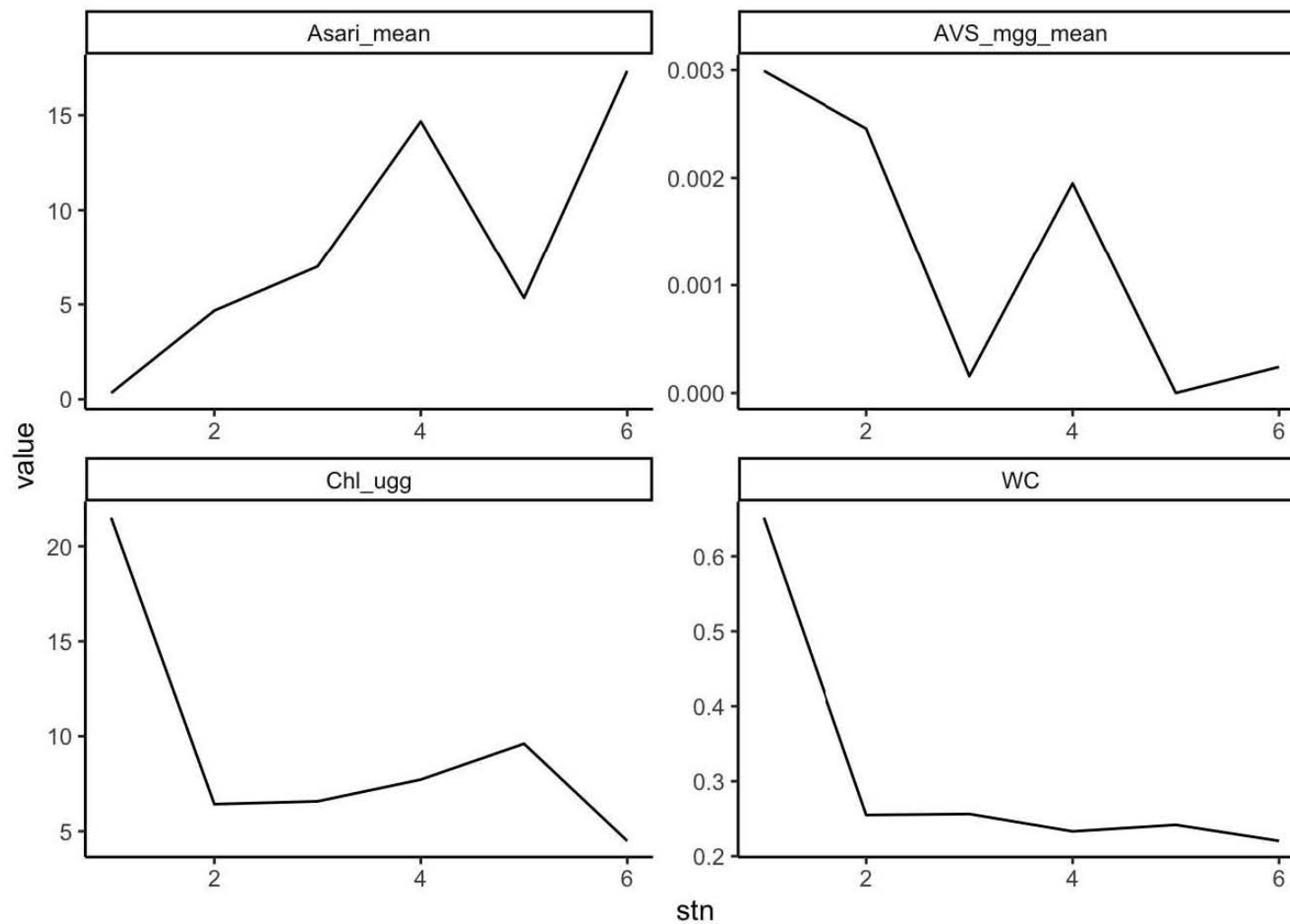


図 1-4 山国川流域地質図

2.4 定常的な汀線際の堆泥域

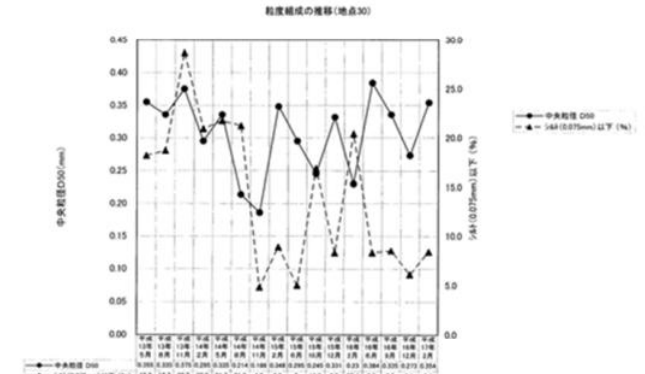
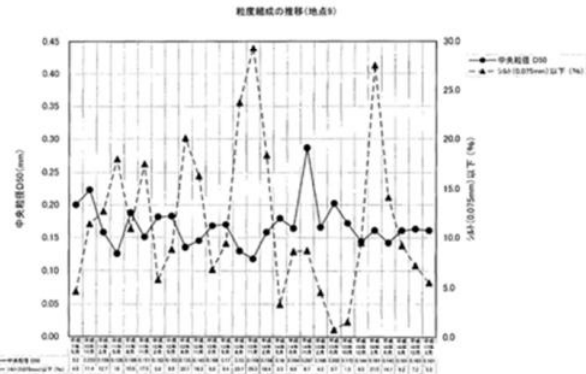
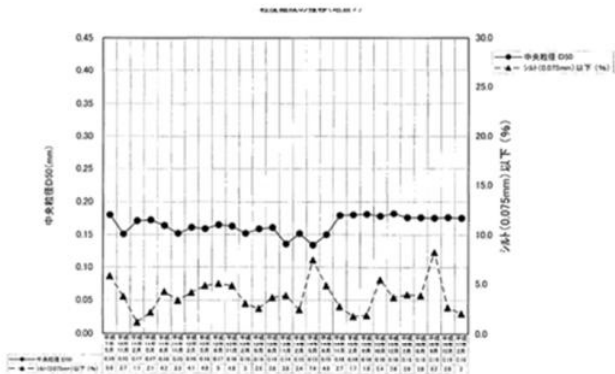
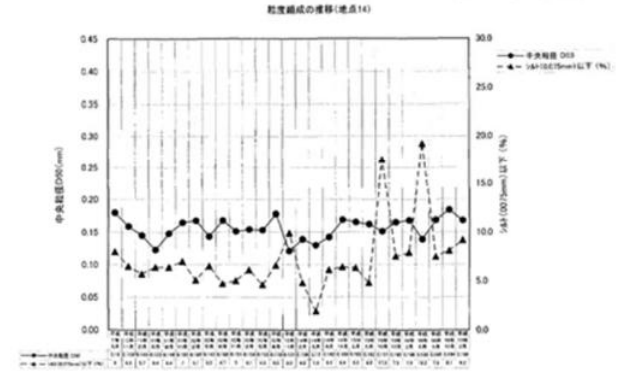
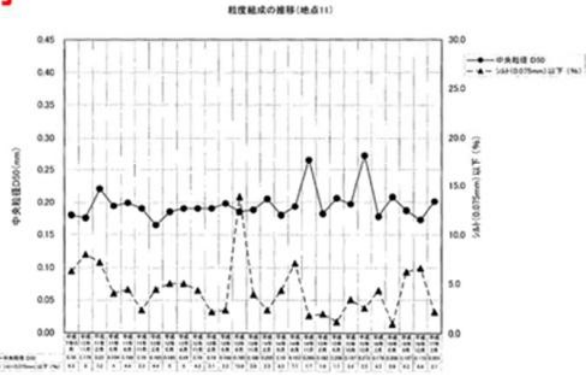
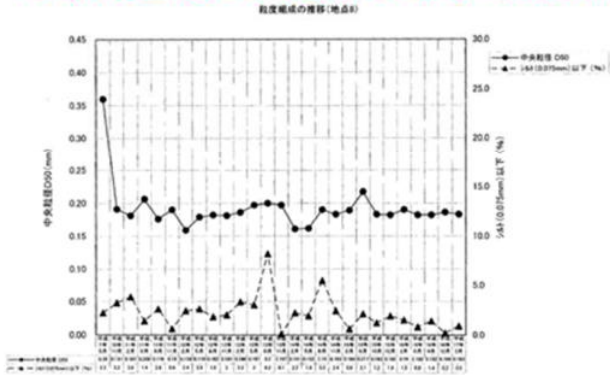


測線7付近における底質特性の岸沖分布(九州大学・熊本県立大学提供, 2021.6)

東側

出水による河口域・沖合堆泥 汀線近傍の定常的な堆泥傾向

中津港



中央

沖側

自見川河口

大新田地区の年間粒度分布変化(2003, 中津土木報告書) 岸側

3. 数値計算 – 山国川・中津干潟, gRSM+WDM-POM

3.1 計算概要

WDM-POM: Wet and Drying Morphodynamic POM

「潮汐と風波を考慮した泥と砂による広域土砂動態・海浜変形モデル」,
Uchiyama (2005), 鵜崎ら (2007, 2018)

■ 潮流+吹送流: POM=Princeton Ocean Model=NS方程式+連続式

■ 風波: Energy平衡方程式 → 海浜流: Radiation Stress Model

■ 泥の輸送: 移流拡散方程式, 砂の輸送: Bailard Model

■ 地形変化: 土砂収支式

■ 河川からの土砂供給Model

■ 干出・冠水Scheme

■ 極浅海域用の拡張対数則

gRSM：実測データと簡易モデルによる河川の土砂量算定モデル

□流量ハイドロ・グラフ



■Manning式 ※径深Rにて断面形状を考慮

■無次元掃流力 τ^* 算定

■ADCPによる鉛直流速分布による τ^* 算定

→計算値の補正

–対数分布則Fitting

■掃流砂量算定 – 芦田・道上式, 限界掃流力 – 岩垣式

↑ 妥当性検証・係数調整

■ADCP：Bottom Track Speedの応用による掃流砂量算定

– Rennie et al. (2002), 江頭の式

■浮遊砂輸送量に関するLQ式による浮遊砂量算定
– 採水試料のSS分析, 光学式濁度計

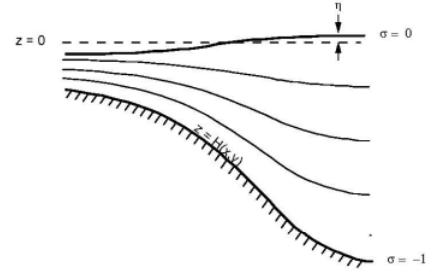


Figure 1. The sigma coordinate system.

$$\frac{\partial DU}{\partial x} + \frac{\partial DV}{\partial y} + \frac{\partial \omega}{\partial \sigma} + \frac{\partial \eta}{\partial t} = 0 \quad (2)$$

$$\begin{aligned} \frac{\partial UD}{\partial t} + \frac{\partial U^2 D}{\partial x} + \frac{\partial UV D}{\partial y} + \frac{\partial U \omega}{\partial \sigma} - fVD + gD \frac{\partial \eta}{\partial x} \\ + \frac{gD^2}{\rho_0} \int_{\sigma}^{\sigma'} \left[\frac{\partial \rho'}{\partial x} - \frac{\sigma'}{D} \frac{\partial D}{\partial x} \frac{\partial \rho'}{\partial \sigma'} \right] d\sigma' = \frac{\partial}{\partial \sigma} \left[\frac{K_M}{D} \frac{\partial U}{\partial \sigma} \right] + F_x \end{aligned} \quad (3)$$

$$\begin{aligned} \frac{\partial VD}{\partial t} + \frac{\partial UV D}{\partial x} + \frac{\partial V^2 D}{\partial y} + \frac{\partial V \omega}{\partial \sigma} + fUD + gD \frac{\partial \eta}{\partial y} \\ + \frac{gD^2}{\rho_0} \int_{\sigma}^{\sigma'} \left[\frac{\partial \rho'}{\partial y} - \frac{\sigma'}{D} \frac{\partial D}{\partial y} \frac{\partial \rho'}{\partial \sigma'} \right] d\sigma' = \frac{\partial}{\partial \sigma} \left[\frac{K_M}{D} \frac{\partial V}{\partial \sigma} \right] + F_y \end{aligned} \quad (4)$$

$$\frac{\partial TD}{\partial t} + \frac{\partial TUD}{\partial x} + \frac{\partial TVD}{\partial y} + \frac{\partial T \omega}{\partial \sigma} = \frac{\partial}{\partial \sigma} \left[\frac{K_H}{D} \frac{\partial T}{\partial \sigma} \right] + F_T - \frac{\partial R}{\partial z} \quad (5)$$

$$\frac{\partial SD}{\partial t} + \frac{\partial SUD}{\partial x} + \frac{\partial SV D}{\partial y} + \frac{\partial S \omega}{\partial \sigma} = \frac{\partial}{\partial \sigma} \left[\frac{K_H}{D} \frac{\partial S}{\partial \sigma} \right] + F_S \quad (6)$$

$$\frac{\partial q^2 D}{\partial t} + \frac{\partial U q^2 D}{\partial x} + \frac{\partial V q^2 D}{\partial y} + \frac{\partial \omega q^2}{\partial \sigma} = \frac{\partial}{\partial \sigma} \left[\frac{K_q}{D} \frac{\partial q^2}{\partial \sigma} \right]$$

$$+ \frac{2K_M}{D} \left[\left(\frac{\partial U}{\partial \sigma} \right)^2 + \left(\frac{\partial V}{\partial \sigma} \right)^2 \right] + \frac{2g}{\rho_o} K_H \frac{\partial \bar{p}}{\partial \sigma} - \frac{2Dq^2}{B_1 \ell} + F_q \quad (7)$$

$$\frac{\partial q^2 \ell D}{\partial t} + \frac{\partial U q^2 \ell D}{\partial x} + \frac{\partial V q^2 \ell D}{\partial y} + \frac{\partial \omega q^2 \ell}{\partial \sigma} = \frac{\partial}{\partial \sigma} \left[\frac{K_q}{D} \frac{\partial q^2 \ell}{\partial \sigma} \right] \\ + E_1 \ell \left(\frac{K_M}{D} \left[\left(\frac{\partial U}{\partial \sigma} \right)^2 + \left(\frac{\partial V}{\partial \sigma} \right)^2 \right] + E_3 \frac{g}{\rho_o} K_H \frac{\partial \bar{p}}{\partial \sigma} \right) - \frac{Dq^2}{B_1} \tilde{W} + F_t \quad (8)$$

Definitions of the variables are contained in section 3. Note that ω is the transformed vertical velocity; physically, ω is the velocity component normal to sigma surfaces. The transformation to the Cartesian vertical velocity is

$$W = \omega + U \left(\sigma \frac{\partial D}{\partial x} + \frac{\partial \eta}{\partial x} \right) + V \left(\sigma \frac{\partial D}{\partial y} + \frac{\partial \eta}{\partial y} \right) + \sigma \frac{\partial D}{\partial t} + \frac{\partial \eta}{\partial t}$$

The so-called wall proximity function is prescribed according to $\tilde{W} = 1 + E_2(\ell / kL)$ where $L^{-1} = (\eta - z)^{-1} + (H - z)^{-1}$. Also, $\partial \bar{p} / \partial \sigma \equiv \partial \rho / \partial \sigma - c_s^{-2} \partial p / \partial \sigma$ (see discussion of static stability in Appendix A) where c_s is the speed of sound. Note that T is potential temperature (see Appendix A).

In equations (3) and (4), ρ_{MEAN} should be subtracted from ρ to form ρ' before the integration is carried out in subroutine BAROPG. ρ_{MEAN} is generally the initial density field which is area averaged on z -levels and then transferred to sigma coordinates in the exact same way as the initial density field. This procedure should reduce the truncation errors associated with the calculation of the pressure gradient term in sigma coordinate over steep topography (see Mellor et al., 1994 and Mellor et al. 1998 for evaluation of this error in POM).

