Supplementary Information for
Poyang and Dongting Lakes, Yangtze River: tributary lakes blocked by main-stem aggradation

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Text S1: Area of large tributary blocked valley lakes (TBVL’s) around the world

In this section, we summarize the area of several of the world’s largest TBVL’s. Here we choose Poyang and Dongting Lakes in China, Lake Murray in Papua New Guinea, and Ria Lakes in Brazil for examples.

- Poyang and Dongting Lakes in the Yangtze River basin, China

Both these two lakes experienced significant shrinkage in area during the past century. The reasons for the shrinkage include levee construction, reclamation, sand dredging, which partially cut through the submerged step, water utilization, dam construction in both the tributary and the main stem, etc.

For Poyang Lake, the area was 5160 km² in 1954, but reduced to 3860 km² in 1998. From 2000 to 2010, the annual mean and minimum inundation areas showed statistically significant declining trends (~30.2 km²/year and ~23.9 km²/year, p<0.05), whereas no significant trend was found in the annual maximum inundation areas. From 2011 to 2016, both the annual maximum and minimum area showed an increasing trend due to the conservation efforts, with the annual minimum area increased by 60.51 km²/year.

For Dongting Lake, the area was 6200 km² in the 1860's, and then saw a decline in the 20th century with the area being 4905 km² in the 1930's, 3962 km² in the 1950's, 2960 km² in the 1970's, and 2623 km² in 1995. In the 2000's, the annual mean inundation area of Dongting Lake maintained a decreasing trend of 3.6%/year. Like Poyang Lake, Dongting Lake also revealed an increasing trend from 2011 to 2016, with annual minimum area increasing by 51.64 km²/year.

- Lake Murray in the Fly-Strickland River system, Papua New Guinea

The area of Lake Murray, the largest tributary blocked valley lake in Papua New Guinea, has been reported as 647 km².

- Ria Lakes in the Amazon River basin, Brazil

As for the Ria Lakes in the Amazon River basin, we did not find much information about their areas from the published literature. Therefore, we performed a remote-sensing analysis to obtain their areas. Here we study 7 large Ria Lakes: five of which are distributed along the Amazon River upstream of Manaus, where the Rio Negro joins the Amazon River, one is at the mouth of Tapajós, and one is at the mouth of Xingu. (See Figure S1 for the location of the 7 lakes.) Lake areas were obtained based on satellite images from Google Earth for each year during the period 2007-2016. Table S1 presents a summary of the results. It can be seen that the largest area we estimated for these Ria Lakes is 2225 km².

It should be noted that our calculation of the area of Lake 1 (see Figure S1) might include several small Ria Lakes (e.g., Lago Grande de Manacapuru and Lago Calabiana) as well as the tributary river channel. We regard them as one lake (i.e., Lake 1 in Table S1) with the consideration that the interconnection among these lakes could be enhanced during flood season. This might lead to an overestimation of lake area. For reference, the Lago Grande de Manacapuru in the upper part of Lake 1 is reported to reach ~511 km² during flood season.

Text S2: Morphodynamic formulation
In the section "Dimensionless rate of rise of main stem for tributary blocking" of the main text, we introduce a dimensionless formulation of the problem. Essential background is provided below.

The up-tributary backwater created by main-stem water surface rise is modeled using a standard 1D open channel formulation:

\[
\frac{\partial H}{\partial x} = S - S_1 \quad , \quad S_1 = \frac{C_r q_w^2}{g H^3} \quad , \quad Fr = \sqrt{\frac{q_w^2}{g H^3}} \quad (S1a, b, c)
\]

Here \(H\) = flow depth, \(x\) = downstream coordinate, \(S\) = streamwise bed slope = \(-\partial \eta / \partial x\), where \(\eta\) = bed elevation, \(q_w\) = flood discharge per unit width, \(g\) = gravitational acceleration, \(Fr\) = Froude number and \(C_r = (C_z)^2\) where \(C_z\) is a dimensionless Chezy coefficient of resistance. Equation (S1a) is solved subject to the downstream boundary condition

\[
\frac{\partial \xi}{\partial t}\bigg|_{x=L} = b
\]

where \(\xi = \eta + H\) = water surface elevation, \(t =\) time, \(x = L\) = downstream end of the modeled reach (where the tributary joins the main stem), and \(b\) = downstream rate of increase in water surface elevation.