The horizontal viscosity and diffusion terms are defined according to:

$$F_x \equiv \frac{\partial}{\partial x} (H\tau_{xx}) + \frac{\partial}{\partial y} (H\tau_{xy}) \quad (9a)$$

$$F_y \equiv \frac{\partial}{\partial x} (H\tau_{xy}) + \frac{\partial}{\partial y} (H\tau_{yy}) \quad (9b)$$

where

$$\tau_{xx} = 2A_M \frac{\partial U}{\partial x}, \quad \tau_{xy} = \tau_{yx} = A_M \left(\frac{\partial U}{\partial y} + \frac{\partial V}{\partial x} \right), \quad \tau_{yy} = 2A_M \frac{\partial V}{\partial y} \quad (10a,b,c)$$

Also,

$$F_\phi \equiv \frac{\partial}{\partial x} (Hq_x) + \frac{\partial}{\partial y} (Hq_y) \quad (11)$$

where

$$q_x \equiv A_H \frac{\partial \phi}{\partial x}, \quad q_y \equiv A_H \frac{\partial \phi}{\partial y} \quad (12a,b)$$

and where ϕ , represents T , S , q^2 or $q^2 \ell$. It should be noted that these horizontal diffusion terms are not what one would obtain by transforming the conventional forms to the sigma coordinate system. Justification for the present forms will be found in Mellor and Blumberg (1985) and relate to the fact that we wish to maintain a valid bottom boundary layer simulation in the face of horizontal diffusion which may be large. The penalty for this is that (12a,b) in sigma coordinates can introduce vertical fluxes even when isotherms and isohalines are flat in cartesian coordinates. The remedy for this is, first, the use of a Smagorinsky diffusivity (see below) so that, at least when velocities are small or nil, so are the values of q_x and q_y . The second remedy is that, before executing (12a, b) for temperature or salinity, we first subtract T_{CLIM} and S_{CLIM} which are "climatologies" of T and S . The latter may be true climatologies (e.g.; Levitus) or approximations such as temperature and salinities which are area averaged prior to transfer to sigma coordinates (in which case, they are treated the same as ρ_{MEAN}). If something like a Levitus climatology is used, then most of the vertical component of the diffusion is removed; furthermore, the diffusion terms tend to slowly drive the scalars back to climatology rather than to a horizontally homogeneous state as in the case of z - level models. The third remedy is make use of a small diffusivity relative to viscosity. Thus, the value, $TPRNI \equiv A_H / A_M$, can generally set to a small number, say 0.2, or even zero in some cases.

It should be noted that the treatment in (9a,b), (10a,b), (11) and (12a,b) allows for a realistic treatment of bottom boundary layers. The bottom boundary layer is important in tidally driven regions, in wind driven coastal regions and according to Mellor and Wang (1996), in deep ocean basins.

In (9a, b) and (11), H is used in place of D for the small algorithmic simplification it offers for terms whose physical significance is questionable.

The Smagorinsky Diffusivity

We generally use the Smagorinsky diffusivity for horizontal diffusion although a constant or biharmonic diffusion can and has been used instead. The Smagorinsky formula is,

$$A_M = C\Delta x\Delta y \frac{1}{2} \left| \nabla \mathbf{V} + (\nabla \mathbf{V})^T \right|$$

where $\left| \nabla \mathbf{V} + (\nabla \mathbf{V})^T \right| = [(\partial u / \partial x)^2 + (\partial v / \partial x + \partial u / \partial y)^2 / 2 + (\partial v / \partial y)^2]^{1/2}$. Values of C (the HORCON parameter) in the range, 0.10 to 0.20 seem to work well, but, if the grid spacing is small enough (Oey *et al.*, 1985a,b), C can be nil. An advantage of the Smagorinsky relation is that C is non-dimensional; related advantages are that A_M decreases as resolution improves and that A_M is small if velocity gradients are small.

Vertical Boundary Conditions.

The vertical boundary conditions for (2) are

$$\omega(0) = \omega(-1) = 0 \quad (13a,b)$$

However, if there is to be surface throughflow of (usually fresh) water, $\omega(0) \neq 0$.

The surface boundary conditions for (3) and (4) are

$$\frac{K_M}{D} \left(\frac{\partial U}{\partial \sigma}, \frac{\partial V}{\partial \sigma} \right) = - \langle wu(0) \rangle, \langle wv(0) \rangle, \sigma \rightarrow 0 \quad (14a,b)$$

where the right hand side of (14a,b) is the input values of the surface turbulence momentum flux (the stress components are opposite in sign). The bottom boundary conditions are

$$\frac{K_M}{D} \left(\frac{\partial U}{\partial \sigma}, \frac{\partial V}{\partial \sigma} \right) = C_z [U^2 + V^2]^{1/2} (U, V), \sigma \rightarrow -1 \quad (14c,d)$$

where

$$C_z = \text{MAX} \left[\frac{\kappa^2}{\left[\ln \left\{ (1 + \sigma_{sb-1}) H / z_0 \right\} \right]^2}, 0.0025 \right] \quad (14e)$$

$\kappa = 0.4$ is the von Karman constant and z_0 is the roughness parameter. Equations (14c,d,e) can be derived by matching the numerical solution to the "law of the wall". Numerically, they are applied to the first grid points nearest the bottom. Where the

bottom is not well resolved, $(1+\sigma_{kb-1})H/z_0$ is large and (14e) reverts to a constant drag coefficient, 0.0025. The boundary conditions on (5) and (6) are

$$\frac{K_H}{D} \left(\frac{\partial T}{\partial \sigma}, \frac{\partial S}{\partial \sigma} \right) = - \langle w\theta(0) \rangle, \quad \sigma \rightarrow 0 \quad (15a,b)$$

$$\frac{K_H}{D} \left(\frac{\partial T}{\partial \sigma}, \frac{\partial S}{\partial \sigma} \right) = 0, \quad \sigma \rightarrow -1 \quad (15c,d)$$

The boundary conditions for (7) and (8) are

$$(q^2(0), q^2 \ell(0)) = (B_1^{2/3} u_*^2(0), 0) \quad (16a,b)$$

$$(q^2(-1), q^2 \ell(-1)) = (B_1^{2/3} u_*^2(-1), 0) \quad (16c,d)$$

where B_1 is one of the turbulence closure constants and u_* is the friction velocity at the top or bottom as denoted in (16a) and (16c). In pom97.f and later versions, (16a) has been replaced by $q^2 \ell(\sigma_1) = q^2(\sigma_1) \kappa D \sigma_1$ where σ_1 is the value of σ corresponding to $k=1$, it is believed that this averts some numerical noise in some applications.

The Vertically Integrated Equations

The equations, governing the dynamics of coastal circulation, contain fast moving external gravity waves and slow moving internal gravity waves. It is desirable in terms of computer economy to separate the vertically integrated equations (external mode) from the vertical structure equations (internal mode). This technique, known as mode splitting (Simons, 1974; Madala and Piacsek, 1977) permits the calculation of the free surface elevation with little sacrifice in computational time by solving the velocity transport separately from the three-dimensional calculation of the velocity and the thermodynamic properties.

The velocity external mode equations are obtained by integrating the internal mode equations over the depth, thereby eliminating all vertical structure. Thus, by integrating Equation (2) from $\sigma = -1$ to $\sigma = 0$ and using the boundary conditions (13a,b), an equation for the surface elevation can be written as

$$\frac{\partial \eta}{\partial t} + \frac{\partial \bar{U} D}{\partial x} + \frac{\partial \bar{V} D}{\partial y} = 0 \quad (17)$$

After integration, the momentum equations, (3) and (4), become

$$\begin{aligned} \frac{\partial \bar{U}D}{\partial t} + \frac{\partial \bar{U}^2D}{\partial x} + \frac{\partial \bar{U}\bar{V}D}{\partial y} - \tilde{F}_x - f\bar{V}D + gD \frac{\partial \eta}{\partial x} = -\langle wu(0) \rangle + \langle wu(-1) \rangle \\ + G_x - \frac{gD}{\rho_o} \int_{-1}^0 \int_{\sigma}^{\sigma'} \left[D \frac{\partial \rho'}{\partial x} - \frac{\partial D}{\partial x} \sigma' \frac{\partial \rho'}{\partial \sigma} \right] d\sigma' d\sigma \end{aligned} \quad (18)$$

$$\begin{aligned} \frac{\partial \bar{V}D}{\partial t} + \frac{\partial \bar{U}\bar{V}D}{\partial x} + \frac{\partial \bar{V}^2D}{\partial y} - \tilde{F}_y + f\bar{U}D + gD \frac{\partial \eta}{\partial y} = -\langle wv(0) \rangle + \langle wv(-1) \rangle \\ + G_y - \frac{gD}{\rho_o} \int_{-1}^0 \int_{\sigma}^{\sigma'} \left[D \frac{\partial \rho'}{\partial y} - \frac{\partial D}{\partial y} \sigma' \frac{\partial \rho'}{\partial \sigma} \right] d\sigma' d\sigma \end{aligned} \quad (19)$$

The overbars denote vertically integrated velocities such as

$$\bar{U} \equiv \int_{-1}^0 U d\sigma. \quad (20)$$

The wind stress components are $-\langle wu(0) \rangle$ and $-\langle wv(0) \rangle$, and the bottom stress components are $-\langle wu(-1) \rangle$ and $-\langle wv(-1) \rangle$. The quantities \tilde{F}_x and \tilde{F}_y are defined according to

$$\tilde{F}_x = \frac{\partial}{\partial x} \left[H2\bar{A}_M \frac{\partial \bar{U}}{\partial x} \right] + \frac{\partial}{\partial y} \left[H\bar{A}_M \left(\frac{\partial \bar{U}}{\partial y} + \frac{\partial \bar{V}}{\partial x} \right) \right] \quad (21a)$$

and

$$\tilde{F}_y = \frac{\partial}{\partial y} \left[H2\bar{A}_M \frac{\partial \bar{V}}{\partial y} \right] + \frac{\partial}{\partial x} \left[H\bar{A}_M \left(\frac{\partial \bar{U}}{\partial y} + \frac{\partial \bar{V}}{\partial x} \right) \right] \quad (21b)$$

The so-called dispersion terms are defined according to

$$G_x = \frac{\partial \bar{U}^2D}{\partial x} + \frac{\partial \bar{U}\bar{V}D}{\partial y} - \tilde{F}_x - \frac{\partial \bar{U}^2D}{\partial x} - \frac{\partial \bar{U}\bar{V}D}{\partial y} + \bar{F}_x \quad (22a)$$

$$G_y = \frac{\partial \bar{U}\bar{V}D}{\partial x} + \frac{\partial \bar{V}^2D}{\partial y} - \tilde{F}_y - \frac{\partial \bar{U}\bar{V}D}{\partial x} - \frac{\partial \bar{V}^2D}{\partial y} + \bar{F}_y \quad (22b)$$

Note that, if A_M is constant in the vertical, then the "F" terms in (22a) and (22b) cancel. However, we account for possible vertical variability in the horizontal diffusivity; such is the case when a Smagorinsky type diffusivity is used. As detailed below, all of the terms on the right side of (18) and (19) are evaluated at each internal time step and then held constant throughout the many external time steps. If the external mode is executed *cum sole*, then $G_x = G_y = 0$.

$$\vec{q}_c = A_c(\tau_m - \tau_c)\vec{u}_c/\rho g \quad (2.2.1)$$

$$\vec{q}_w = A_w F_D(\tau_m - \tau_c)\vec{u}_b/\rho g \quad (2.2.2)$$

$$F_D = 1 - [1 + \tanh\{\alpha(\psi_m - \psi_{c1})\}] \\ \times [1 + \tanh\{\alpha(\psi_m - \psi_{c2})\}]/2 \quad (2.2.6)$$

渡辺モデル

$$\begin{aligned}
 i &= \rho s' g (1 - \lambda) q \\
 &= \rho C_f \frac{\epsilon_B}{\tan \phi} \left[|u(t)|^2 - \frac{\tan \beta}{\tan \phi} |u(t)|^3 \right] \\
 &\quad + \rho C_f \frac{\epsilon_s}{w_0} \left[|u(t)|^3 - \frac{\epsilon_s}{w_0} \tan \beta |u(t)|^5 \right]
 \end{aligned}
 \tag{2.2.25}$$

Bailardモデル

$$\frac{\partial(v_x S)}{\partial x} + \frac{\partial(v_y S)}{\partial y} + \frac{\partial(v_\theta S)}{\partial \theta} = -\varepsilon_b S \quad (5.4.15)$$

ここで、 S は波の方向スペクトル、 (x, y) は水平座標、 θ は x 軸から反時計回りに測った波向角、 ε_b はエネルギー減衰係数であり、特性速度 (v_x, v_y, v_θ) は

$$(v_x, v_y, v_\theta) = \left\{ \begin{array}{l} C_g \cos \theta, C_g \sin \theta, \\ \frac{C_g}{C} \left(\sin \theta \frac{\partial C}{\partial x} - \cos \theta \frac{\partial C}{\partial y} \right) \end{array} \right\} \quad (5.4.16)$$

＋回折項

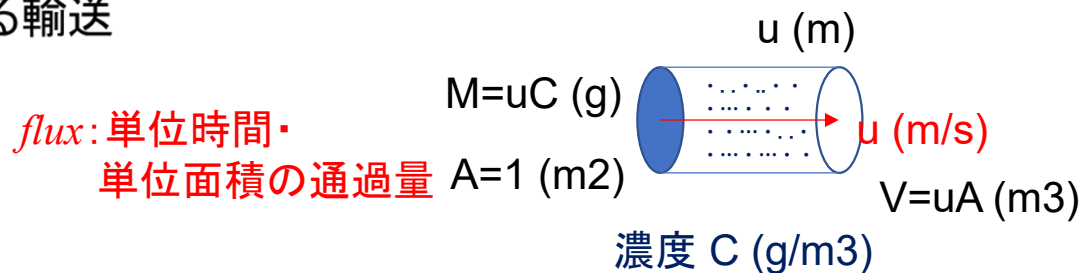
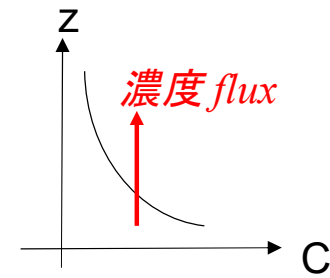
移流拡散方程式 = 濃度fluxの収支式

$$\frac{\partial \bar{C}}{\partial t} + u \frac{\partial C}{\partial x} + v \frac{\partial C}{\partial y} + w \frac{\partial C}{\partial z} \rightarrow \frac{\partial(uC)}{\partial x}$$

$$= \frac{\partial}{\partial x} \left((D_x + K_x) \frac{\partial C}{\partial x} \right) + \frac{\partial}{\partial y} \left((D_y + K_y) \frac{\partial C}{\partial y} \right) + \frac{\partial}{\partial z} \left((D_z + K_z) \frac{\partial C}{\partial z} \right)$$

D: 分子拡散係数, K: 乱流拡散係数
 $w = w - wg$: 沈降速度

- ・(分子) 拡散現象 = 濃度勾配による変化
- ・乱流拡散 = 流れ場の乱れによる濃度変化
- ・移流 = 平均流速による輸送



泥の沈降速度モデル: Burban *et al.*(1990)の実験式

$$W_s = \alpha (CG)^\beta \quad \sim \text{凝集性土砂 cohesive sediment}$$

$$G = \rho K_M \left[\left(\frac{\partial U}{\partial z} \right)^2 + \left(\frac{\partial V}{\partial z} \right)^2 \right]^{1/2}$$

砂の沈降速度式: Rubeyの実験式(1933) cf. Stokesの法則

$$\frac{w_f}{\sqrt{sgd}} = \sqrt{\frac{2}{3} + \frac{36v^2}{sgd^3}} - \sqrt{\frac{36v^2}{sgd^3}} \quad (4.6)$$

ただし, v : 水の動粘性係数 ($\div 0.01 \text{cm}^2/\text{s}$)

s : 砂粒の水中比重 (砂の標準値 1.65)

d : 砂粒の粒径 (cm)

g : 重力加速度 ($980 \text{cm}/\text{s}^2$)

Winterwerp,2002

$$w_{s,r} = \frac{(\rho_s - \rho_w)gD^2}{18\mu} \quad (5)$$

$$w_{s,r} = \frac{\alpha}{18\beta} \frac{(\rho_s - \rho_w)g}{\mu} D_p^{3-n_f} \frac{D^{n_f-1}}{1 + 0.15Re_p^{0.687}} \quad (4)$$

$$w_s = w_{s,r} \frac{(1 - \phi_*)(1 - \phi_p)}{1 + 2.5\phi} \quad (6)$$

$$\phi = \left(\frac{\rho_s - \rho_w}{\rho_f - \rho_w} \right) \frac{c}{\rho_s} = \frac{c}{\rho_s} \left[\frac{D}{D_p} \right]^{3-n_f} = f_s N D^3$$

$w_{s,r}$ はStokes式から求められる粒子単体での沈降速度, $\phi = (\rho_s - \rho_w) / (\rho_f - \rho_w) c / \rho_s$, $\phi_* = \min(1, \phi)$, $\phi_p = c / \rho_s$: いずれも体積濃度, ρ_s, ρ_w, ρ_f : 粒子密度, 水の密度, フロックの密度, μ : 分子粘性係数, D : フロック径.

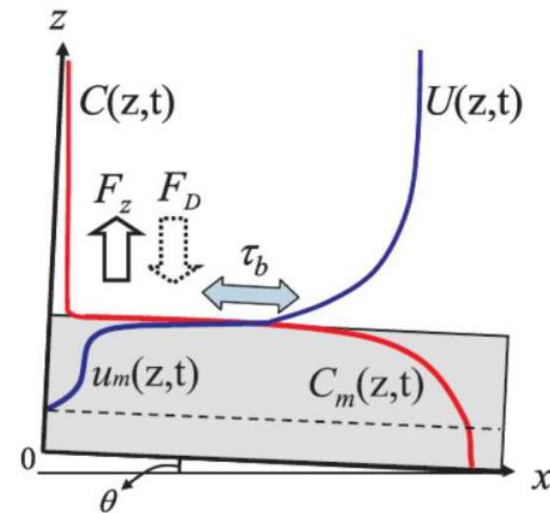


図-4 底泥界面近傍での密度（濃度）分布と輸送過程

$$\begin{aligned}
& \frac{\partial N}{\partial t} + \frac{\partial}{\partial x_i} \left(\left(u_i - \delta_{i,3} \frac{(1 - \phi_*)(1 - \phi_p)}{(1 + 2.5\phi)} w_{s,r} \right) N \right) \\
& - \frac{\partial}{\partial x_i} \left((D_s + \Gamma_T) \frac{\partial N}{\partial x_i} \right) \\
& = -k'_A (1 - \phi_*) G D^3 N^2 \\
& \quad + k_B G^{q+1} (D - D_p)^p D^{2q} N + E_{b,N}, \quad (15)
\end{aligned}$$

in which the parameters k'_A and k_B are defined as follows (e.g. Winterwerp, 1998).

$$\begin{aligned}
k'_A &= \frac{3}{2} e_c \pi e_d \quad \text{and} \\
k_B &= a e_b D_p^{-p} \left(\frac{\mu}{F_y} \right)^q,
\end{aligned}$$

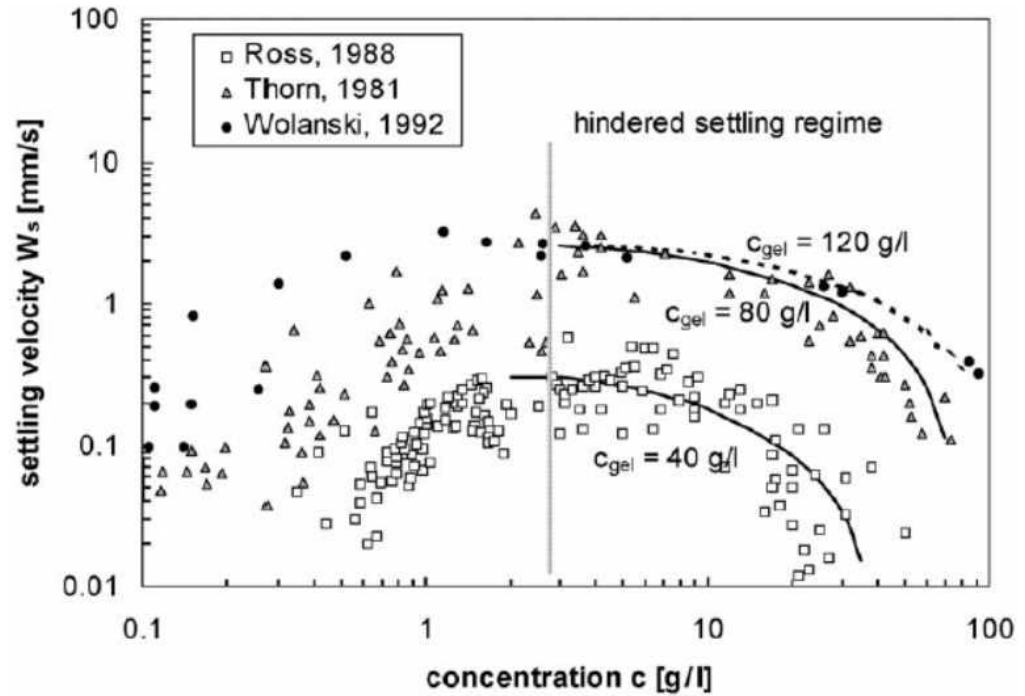
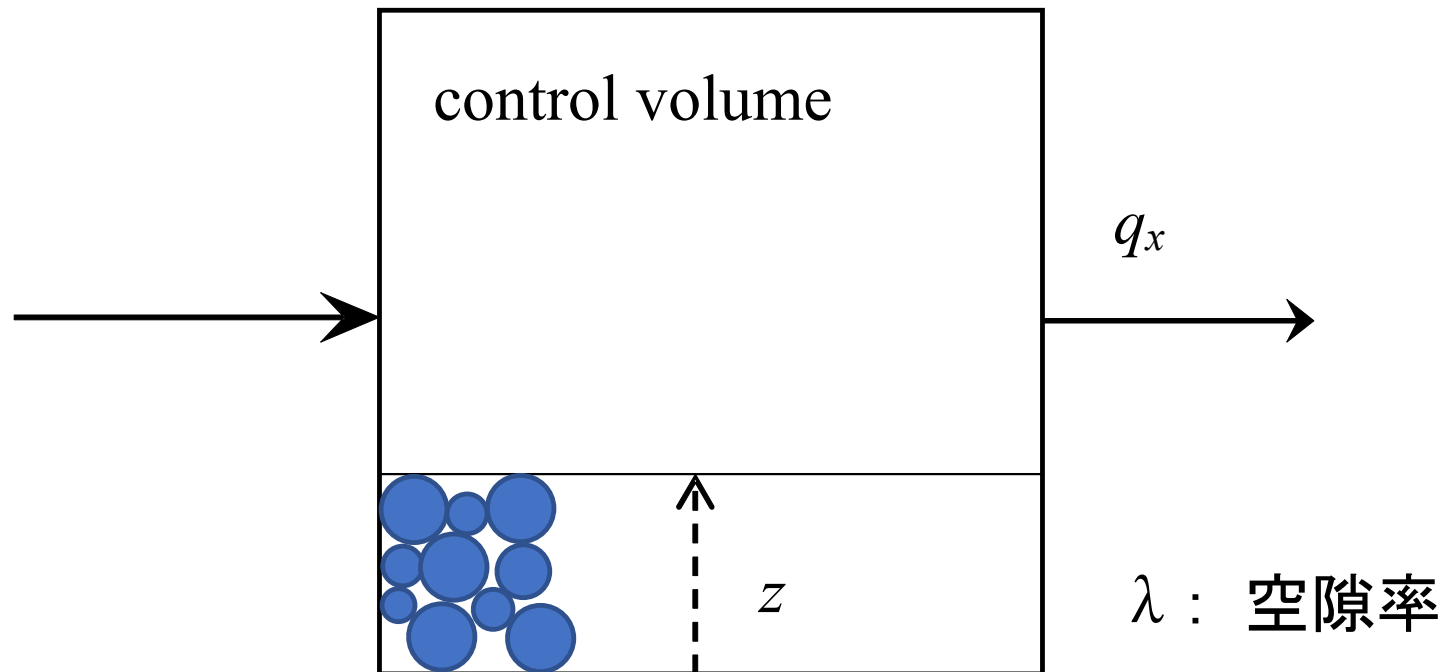


Fig. 1. Hindered settling; comparison of Eq. (6) with data from literature.

土砂収支式：
$$\frac{\partial z}{\partial t} + \frac{1}{1 - \lambda} \left(\frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} \right) = 0$$

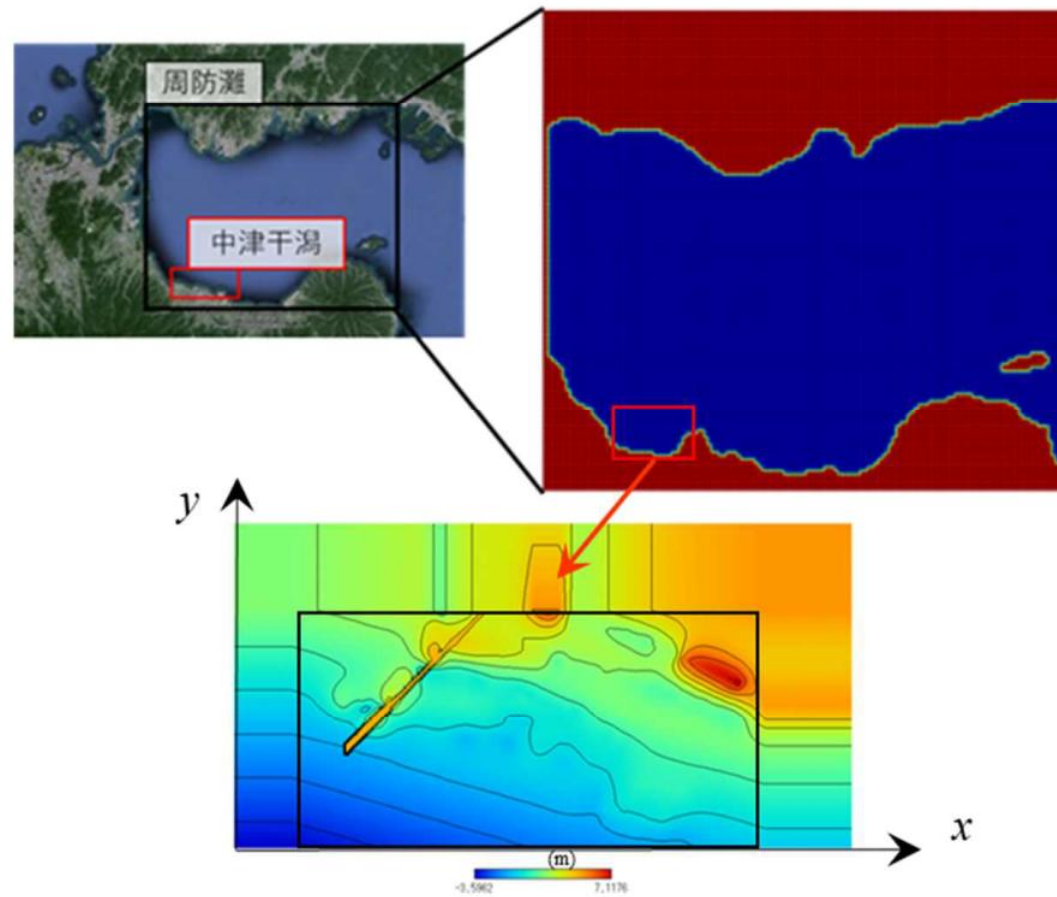


2017年出水による泥質化の再現計算

WDM-POM: Uchiyama (2005), 鵜崎ら (2007, 2018)

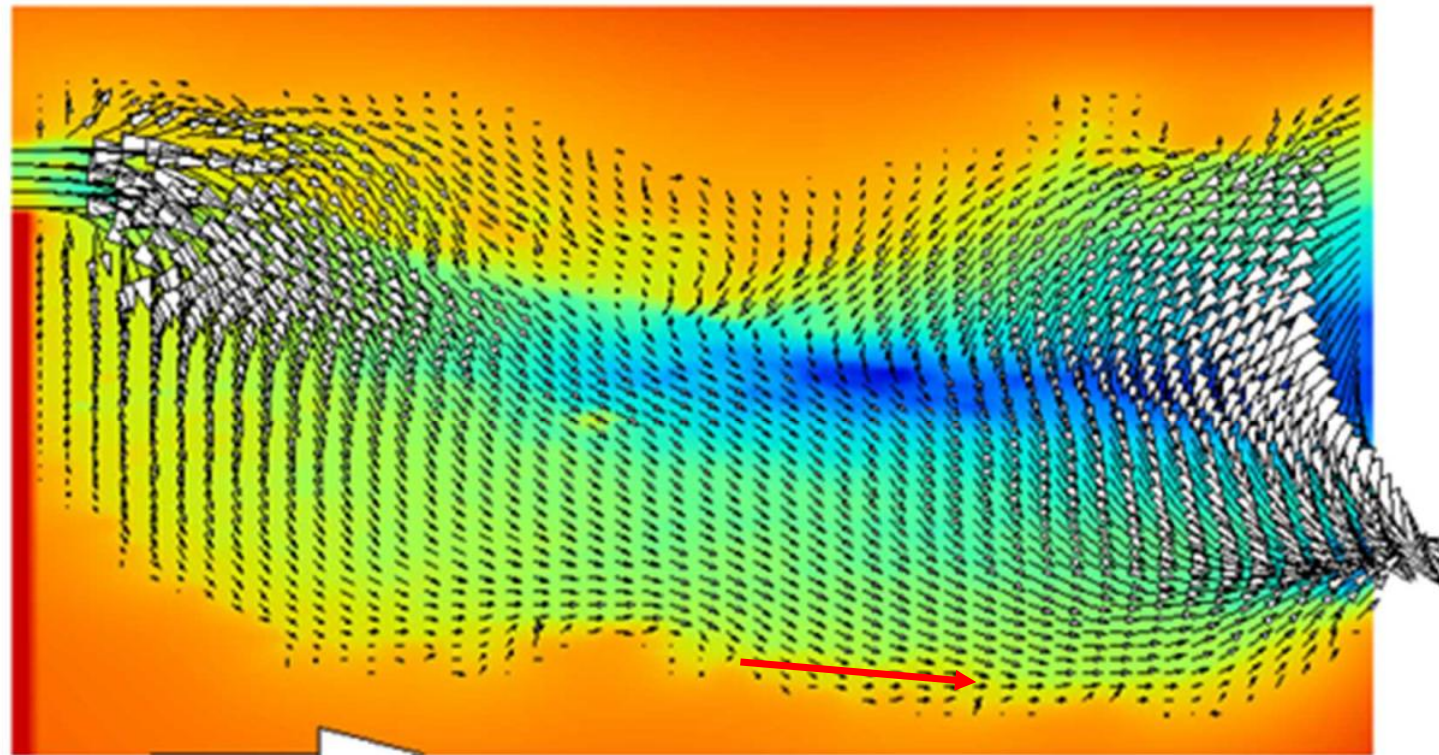
計算条件表

		大領域	小領域	
メッシュ数	nx	146	216	
	ny	131	114	
メッシュサイズ	dx	450	55.66	m
	dy	550	45.34	m
時間刻み	dte	1.0	0.1	s
	dte	0.1	0.01	s
	計算領域	流れ		
Run 1	大領域	潮流		
Run 2	小領域	潮流		
Run 3	小領域	潮流	海浜流	