The above formulation is coupled to morphodynamics by adding the Exner equation of bed sediment conservation:

\[
(1 - \lambda_p) \frac{\partial \eta}{\partial t} = -I_f \frac{\partial q_b}{\partial x} \quad (S3)
\]

where \(\lambda_p\) = porosity of bed sediment, \(q_b\) = volume transport rate of bed material per unit width, and \(I_f\) is a flood intermittency factor\(^\text{12}\). Equation (S3) is solved subject to the upstream boundary condition

\[
q_b\big|_{x=0} = q_{sf}
\]

where \(q_{sf}\) is the upstream sediment feed rate.

The transport rate of sediment is evaluated with a calibrated version of the Engelund-Hansen formulation\(^\text{13}\)

\[
q_b = \beta \frac{0.05 \sqrt{RgDD} (\tau^\ast)^{5/2}}{C_r q_w^2} \quad , \quad \tau^\ast = \frac{C_r q_w^2}{RgDH^2} \quad (S5a, b)
\]

where \(D\) is characteristic bed material grain size, \(R\) is the submerged specific gravity of the sediment \((1.65)\), \(\tau^\ast\) is the Shields number and \(\beta\) is an order-one calibration factor \((= 1.7; \text{see below})\). This relation has been verified for the middle Mississippi River, but with a different value \(\beta = 0.64\)\(^\text{12}\).

The above relations are placed in a non-dimensional context in the text of the paper.

**Text S3: Water surface rise of Yangtze River in response to the sea level rise**

In this section, we simulate the pattern of water surface rise of the Yangtze River in response to sea level rise. We implement the dimensional morphodynamic model described in the main text and immediately above. The simulation has the same settings as the calculational examples in the main text, except that \(\varphi = 1\) is applied here. This is because in this section we simulate the morphodynamics of the main stem, whereas in the main text we simulate the morphodynamics of a tributary. Details of the simulation parameters are presented in Table 1 in the main text.
The computational domain is specified as from the river mouth to 4000 river km upstream, which is sufficiently large to cover the locations of Poyang and Dongting Lakes. The duration of the calculation is 9000 years. At initial conditions, the river is in grade at normal flow. The sediment feed rate \( q_{sf} \) is kept constant during the simulation, and the water surface level at the downstream end rises with a constant rate of 10 mm/a to mimic Holocene sea level rise.\(^{14}\)

Figure S2 presents the predicted variation of water surface level at 600, 700, and 800 km upstream from the sea. Poyang Lake is about 700 km upstream of the river mouth. For all three sites, the water surface level of the river increases in response to sea level rise. The increase rates of water surface level at the three sites starts from around 7 mm/a in the first 1500 years, and eventually approaches 10 mm/a (the rate of sea level rise) at the end of 9000 years. The results indicate that for the Yangtze River, despite the fact that the increase rate of water surface level attenuates upstream of the river mouth, is still at least 7 mm/a 700 km upstream of the river mouth (which is near the location of Poyang Lake).

**Text S4: Calibration of the Engelund-Hansen relation against the Yangtze River data**

In this section the Engelund-Hansen relation\(^ {13}\) for sediment transport is calibrated against data for the Yangtze River. The calibrated version of the relation is formulated above as equation (S5). Here we apply the normal (steady and uniform) flow approximation for flow in the middle Yangtze River, so that \( H \) in equation (S1) can be calculated as the normal flow depth \( H_n \). (See equation (S6a) in the following section for the calculation of normal flow).

Data for daily water and sediment discharge of the Yangtze River at the Hankou hydrological station\(^ {15}\) (which is near the Wuhan city) is used for calibration. Data are shown in Figure S3. Here we implement the data of 2001 and 2002, so as to exclude the effects of Three Gorges Dam, which was impounded in 2003. Based on the bed and load sediment grain size distributions of the middle Yangtze River\(^ {16-17}\), the standard value of 62.5 \( \mu m \) is found to be applicable for the cutoff of wash load. The bed material load constitutes about 22.5% of the total sediment load of the middle Yangtze River. The daily discharge of bed material load at Hankou station during 2001 and 2002 can thus be obtained.

In the calibration, \( C_f, D, B, \) and \( S \) are specified as constants with the values presented in Table 1 in the main text. Using equation (S5) and the measured daily water discharge, we calculate the daily discharge of bed material load, and then sum over 2001 and 2002 to get the total volume of bed material load. The value of \( \beta \) is calibrated so that the calculated volume of bed material load equals the measured value for the period including 2001 and 2002. We eventually get a calibration factor of \( \beta = 1.70 \). Figure S4 shows how the calibrated version of Engelund-Hansen relation fits the daily measured data of 2001 and 2002.

**Text S5: Ratio of tributary sediment concentration to main-stem sediment concentration**

In this section, we study the ratio of tributary sediment concentration to main-stem sediment concentration \( c_{trib}/c_{main} \). Sediment concentration is calculated by the Engelund-Hansen relation calibrated against the Yangtze River data (see equation (S5)). In the calculation, flow depth is predicted using the normal flow condition. The same bed slope \( S \), sediment grain size \( D \), and dimensionless Chezy coefficient \( C_z \) (shown in Table 1 in the main text) are implemented for both the tributary and the main stem. The bankfull discharge per unit width near Wuhan (see Table 1 in the main text) is implemented for the main stem, whereas for the tributary a series value of \( \varphi < 1 \) are applied to set a range of tributary flow discharge per unit width.

Figure S5 presents the calculational results. It is seen that \( c_{trib}/c_{main} \) drops nonlinearly as \( \varphi \) decreases below unity, indicating that the tributary has a lower sediment concentration than the main stem. The value of \( c_{trib}/c_{main} \) becomes substantially lower as \( \varphi \to 0 \). For \( \varphi = 0.48 \) (as estimated in the main text for Poyang Lake), the sediment concentration of the tributary is about 61% of the
sediment concentration of the main stem, which contributes to the inability of the tributary to compete with the main stem as base level rises.

Text S6: Estimation of the flood intermittency factor

In this section, we estimate the flood intermittency factor of the middle Yangtze River based on measured data and the calibrated Engelund-Hansen relation. The flood intermittency factor $I_f$ is introduced so that the full hydrograph can be replaced by a characteristic constant discharge, for which we apply bankfull discharge. Under such assumption, the river is assumed to be at low flow and not transporting significant amounts of sediment for time fraction $1 - I_f$, and is at bankfull discharge and active morphodynamically for time fraction $I_f$.

We first calculate the transport rate of bed material load at bankfull discharge, using the Engelund-Hansen relation calibrated against the Yangtze River data (see equation (S5)). The values of $C_f$, $D$, $B$, and $S$ are specified as constant (see Table 1 in the main text for the values). For a given value of $I_f$, the annual bed material load can then be calculated. The value of $I_f$ is estimated by making the predicted annual bed material load equal to the long term measured data, which is 90.1 Mt/a as presented in Table 1 in the main text. With this method, we get an estimation of the flood intermittency factor $I_f = 0.73$.

Text S7: Calculational examples with different values of $\varphi$

In the main text, we present calculational results with a fixed value of $\varphi$ (= 0.48 based on the Yangtze and lumped Poyang feeder channel bankfull discharges) and different rates of base level rise $b$ (= 0, 2, 4.5, and 6 mm/a). In this section, we implement calculational examples with a fixed rate of base level rise $b$ (= 6 mm/a), but different values of $\varphi$ (= 0.3, 0.48, 0.7), to investigate the effect of tributary flow discharge on the formation of TBVL.