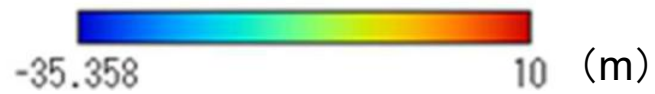


計算領域

□開境界条件：別府・徳山・関門：実測潮位，線形補間

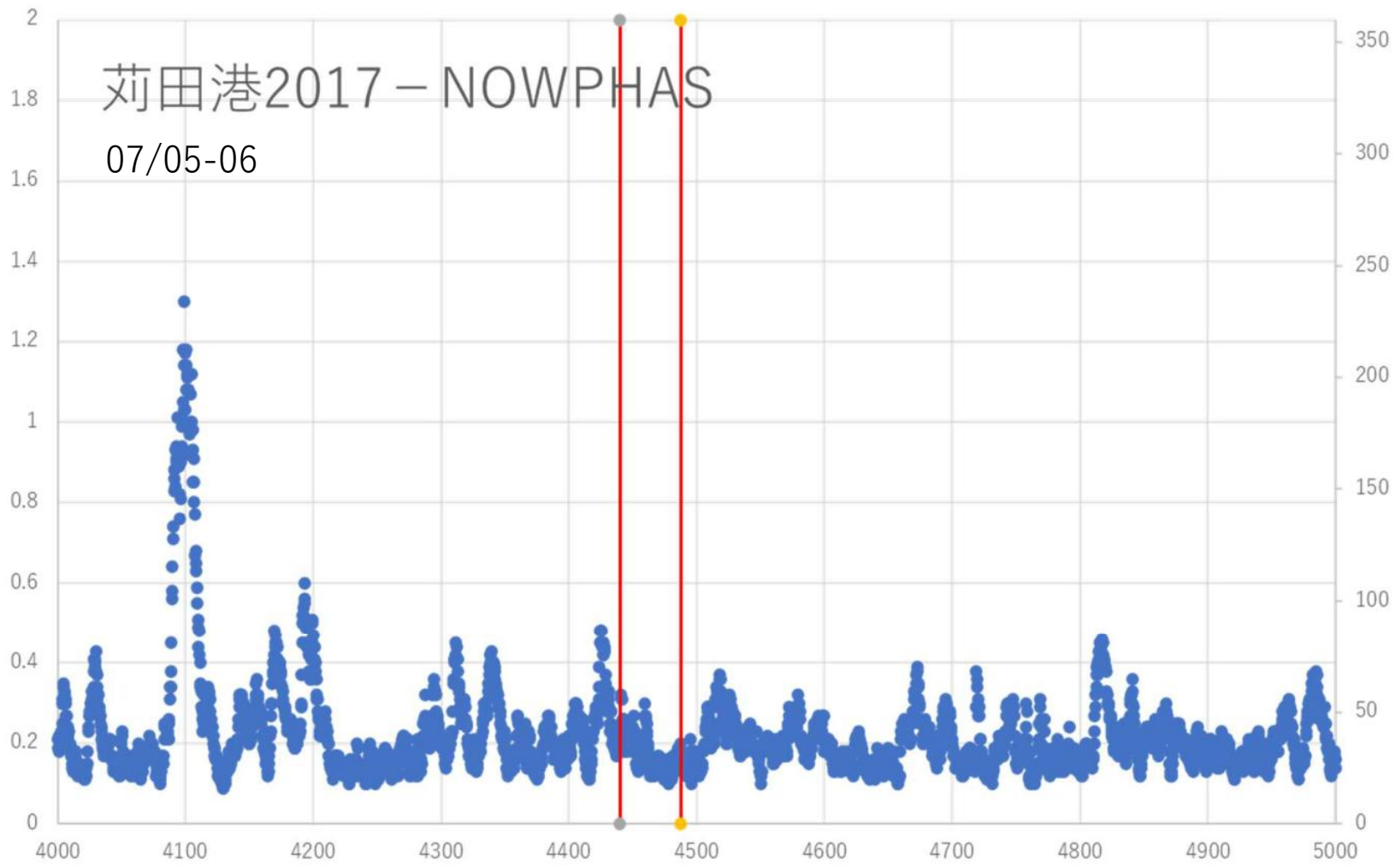


1.0 m/s



周防灘における潮汐残差流

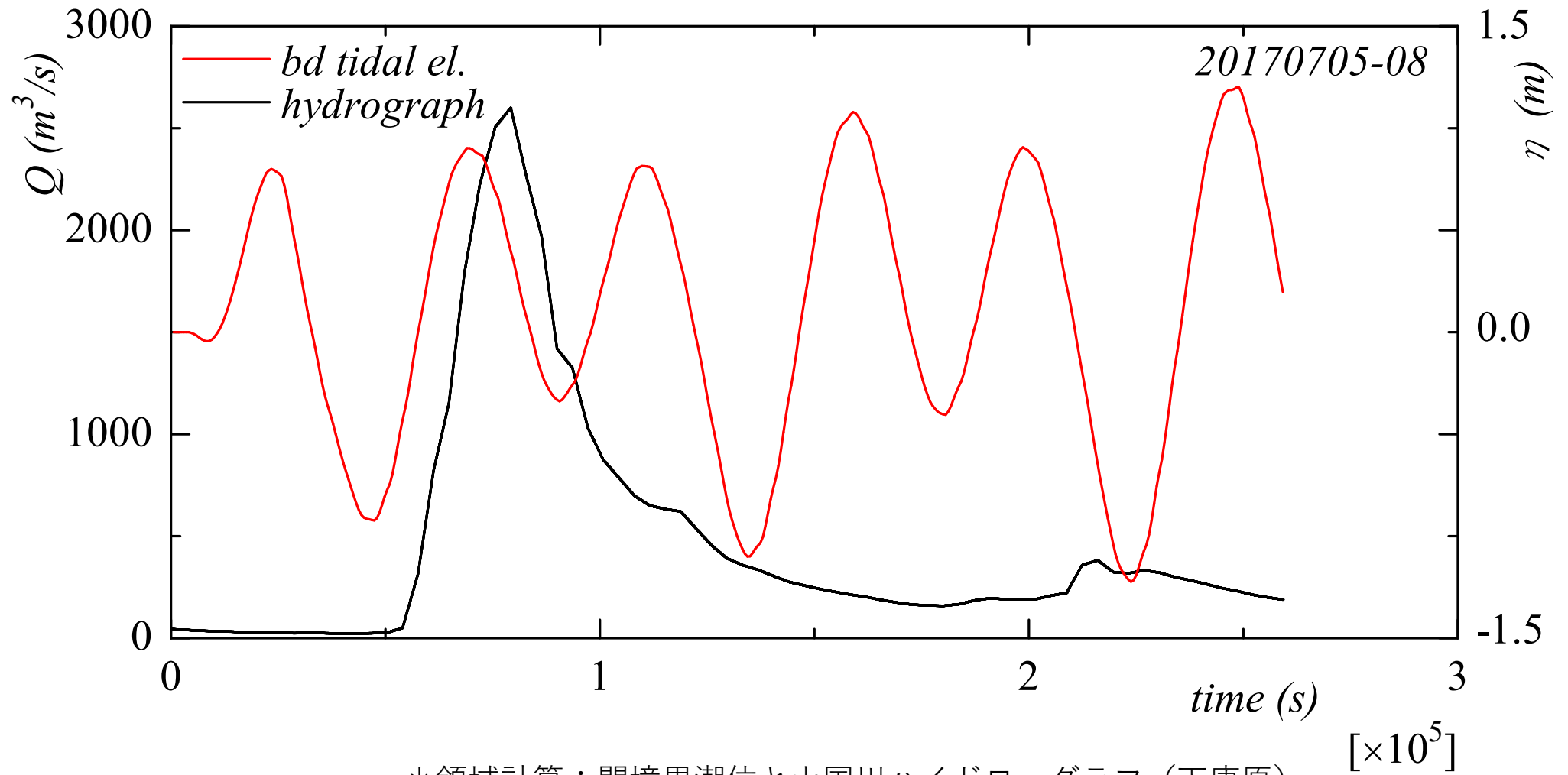
有義波高
(m)



荻田港2017 - NOWPHAS

07/05-06

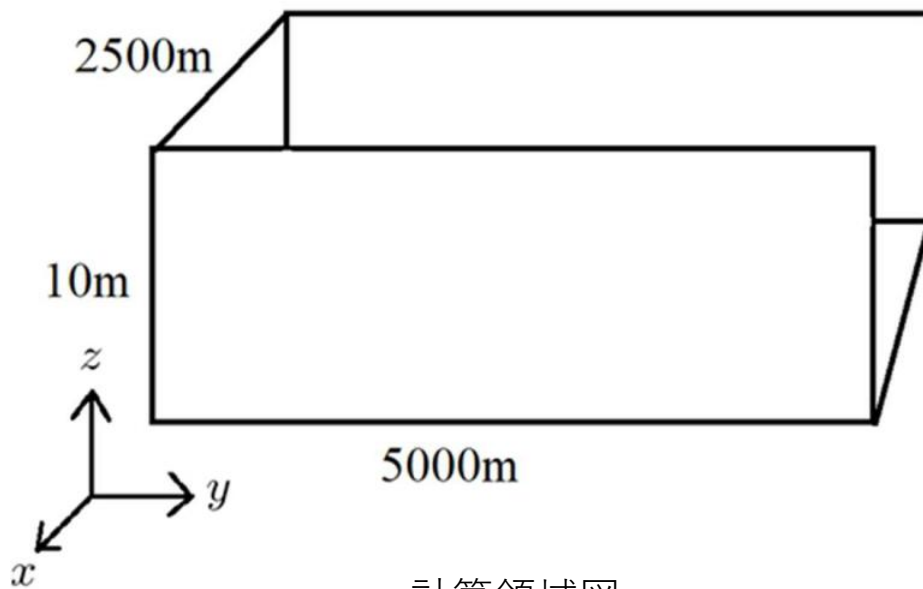
時間



小領域計算：開境界潮位と山国川ハイドロ・グラフ（下唐原）

■定常的な汀線際の堆泥域形成

参考：蔵田ら（2017），干潟域への干拓堤防建設と凝集性懸濁物輸送特性の関係に関する数値計算，流体力学学会年会2017



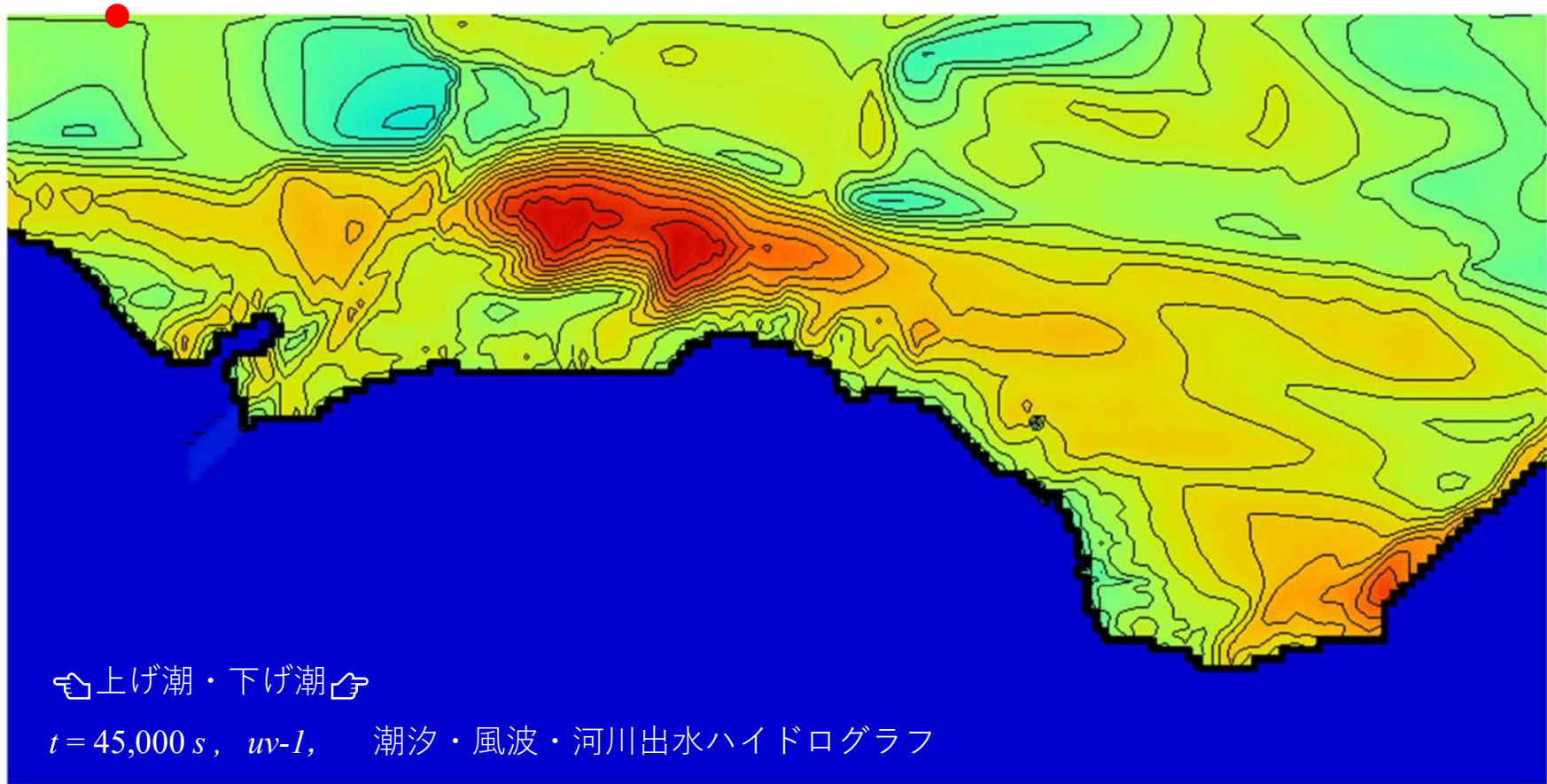
計算領域図

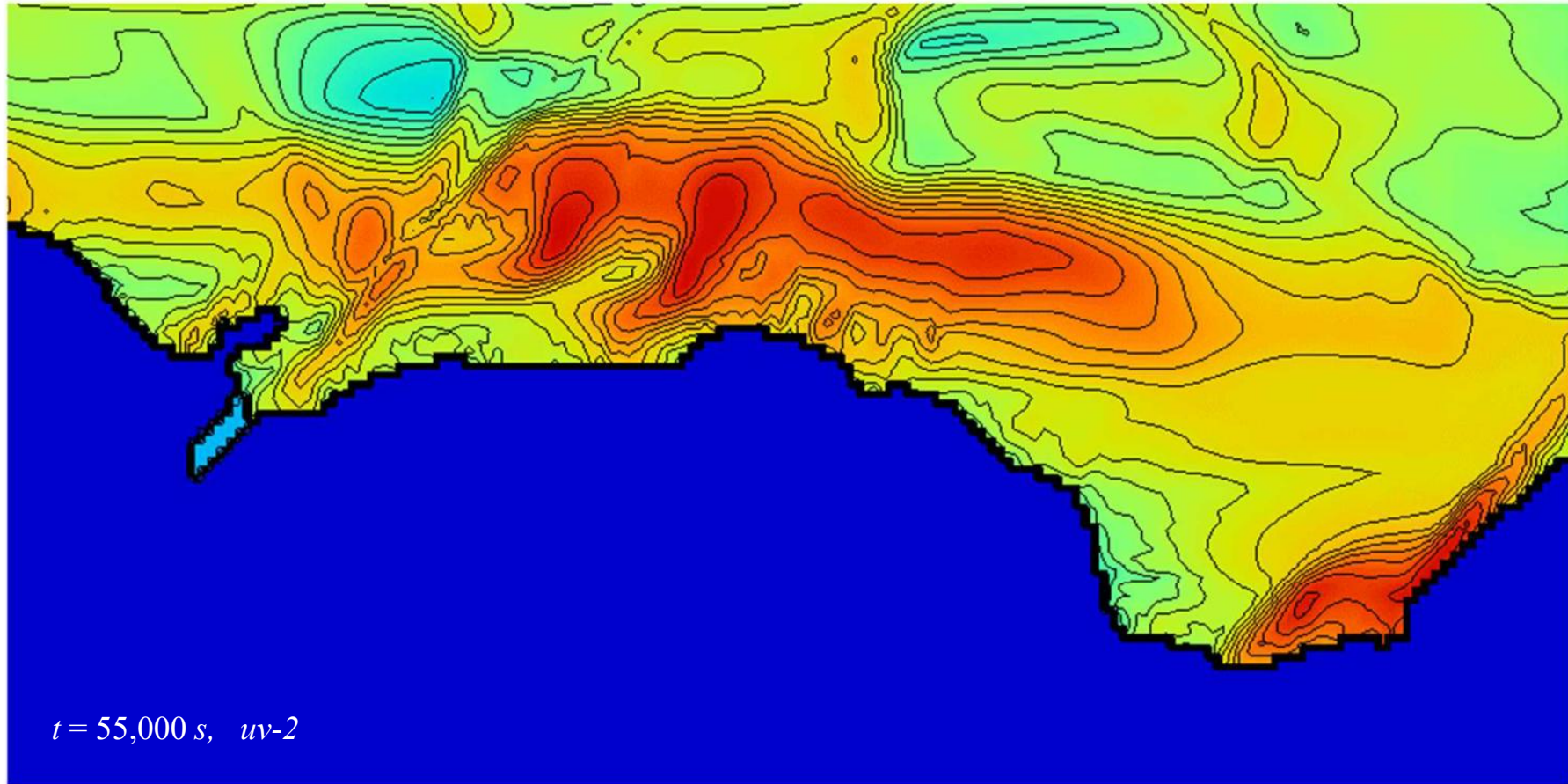
計算条件表

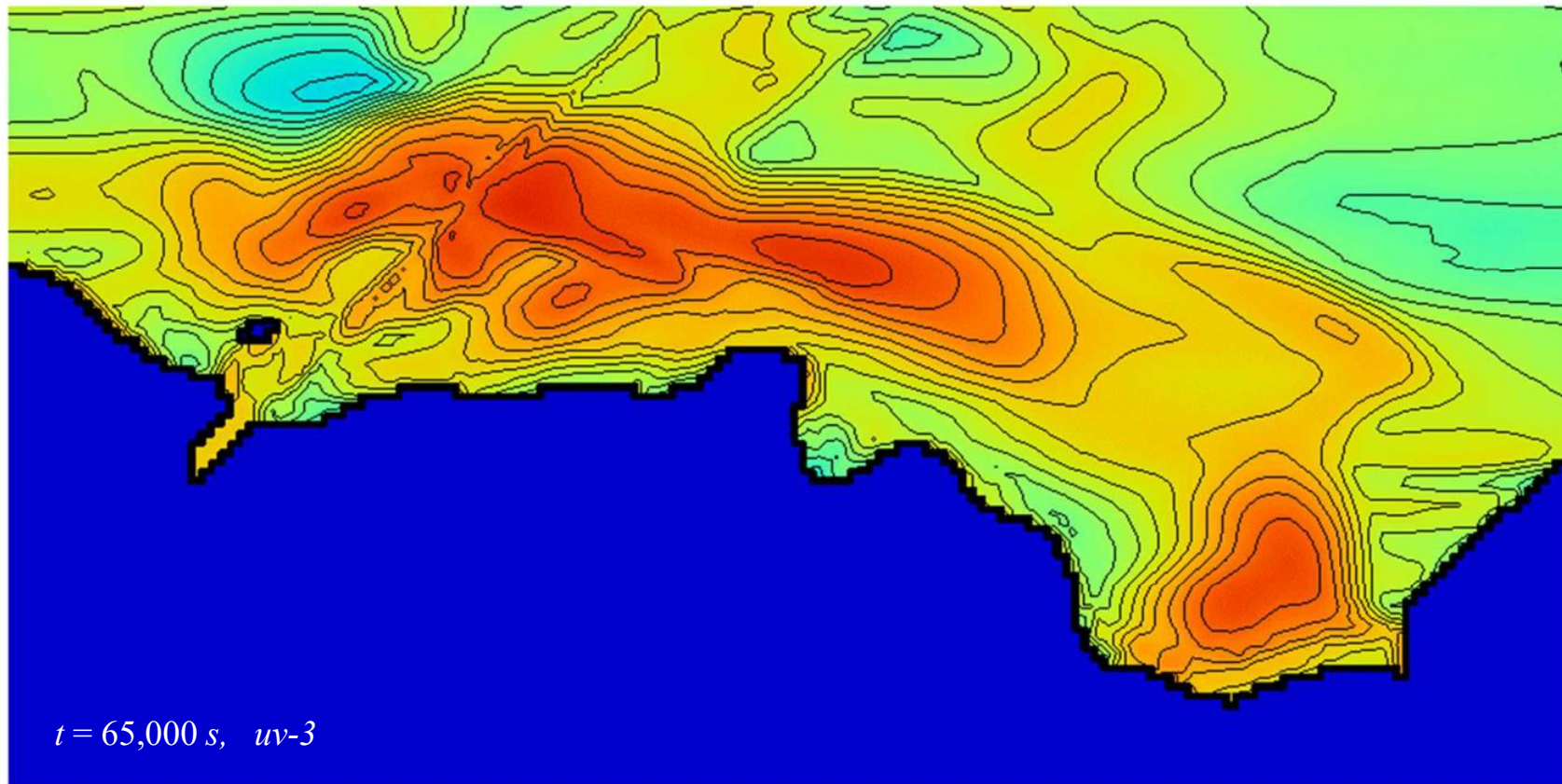
	記号	値	単位
土粒子密度	ρ	$1.2 * 10^3$	kg / m^3
含泥率	-	30	%
限界せん断応力	τ_c	0.1	N / m^2
侵食速度係数	P_e	2.5	$\text{kg} / (\text{m}^2 * \text{s})$
最小せん断応力	$\tau_{b, \min}$	$* 10^{-5}$	N / m^2
最小沈降速度	-	0.15	m / s
		6.6	
		$* 10^{-6}$	