Figure S6 shows the calculational results. In the case $\varphi = 0.48$ (Figure S6b, which is identical to Figure 3d in the main text), the formation of a zone of ponded water is clear at the downstream end, where the tributary flows into the main stem. For a smaller value of $\varphi = 0.3$, the zone of ponded water is larger and deeper compared with that for $\varphi = 0.48$. Whereas for a larger value of $\varphi = 0.7$, the zone of ponded water barely forms at the downstream end. Here we provide an explanation: with a larger value of $\varphi$ (i.e., a larger tributary bankfull discharge per unit width), the sediment concentration of the tributary is larger (see Figure S5 for reference), so that the tributary is more able to keep up with the main stem as base level rises, and a TBVL is less likely to form.

Text S8: Dimensionless formulation for the formation of TBVL

In this section, we cast the dimensional model as presented in the main text in a dimensionless, and therefore quasi-universal form. To implement the dimensionless analysis, we first define the river equilibrium under normal flow conditions,

$$H_n = \left( \frac{C_f q_w^2}{g S_n} \right)^{1/3}, \quad Fr_n = \frac{q_w^2}{g H_n^3}$$  \hspace{1cm} (S6a,b)

where $H_n$ denotes the depth and $Fr_n$ denotes the Froude number, both under normal flow conditions. We further divide the bed elevation $\eta$ into the equilibrium bed elevation $\eta_n$ and the deviatoric bed elevation $\eta_d$:

$$\eta = \eta_n + \eta_d, \quad S_n = -\frac{\partial \eta_n}{\partial x}, \quad S = -\frac{\partial \eta}{\partial x} = S_n - \frac{\partial \eta_d}{\partial x}$$  \hspace{1cm} (S7a,b,c)
Based on the above-defined normal flow condition and river equilibrium, dimensionless parameters are defined below.

\[
\tilde{H} = \frac{H}{H_n}, \quad \tilde{\eta}_d = \frac{\eta_d}{H_n}, \quad \tilde{S} = \frac{\partial \tilde{\eta}_d}{\partial \tilde{x}} = \frac{S}{S_n} - 1, \quad \tilde{q}_s = \frac{q_s}{q_{sf}} \quad (S8a,b,c,d)
\]

\[
\tilde{x} = \frac{x S_n}{H_n}, \quad \tilde{t} = \frac{q_{sf} S_n}{(1 - \lambda_p) H_n^2} t \quad (S8e,f)
\]

In the equation for dimensionless time, \(H_n^2/S_n\) is a backwater volume (/unit channel width) scale, and \(q_{sf} H_t/(1 - \lambda_p)\) scales the volume input of sediment per unit width by time \(t\). The dimensionless time to blockage \(T_b\) is related to the dimensioned time to blockage \(t_b\) as

\[
T_b = \frac{t_b q_{sf} S_n}{(1 - \lambda_p) H_n^2} \quad (S9)
\]

Substituting equation (S8) into the backwater equation (equation (S1)), we obtain the dimensionless form of the backwater equation,

\[
\frac{\partial \tilde{H}}{\partial \tilde{x}} = \frac{1 - \frac{\partial \tilde{\eta}_d}{\partial \tilde{x}}}{1 - Fr_n^2 \tilde{H}^{-3}} \quad (S10)
\]

with the dimensionless form of the downstream boundary condition, given as (from equation (S2))

\[
\left. \frac{\partial \left( \tilde{\eta}_d + \tilde{H} \right) }{\partial \tilde{t}} \right|_{\tilde{x} = \tilde{L}} = BL, \quad \tilde{L} = \frac{S_n L}{H_n}, \quad BL = \frac{(1 - \lambda_p) b H_n}{q_{sf} S_n t} \quad (S11a,b,c)
\]

Here \(\tilde{L}\) is dimensionless domain length and \(BL\) is the dimensionless Blocking Number, which corresponds to the dimensionless increase rate of the water surface level of the main stem.

Substituting equation (S8) into the Exner equation (S3), we obtain the dimensionless form of the Exner equation,

\[
\frac{\partial \tilde{\eta}_d}{\partial \tilde{t}} = - \frac{\partial \tilde{q}_s}{\partial \tilde{x}} \quad (S12)
\]

with the dimensionless form of the upstream boundary condition (equation (S4)) given as,

\[
\tilde{q}_s \bigg|_{\tilde{x} = 0} = 1 \quad (S13)
\]

Substituting equation (S8) into the calibrated Engelund-Hansen relation, the dimensionless sediment transport rate \(\tilde{q}_s\) can be calculated as,

\[
\tilde{q}_s = \frac{q_s}{q_{sf}} = \left( \frac{\tau}{\tau_n} \right)^{2.5} \tilde{H}^{-5} \quad (S14)
\]

In summary, in the dimensionless formulation equations (S10) and (S12) are the governing equations, equations (S11a) and (S13) are the boundary conditions, and equation (S14) is the closure relation. When applying the dimensionless formulation to study the formation of TBVL, we specify the initial condition as the equilibrium river channel under normal flow conditions; that is,
\[ \tilde{H}|_{t=0} = 1, \quad \tilde{n}_d|_{t=0} = 0 \]  

(S15)

It should be noted that in the dimensionless formulation, there are only two parameters that need to be specified, i.e. \( F_r \) and \( Bl \). Therefore, in the main text we implement various values of \( F_r \) and \( Bl \) in the calculation to study the formation of TBVL under different scenarios. For each case, we implement a calculational domain of \( L = 10 \) and a dimensionless calculational duration of 50. For the case of Poyang Lake with \( \phi = 0.48 \), \( \tilde{t} = 50 \) corresponds to a time duration of \( \sim 40,000 \) years, and is much larger than the duration of Holocene sea level rise.

In the dimensionless analysis, the following condition is implemented to determine the dimensionless time \( T_b \) for incipient of blocking

\[ \frac{\partial S}{\partial x}|_{x=L} = 0.01 \]  

(S16)

It should be noted that equation (S16) can be regarded as the dimensionless form of equation (4) in the main text, except that a small number of 0.01 is applied to the right hand side of equation (S16), instead of 0 in the right hand side of equation (4) in the main text. Such a treatment is purely numerical in order to avoid the influence of numerical instability on the calculation of \( T_b \). Using equation (S8), equation (S16) can be written in the dimensional form

\[ \frac{\partial S}{\partial x}|_{x=L} = 0.01 \frac{S_n^g}{H_n^g} \]  

(S17)

Considering that \( S_n << 1 \) for most alluvial rivers, equation (S16) can be regarded as nearly identical to equation (4) in the main text.

Text S9: Dependence of Time to Incipient Lake Formation on Spatial Scale: Froude Scaling

The relations (5a,b,c) of the main text can be manipulated into the form

\[ \text{An} = F_{\text{new}}(\text{Bl}) \quad ; \quad \text{An} = \text{BIT}_b = \frac{t_b^b}{H_n^b} \quad ; \quad F_{\text{new}} = \text{BI}F^{-1}(\text{Bl}) \]  

(S18a, b, c)

We apply the principles of Froude scale modeling to study the dependence of time to blocking on spatial scale. We first consider a "prototype" case with values \( t_{bp} \), \( b_p \), \( H_{np} \), \( q_{sp} \) and \( S_{np} \). We keep porosity \( \lambda_p \) and flood intermittency \( I_f \) constant. We then decrease all spatial parameters to model parameters (subscript "m") by the same multiplicative factor \( \lambda_m \):

\[ H_{nm} = \lambda_m H_{np} \quad ; \quad S_{nm} = \frac{\lambda_m}{\lambda_n} S_{np} = S_{np} \]  

(S19a,b)