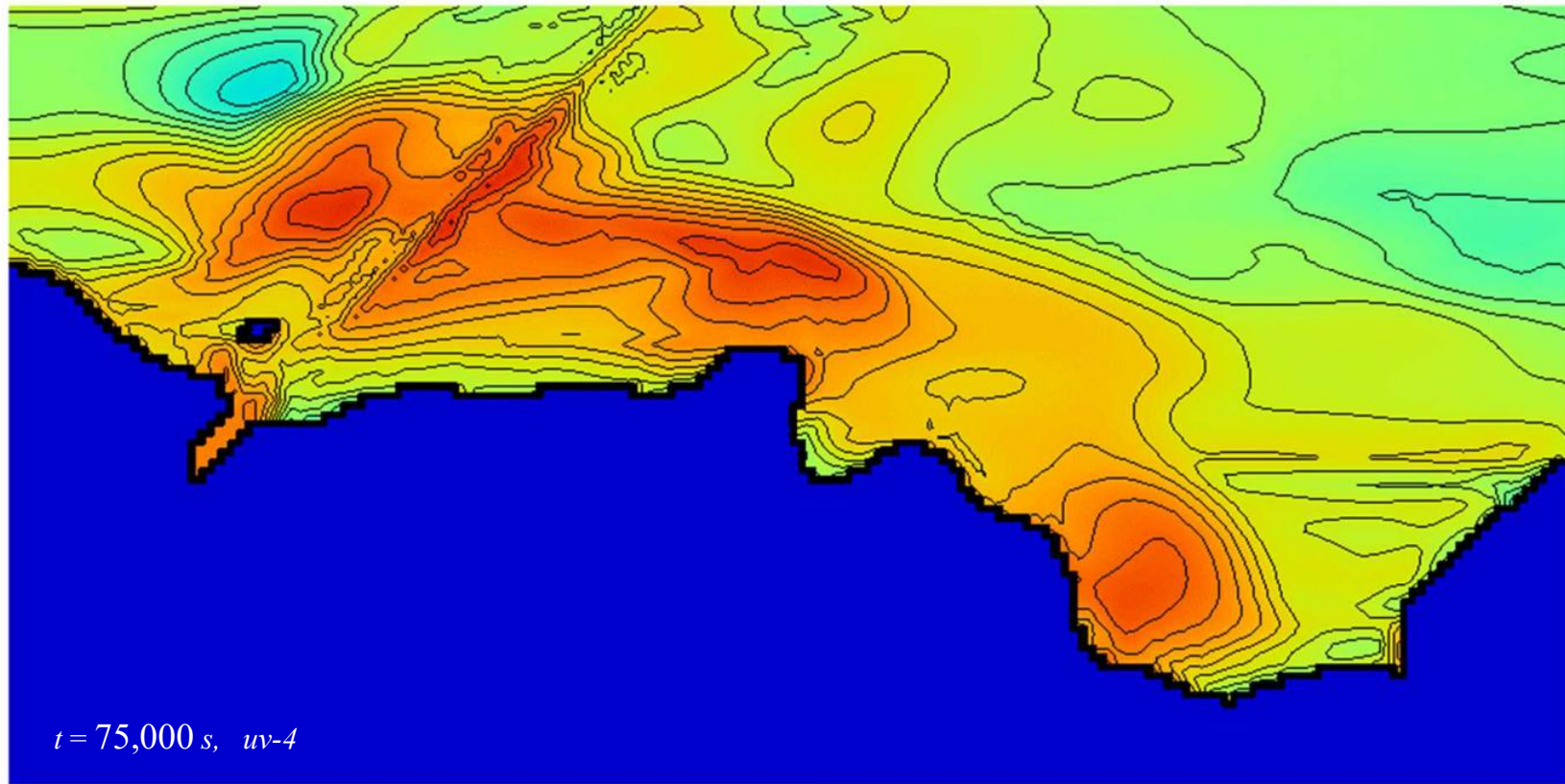
◆開境界振幅： $a=1.0 \text{ m}$ ，周期： $T=12.0\text{h}$ ，sin波