The second relation above states that there is no scale distortion. We require that the Froude number in the model be equal to that in the prototype:

\[ F_{rnm} = \frac{U_{nm}}{\sqrt{gH_{nm}}} = F_{rnp} = \frac{U_{np}}{\sqrt{gH_{np}}} \]  

(S20)

where \( U_n \) is the flow velocity at normal flow. Rearranging with the above two flows gives the velocity scaling:
\[ \frac{U_{nm}}{U_{np}} = \sqrt{\frac{\lambda_s}{\lambda_p}} = \frac{b_m}{b_p} \tag{S21} \]

Time scales as \( \left( \frac{H_{nm}}{H_{np}} \right) / \left( \frac{U_{nm}}{U_{np}} \right) = (\lambda_s)^{1/2} \), in which case \( q_{sfm}/q_{sfp} = (\lambda_s)^{3/2} \). Thus the Blocking Number \( Bl \) is the same in the model as the prototype.

\[ Bl_m = \frac{(1 - \lambda_p)b_m H_{nm}}{\lambda_s q_{sfm} S_{nm} I_f} = \frac{(1 - \lambda_p)\lambda_s^{1/2} b_s \lambda_s H_{np}}{\lambda_s^{3/2} q_{sfp} S_{np} I_f} = Bl_p \tag{S22} \]

Substituting into (S18), model and prototype times to blockage \( t_{bm} \) and \( t_{bp} \) are related as

\[ An_m = \frac{t_{bm} b_m}{H_{nm}} = \frac{t_{bm} \lambda_s^{1/2} b_p}{\lambda_s H_{np}} = An_p = \frac{t_{bp} b_p}{H_{np}} \tag{S23} \]

or therefore

\[ t_{bm} = \lambda_s^{1/2} t_{bp} \tag{S24} \]

Consider the case of Figure 3c. The time for incipient formation of a blocked valley lake is around 9000 years. When scaled down by a spatial factor \( \lambda_s = 1/30 \) (corresponding to a main-stem depth reduced from 21 m to 0.7 m), this time becomes around 1640 years. Both the time for formation of TBVL, as well as the time for their erasure by deposition after sea level stops rising, decline with declining scale.

**Text S10: Case of Sediment Mixtures**

In the main text, we have assumed a uniform bed grain size. In this section, we investigate how the nonuniformity of bed sediment can affect the formation of TBVL. To do this, we run several cases with sediment mixtures. The model implemented in this section is different from the model applied in other sections in that: (1) the Exner equation with an active layer formulation and a substrate storage method is applied to the conservation of sediment mixtures, and (2) a grain size-specific Engelund-Hansen type (SEH) relation is applied to calculate the fractional bed material load of sediment mixtures\(^{19} \).

The values of several key parameters in the model are as follows. The thickness of the active layer is specified as 2.57 m (i.e., 20% of the normal flow depth\(^{19} \)). The adjustment coefficient of the SHE relation is 2.3, based on the calibration against the measured annual bed material load at Hankou station (Table 1 of the main text). The grain size distribution (GSD) of the initial bed surface is based on the measurement near Hankou Station on the Yangtze River, before the implementation of the Three Gorges Dam (Figure S7). The rate and GSD of sediment supply are calculated by the SEH relation, under an assumed equilibrium condition. The initial substrate material is specified as a mixture of the initial bed surface material and sediment supply, with the fraction of each component of 50%. It should be noted that the GSD of the substrate is not important in this study, as the system we simulate is dominated by deposition. The values of other parameters are the same as those applied in the simulation of the main text. The measured grain size distribution, and a renormalized distribution to remove material finer than 62 \( \mu \text{m} \), as required by SEH, are shown in Figure S7.

Figure S8 shows the simulation results for sediment mixtures with four different rates of base level rise: 0, 2, 4.5, and 6 mm/a, which are the same as those applied in Figure 3 of the main text. Comparison between Figure S8 and Figure 3 of the main text shows that the consideration of multiple sediment sizes leads to very little change to the calculational results.

**Text S11: Ancient Tributary Blocked Valley Lakes**
All but one of the TBVL’s referred to in the main text (Yangtze River, China; Amazon River, Brazil; Fly River, Papua New Guinea) are present in the modern. It is reasonable to assume that many more such lakes existed following any given glacial period, but have since filled with sediment.

Examples of ancient TBVL’s can be found in southern Illinois, in valleys of rivers that drained sub-lobes of the Laurentide Ice Sheet during Illinois Episode Glaciations (MIS6) and Wisconsin Episode Glaciation (MIS2-4) (Figure S9). These glacial sediment-rich rivers were tributaries to the Wabash, Ohio, and Mississippi Rivers, and are today filled with lacustrine sediment associated with these recent glaciations. Tributary damming and subsequent lake formation have been interpreted to be a result of aggradation of the respective mainstem rivers. This aggradation may have been driven by upstream-propagating sea level rise, or alternatively by the deposition of outwash sediment in the main stem. In particular, both the Kaskaskia and Embarras River Valleys, Illinois, contain significant slackwater lake deposits from both MIS6 and MIS2-4, indicating that this phenomenon is not unique to the Wisconsin Episode, and suggesting that ancient TBVL’s are common in the geologic record.
Figure S1. Locations of the 7 Ria Lakes in the Amazon basin.
Figure S2. Calculational results for the Yangtze River water surface rise in response to sea level rise. The solid lines denote the cumulative increase of water surface level at three sites. The dashed lines denote the corresponding increase rate of water surface level. For reference, the (average) rate of sea level rise is presented as the black dashed line.
Figure S3. Daily water and sediment discharge of the Yangtze River at Hankou hydrological station during a. 2001 and b. 2002. The black dashed line denotes the bankfull water discharge near Hankou hydrological station. 
Figure S4. Daily discharge of water and bed material load at Hankou hydrological station during 2001 and 2002. The Engelund-Hansen relation is calibrated against the measured data.
Figure S5. Relation between \( \frac{c_{\text{trib}}}{c_{\text{main}}} \) and \( \phi \). \( \frac{c_{\text{trib}}}{c_{\text{main}}} \) = ratio of tributary sediment concentration to main-stem sediment concentration; \( \phi \) = ratio of tributary flow discharge per unit width to main-stem flow discharge per unit width.
Figure S6. Calculational results for three cases of tributary bankfull discharge per unit width: a. \( \varphi = 0.3 \); b. \( \varphi = 0.48 \); c. \( \varphi = 0.7 \). Rate of base level rise \( b \) is fixed at 6 mm/a for the three cases.
Figure S7. Grain size distributions of bed material. The black solid line shows a field measurement near the Hankou Station, Yangtze River, in 2001. The red dash line shows the GSD applied in this study, after renormalization so as to be consistent with the assumption of a cutoff size 62.5 μm for wash load, as required by the SHE model.
Figure S8. Calculational results for an initial bed sediment mixture (Figure S7), with four different rates of base level rise: a. b = 0 mm/a; b. b = 2 mm/a; c. b = 4.5 mm/a; and d. b = 6 mm/a.
Figure S9 The figure shows candidates for ancient TBVL’s, which were present at some time after the end of a glacial period, but which have subsequently filled with sediment\textsuperscript{20-21}.
Table S1. Areas of the 5 Ria Lakes in the Amazon basin during 2007-2016.

| No. | Maximum Area | Minimum Area | Average Area |
|-----|--------------|--------------|--------------|
| 1   | 793 km²      | 761 km²      | 775 km²      |
| 2   | 275 km²      | 241 km²      | 260 km²      |
| 3   | 255 km²      | 248 km²      | 251 km²      |
| 4   | 926 km²      | 879 km²      | 900 km²      |
| 5   | 388 km²      | 356 km²      | 368 km²      |
| 6   | 2225 km²     | 2112 km²     | 2143 km²     |
| 7   | 1176 km²     | 1140 km²     | 1165 km²     |
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