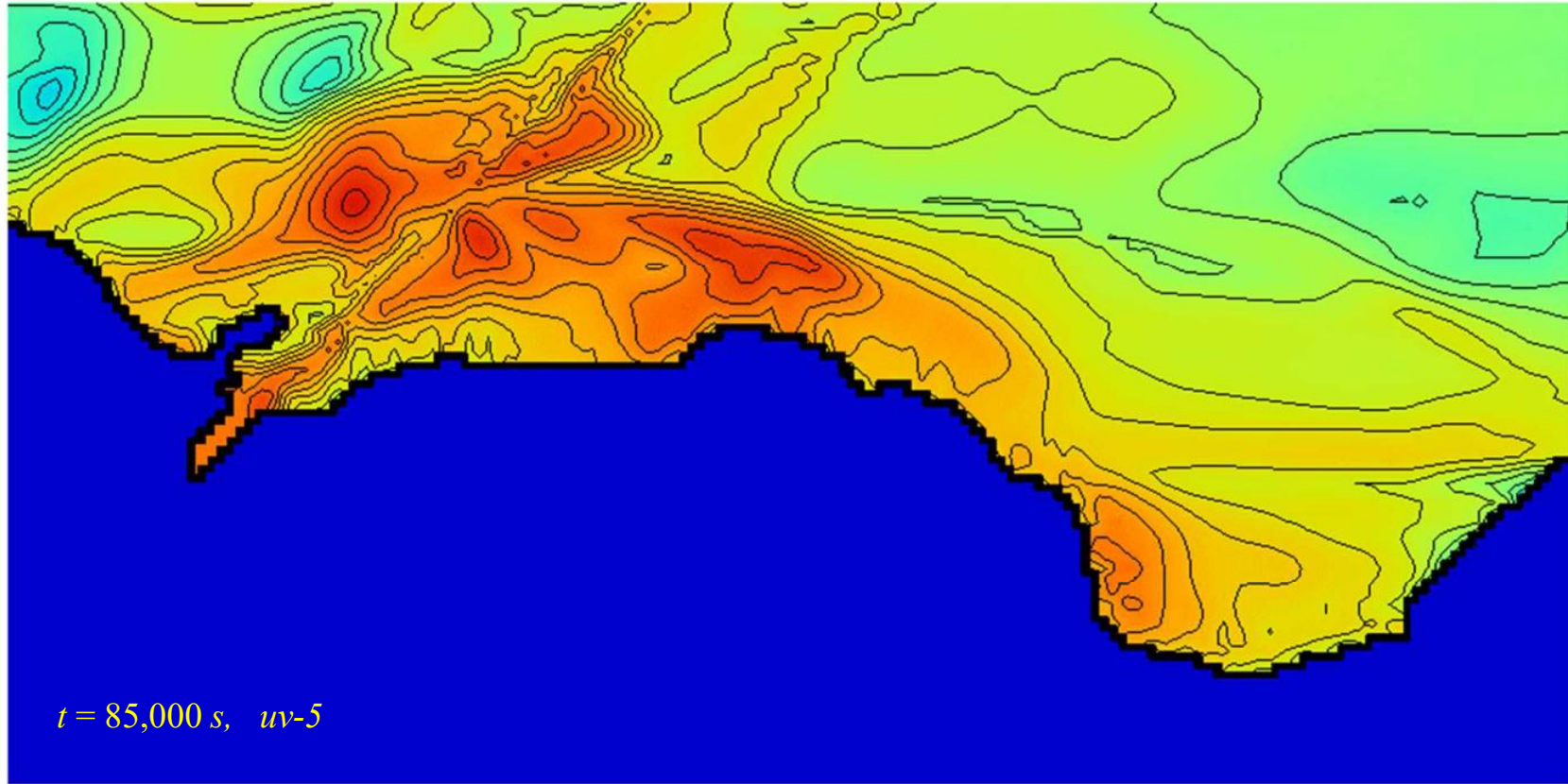
3.2 出水による河口—大新田地区沖の堆泥





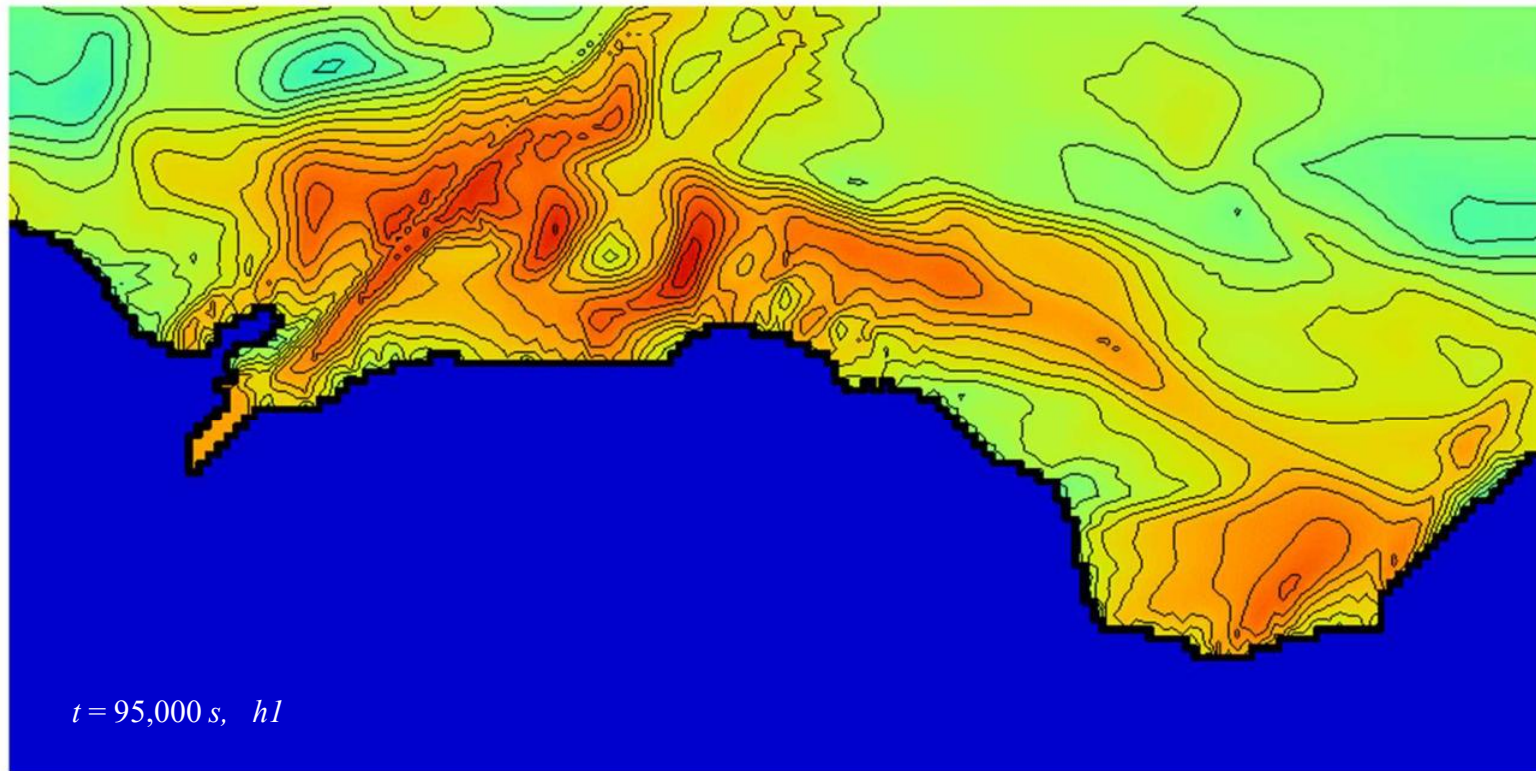


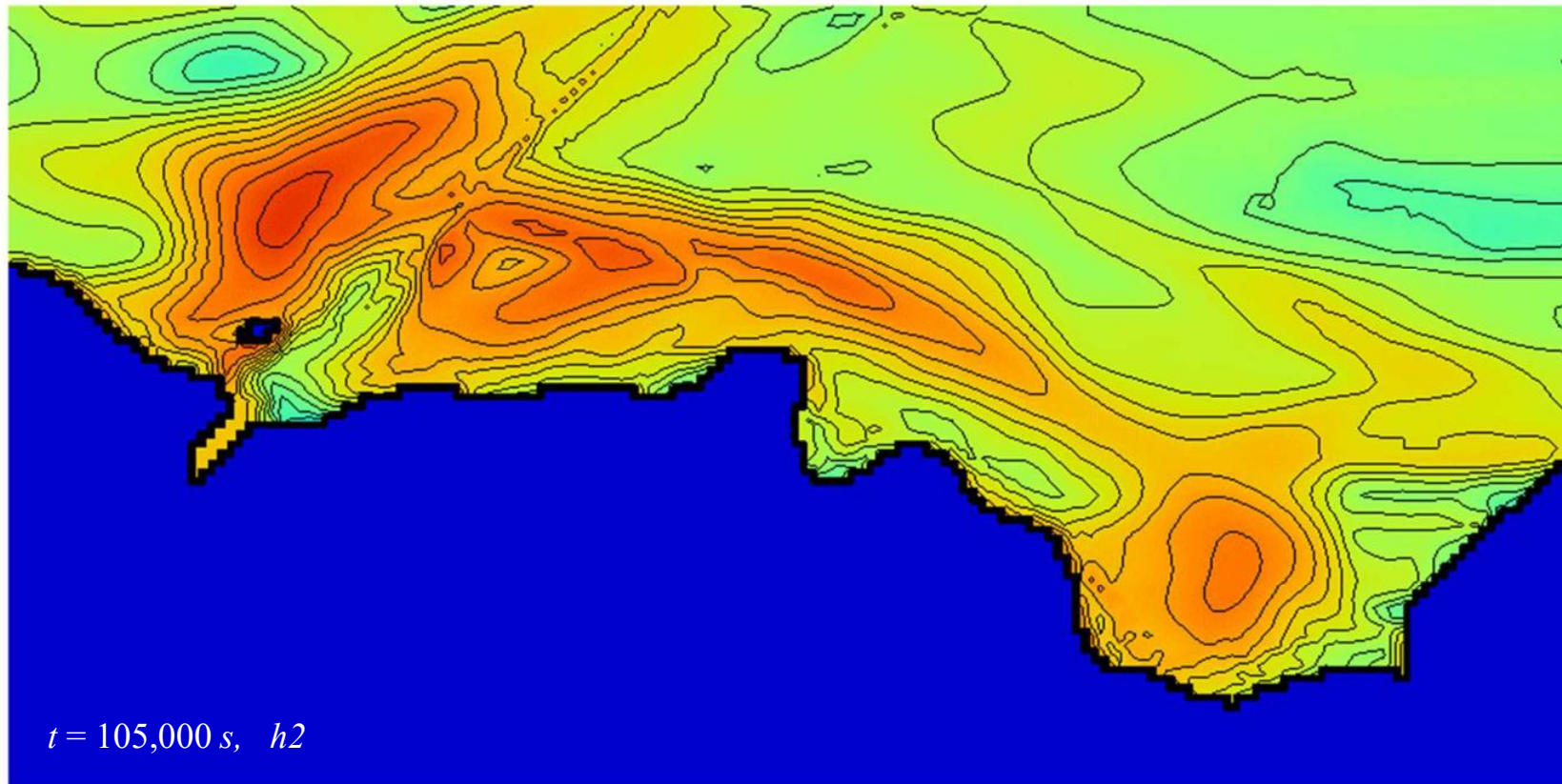


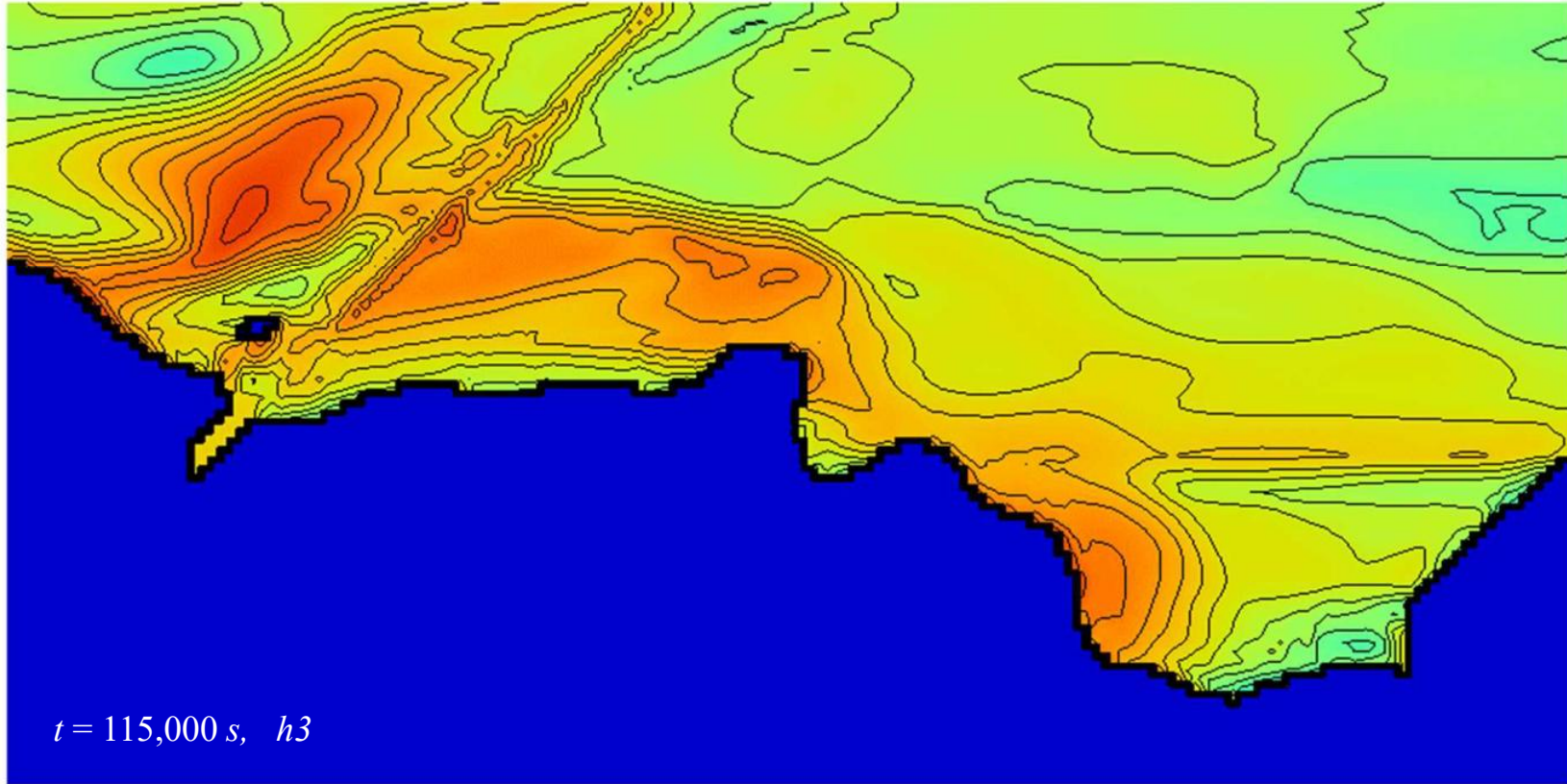


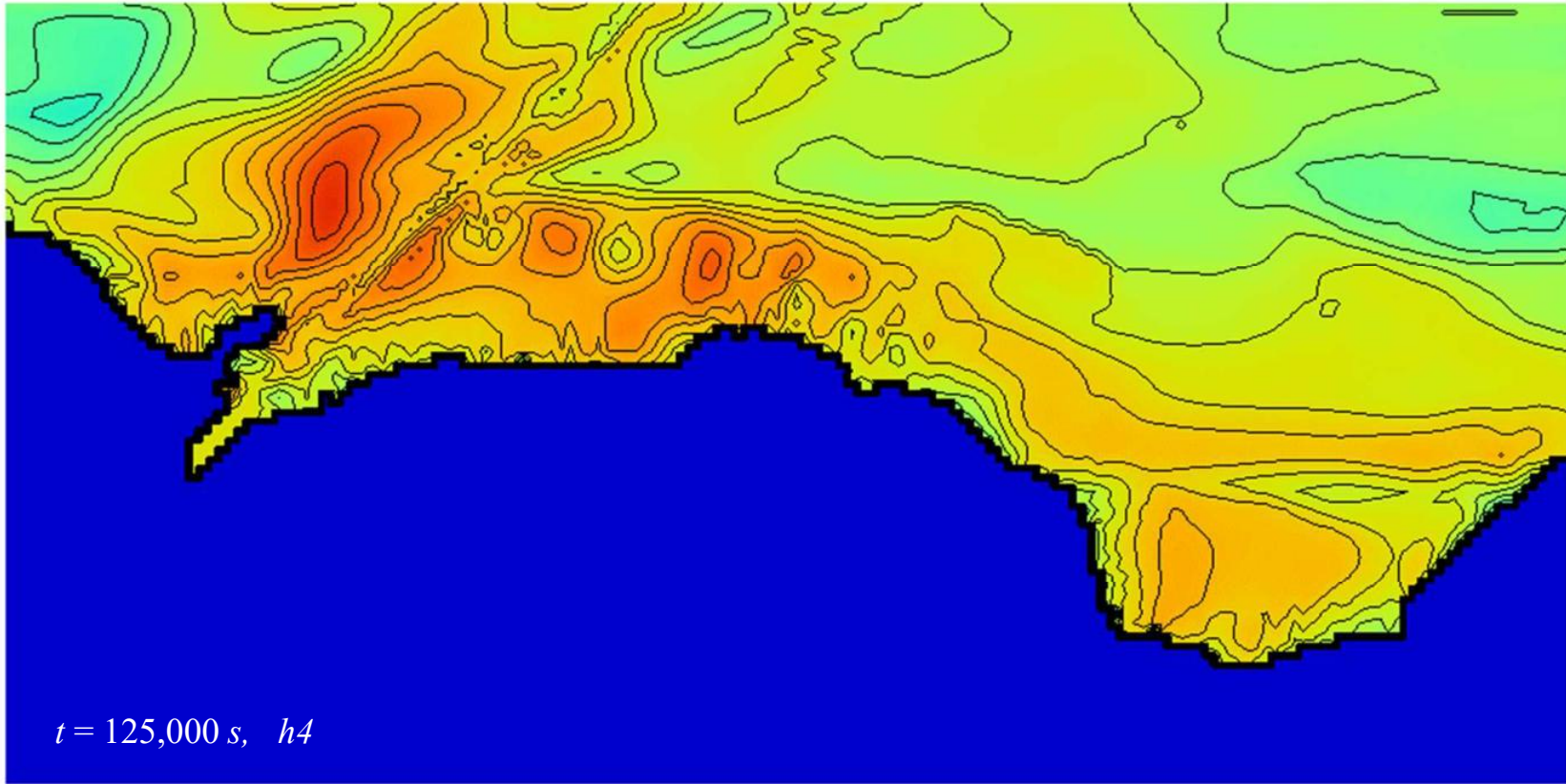
$t = 85,000$ s, $uv-5$

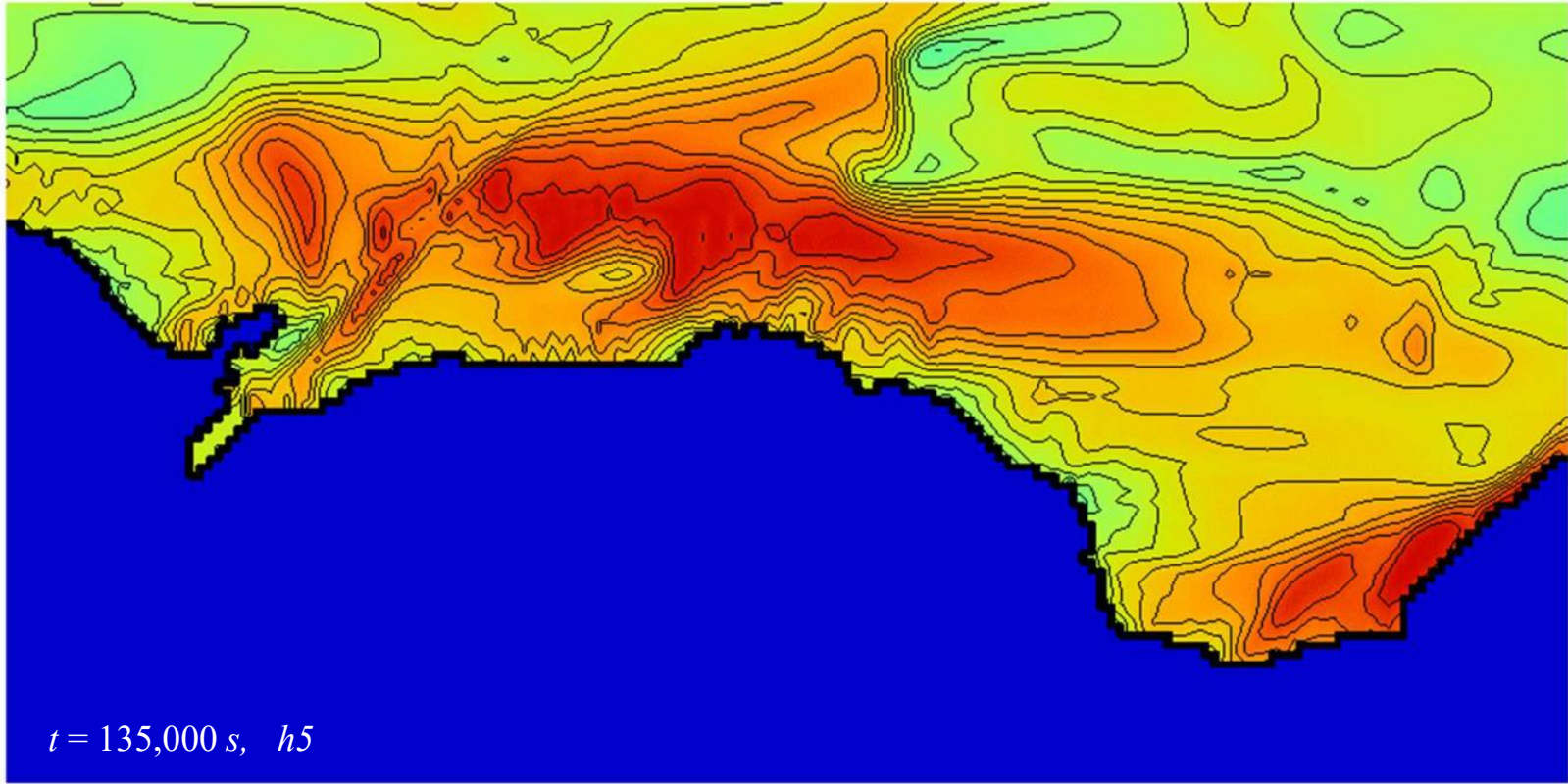


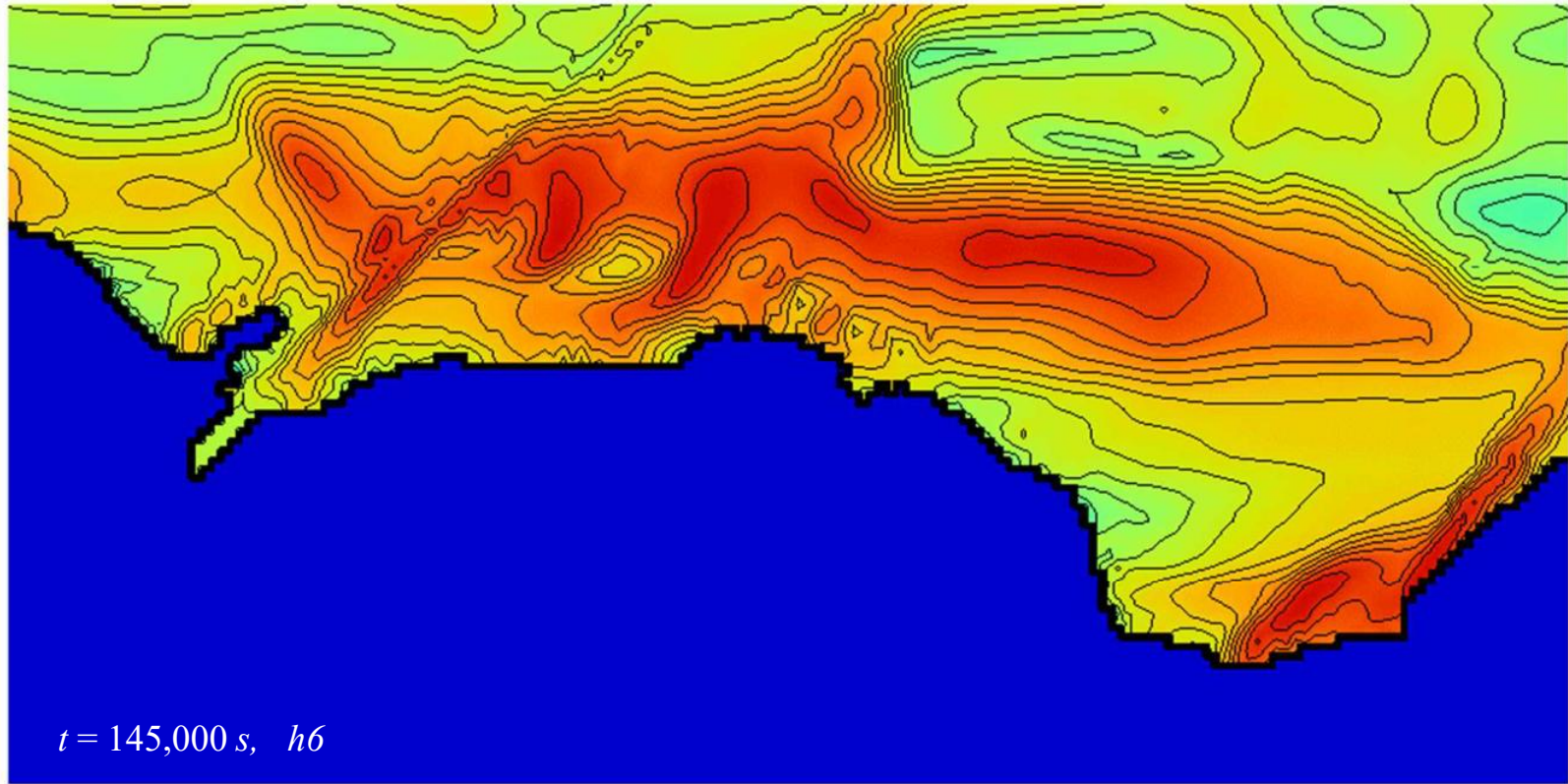


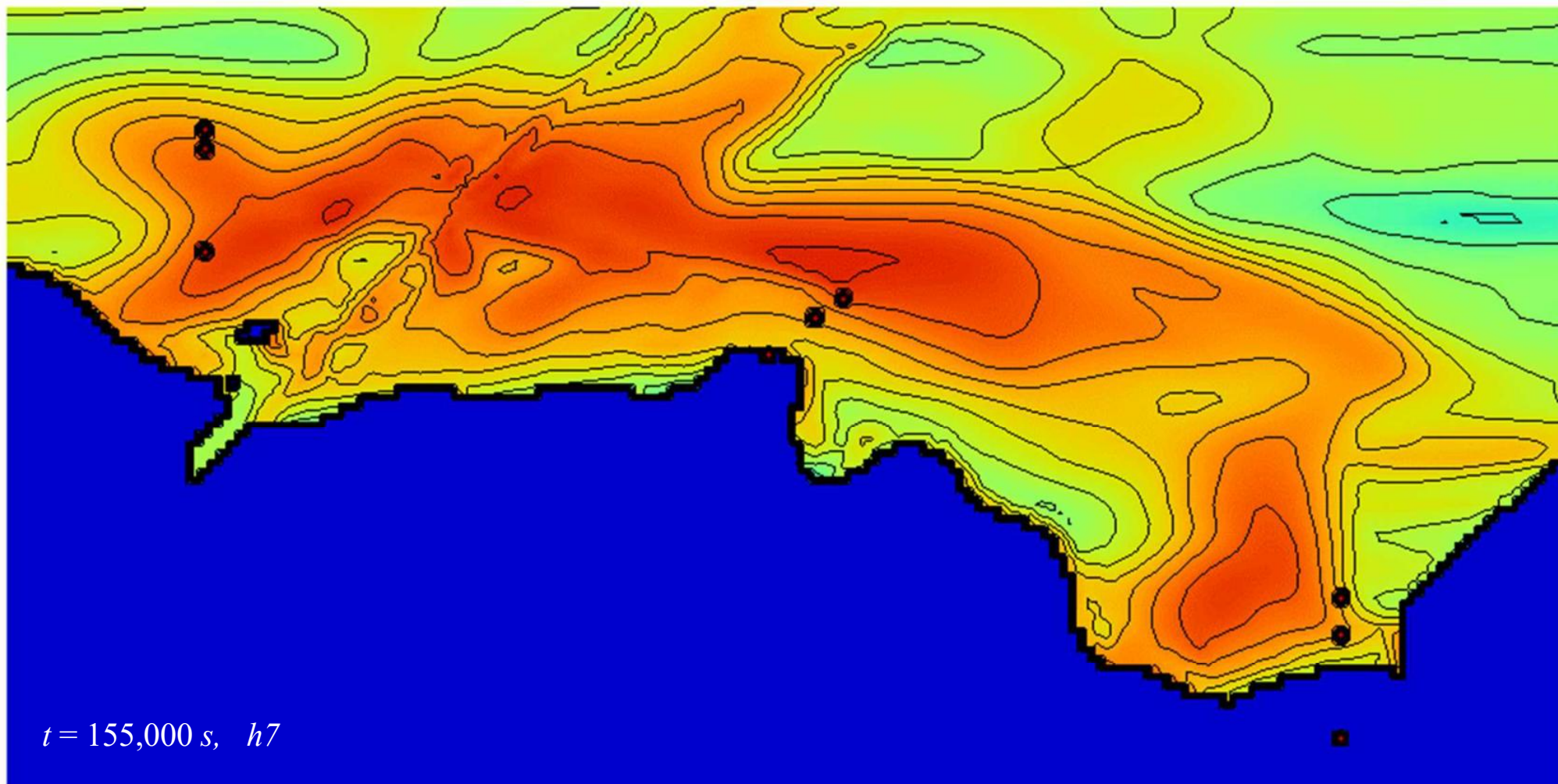












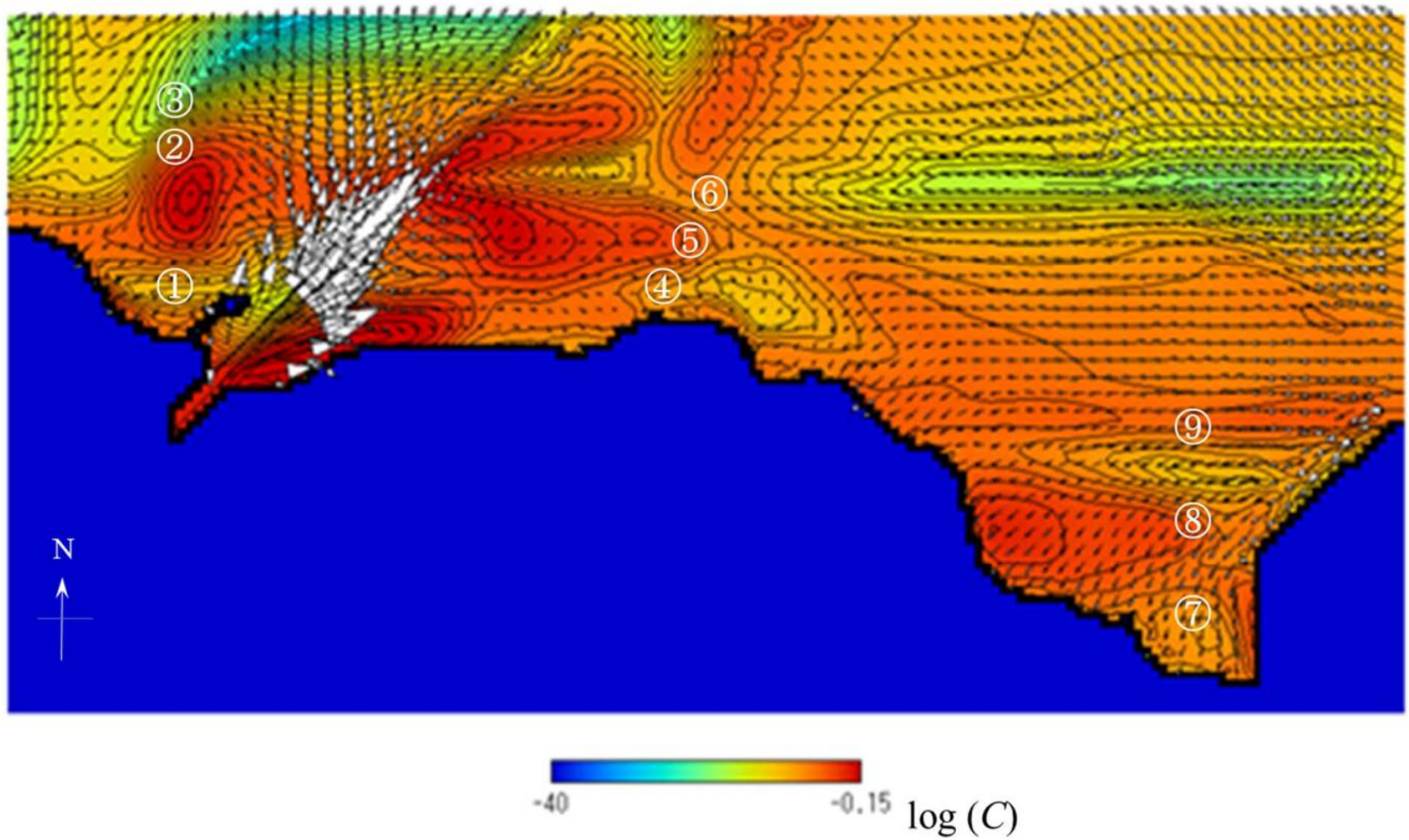
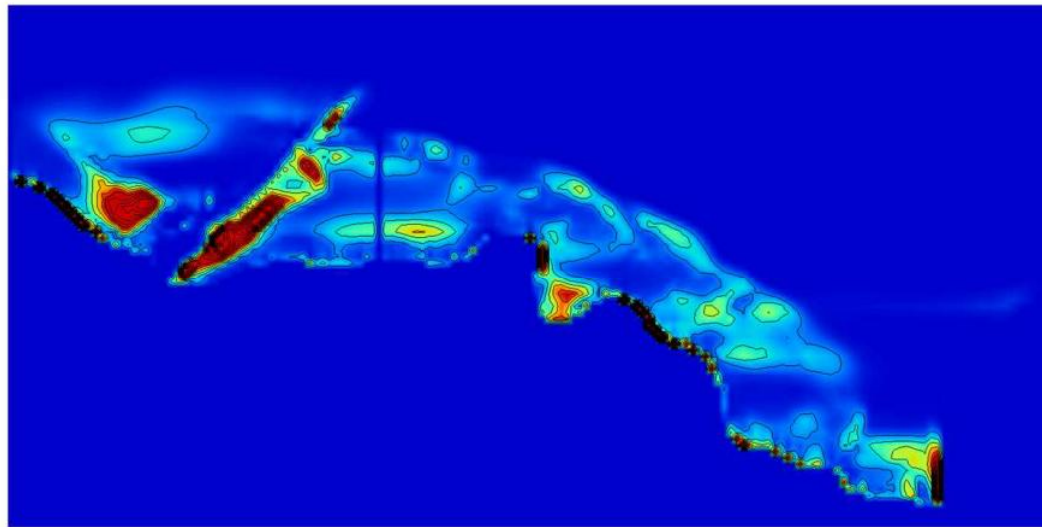
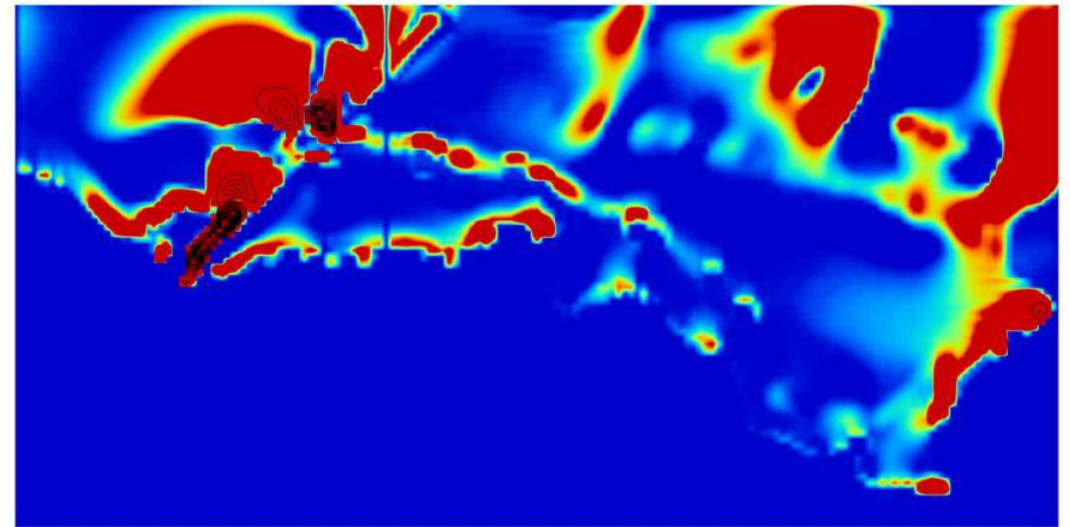


図-5 浮遊泥濃度の空間分布 (Run 4, $t = 30,000$ s)



2.8E-23 0.02 $\log(dhsilt)$

(a) 堆泥厚

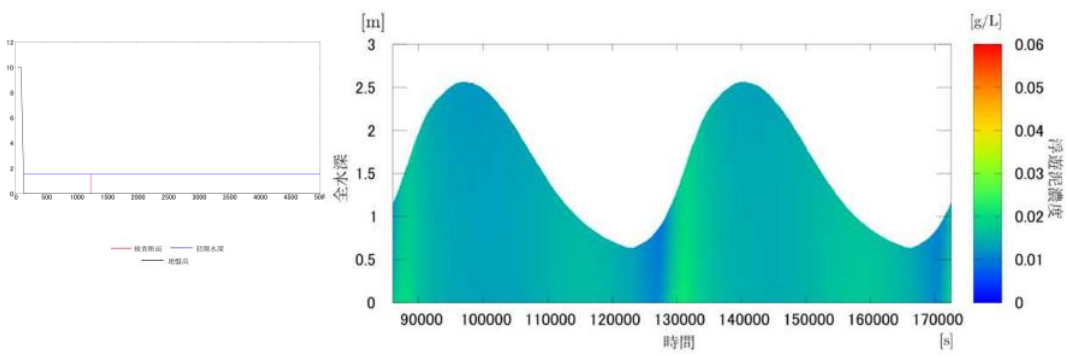


0 0.0001 $\log(dhsed)$

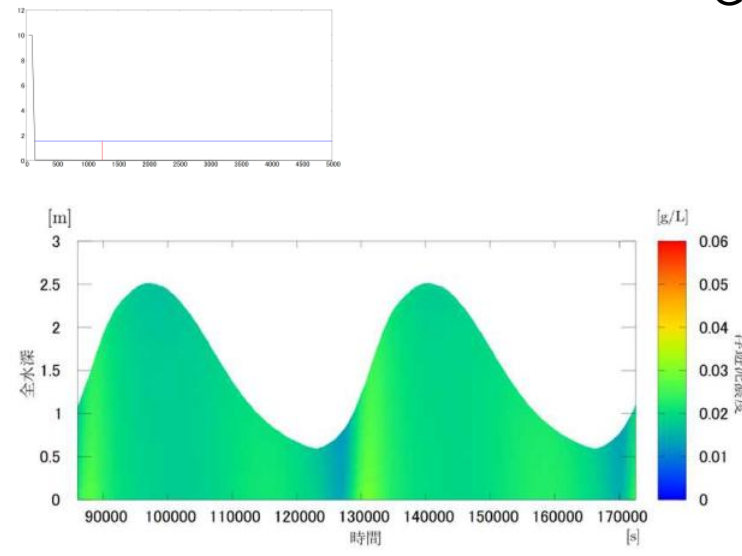
(b) 堆砂厚

図-6 堆泥厚と堆砂厚 ($t = 86,400$ s)

3.3 定常的な汀線際の堆泥－沖合堆泥をもとにした上げ潮による再懸濁・再凝集・再沈降過程

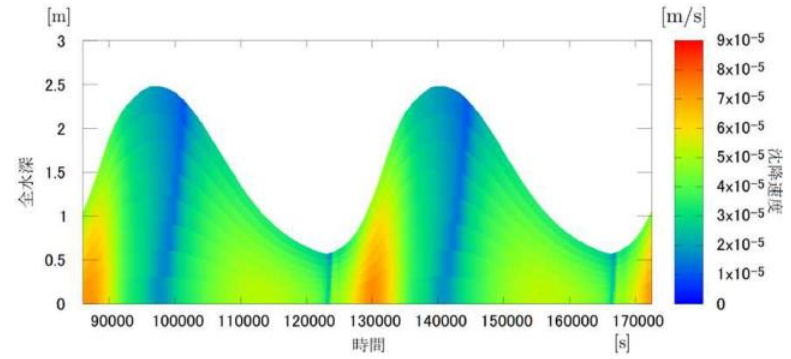
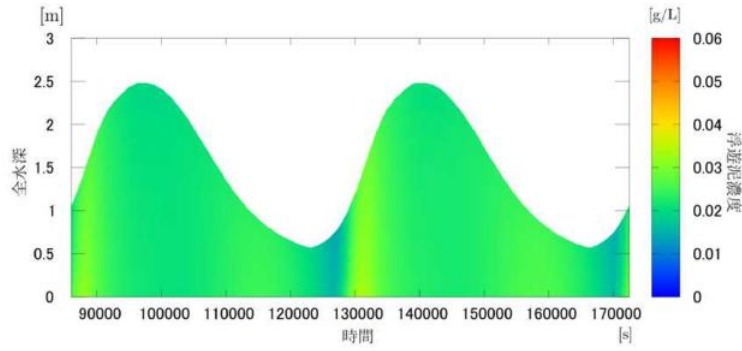
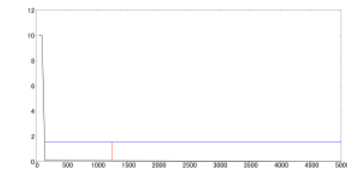


Grad 0

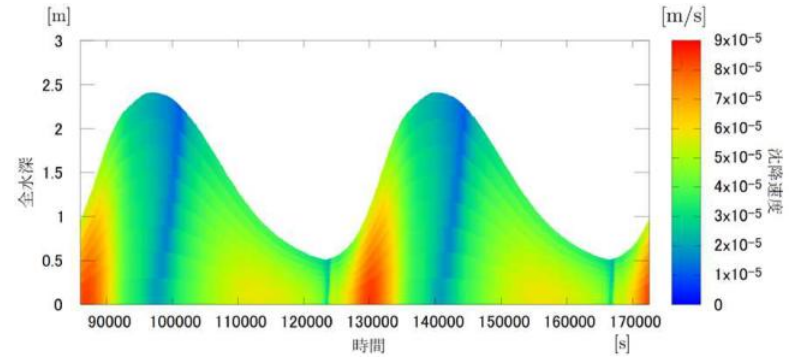
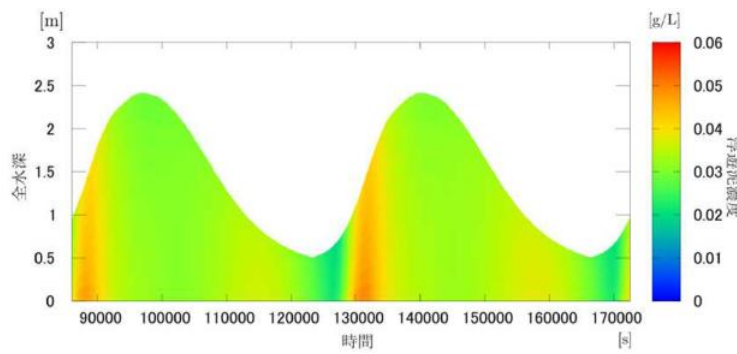
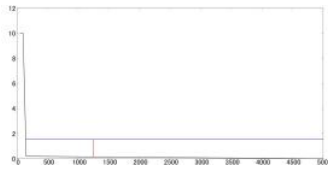


Grad 1/1500

浮遊泥濃度(左)と沈降速度(右)



Grad 1/1000



Grad 1/500

浮遊泥濃度(左)と沈降速度(右)

■ 広域土砂動態・地形変化モデル：SPR-WDMPOM

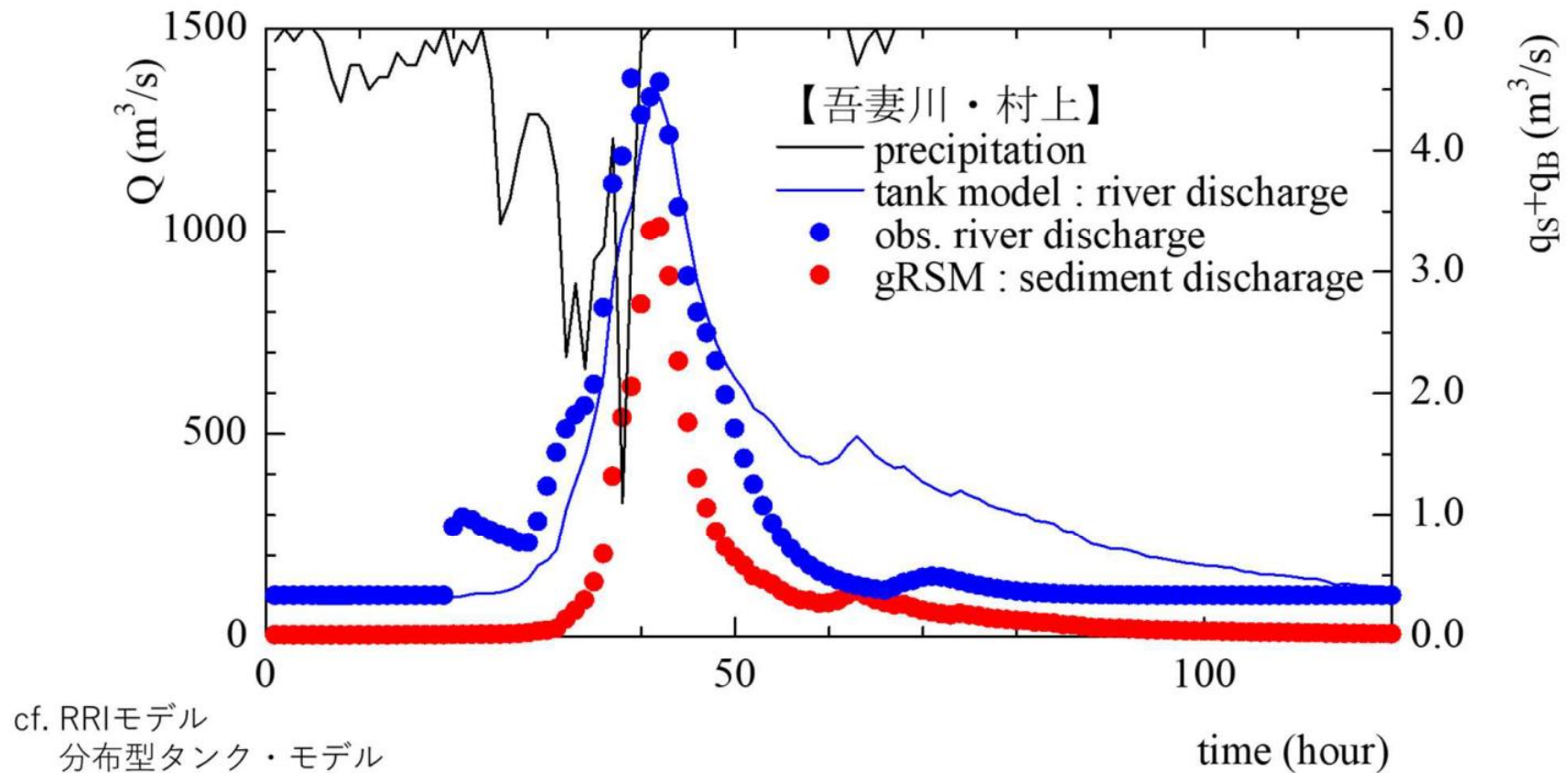
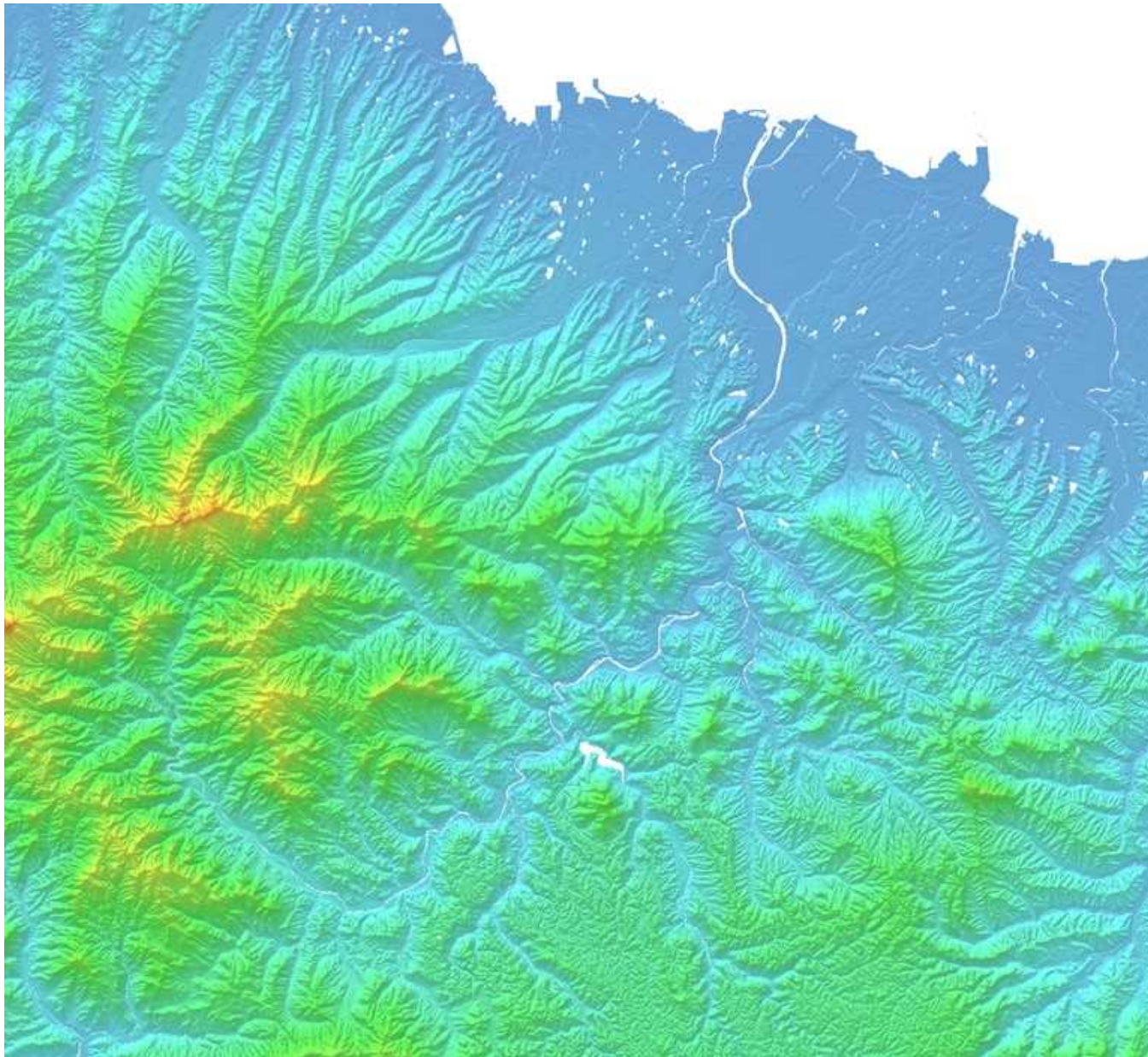
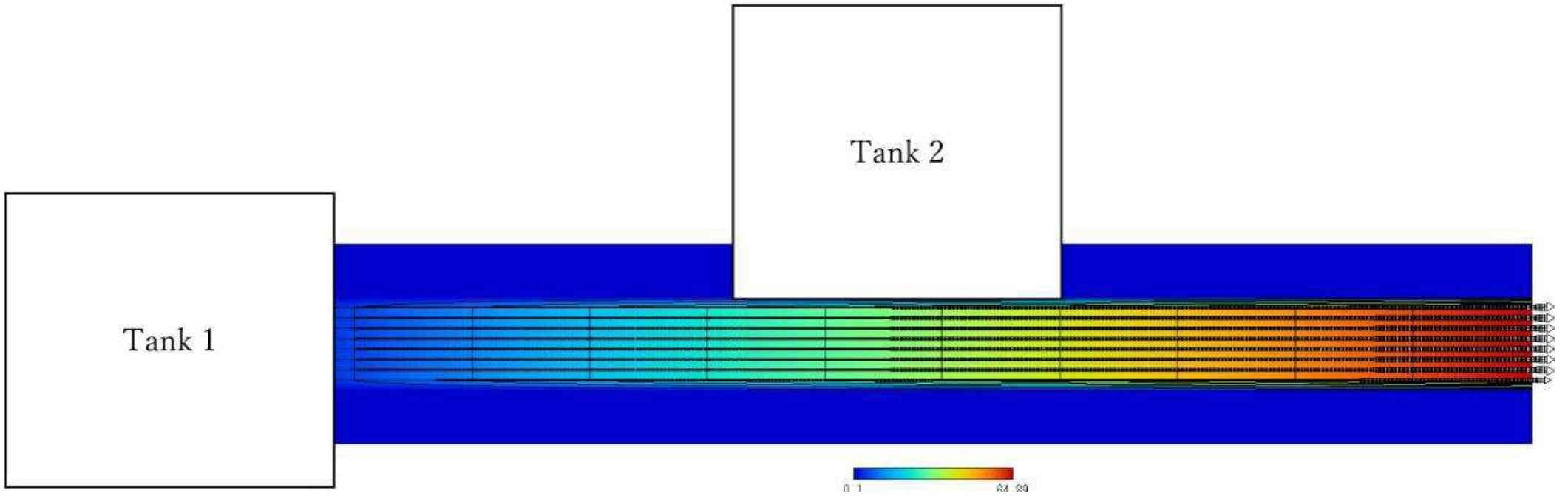


図-7 利根川上流域・吾妻川におけるtank modelとgRSMによる土砂流出解析



山国川流域5mDEM
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【結 論】

山国川・中津干潟と利根川中上流域を対象として現地観測と数値計算を行い、gRSM、WDM-POMと連結して「広域土砂動態・地形変化モデル：SPR-WDMPOM」を構築して現地適用を行った。

■山国川・中津干潟

- ①2000年からの観測データ（国土交通省データも含む）から、中津港拡張事業中以降の侵食・堆積傾向はあまり明瞭ではないが、大規模降雨による堆積とそれ以外の侵食、ならびに近年、侵食・堆積の振幅の増大が認められた。
- ②2017年の水害出水で泥質化、18・22年は細砂・中砂化（細粒化）が示された。
- ③2017年の泥質化は、支川山移川流域に集中した水害による異常な数の斜面崩壊によるものと推察された。
- ④gRSMとWDM-POMによって17年の泥質化の数値計算を行い、河口から大新田 地区沖合への泥質化を再現された。
- ⑤トレンチ調査結果ならびに中津土木事務所資料の再解析から、定常的な汀線際の堆泥域形成が示された。
- ⑥WDM-POMによって、沖合堆泥をソースとした、上げ潮による再懸濁・再凝集・再沈降過程による汀線際の堆泥機構が示唆された。

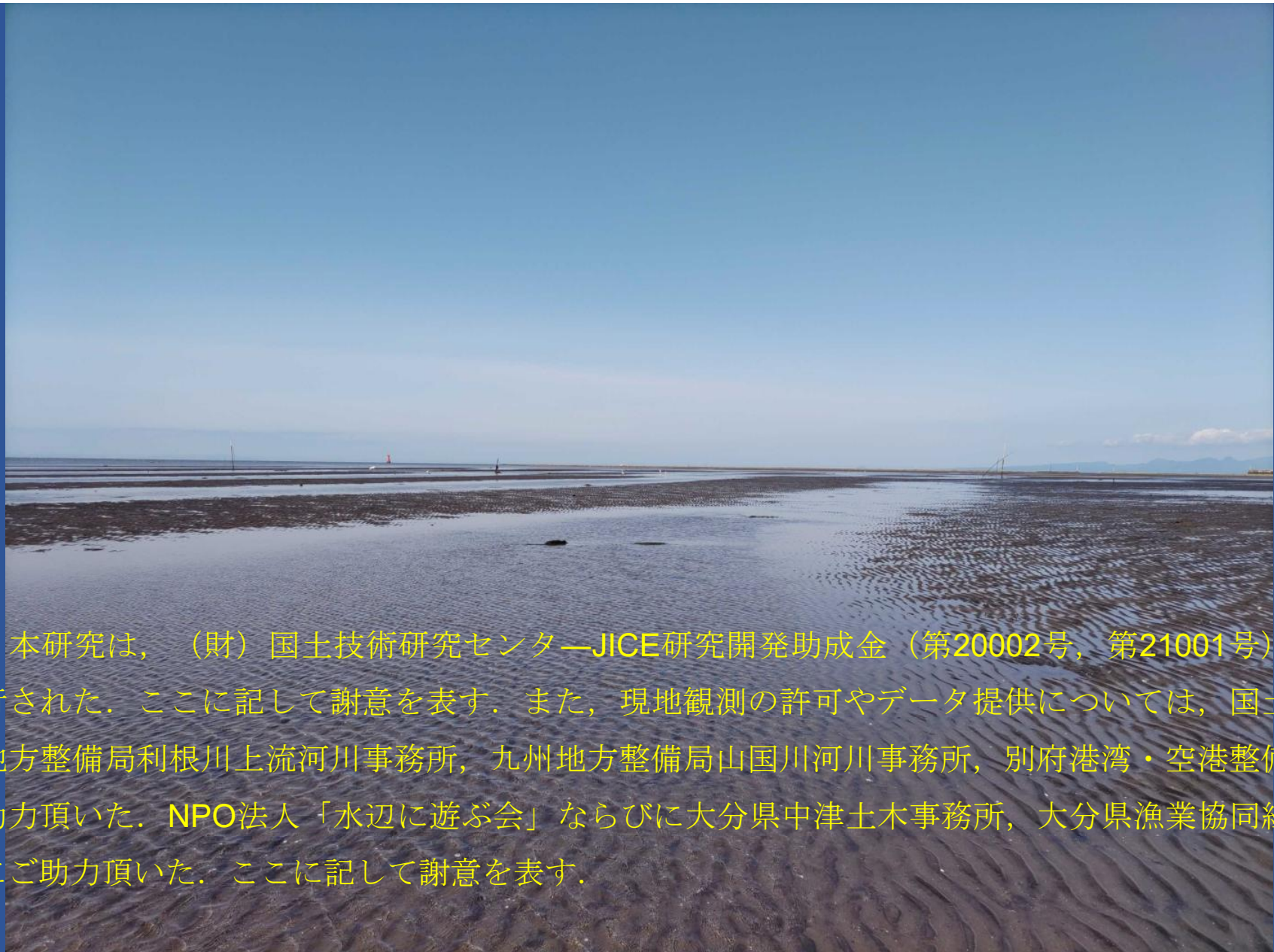
⑦山国川の年最大流量は右肩上がりであるが、gRSMによる過去の土砂供給量算定では、土砂供給量は年総降水量に依存する。年総降水量の最大値はさほど変化がないが、最小値が右肩上がりであり、トレンドとしては緩やかな右肩上がりである。年最大流量は最大降雨強度に依存すると考えられ、その急激な変化は2017年水害の頻発を示唆し、それに伴って大規模に斜面崩壊が起きると干潟の泥質化を招きかねない。出水による細粒化は、表層採泥のみの調査であるため、今後のトレンチ調査、柱状採泥調査結果の精査が必要である。

■利根川中上流域

⑤gRSMとtank modelによって吾妻川による流域土砂流出計算が行われ、ハイドログラフについては再現性が検証された。

⑥WDM-POMで河道計算を行い、4つの小流域を連結した流域土砂流出計算が行われた。

河道モデルを1次元不定流モデルの変更に山国川・中津干潟と利根川流域に現地適用し、現地観測データで妥当性検証を行った上で、現地の広域土砂対策の検討を行う。



謝辞：本研究は、（財）国土技術研究センターJICE研究開発助成金（第20002号，第21001号）によって遂行された。ここに記して謝意を表す。また，現地観測の許可やデータ提供については，国土交通省関東地方整備局利根川上流河川事務所，九州地方整備局山国川河川事務所，別府港湾・空港整備事務所にご助力頂いた。NPO法人「水辺に遊ぶ会」ならびに大分県中津土木事務所，大分県漁業協同組合にも同様にご助力頂いた。ここに記して謝意を表す。