SEDIMENT AND CONTAMINANT
TRANSPORT AND FATE IN RIVERS

by

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ABSTRACT

Contaminated bottom sediments in a river can be a major source of contaminants to the overlying water and biota. In the present paper, recent work on understanding, modeling, and predicting the transport and fate of these bottom sediments and the contaminants associated with them is described. Physical processes that are significant in the accurate determination of the hydrodynamics and sediment dynamics are first discussed. Applications of the resulting numerical models to the Fox River in Wisconsin, the Saginaw River in Michigan, and the Buffalo River in New York are then presented. Significant sediment transport mechanisms in these applications are suspended load, bed load, and the movement of sediments due to slumping. A brief discussion of contaminant fluxes between the bottom sediments and the overlying water is also given.
INTRODUCTION

Contaminated bottom sediments are becoming widely recognized as a major problem in the Great Lakes area as well as elsewhere. Depending on local conditions, which can vary spatially and with time, these sediments can be a major source of contaminants to the overlying water and biota. In order to remediate this problem of contaminated bottom sediments and to evaluate possible management alternatives for the disposition of these sediments, extensive experimental and field work has been done to determine the transport properties (resuspension, flocculation, and settling speeds) and chemical sorption properties of fine-grained sediments; we have then used these results to develop numerical models of the transport and fate of sediments and contaminants in rivers, lakes, estuaries, and near-shore areas of the oceans. In the present paper, the emphasis is on the transport and fate of sediments and contaminants in rivers.

In order to accurately predict the transport and fate of sediments in rivers, it is necessary to have a quantitative and detailed knowledge of (1) the hydrodynamics, including flow rates, currents as a function of these flow rates, seiche effects due to changing lake levels, wave action, and especially bottom shear stresses due to all of the above hydrodynamic processes, and (2) sediment dynamics, including the resuspension properties of the bottom sediments, the settling speeds of the suspended sediments, and the incoming sediment load. If the above processes are understood, quantitative and predictive models of the transport and fate of sediments can then be constructed. These models must include the transport of sediments as suspended load and as bed load as well as their movement due to slumping.

Recent work on understanding and modeling the above processes will be briefly summarized in the following section. Representative applications of the resultant models to the Fox River in Wisconsin, the Saginaw River in Michigan, and the Buffalo River in New York will then be described. The physical properties of each of these rivers is distinctly different from the others, and therefore the transport and fate of sediments and
contaminants also differs significantly from one river to the next. In order to be relatively brief, only a few applications will be given. These applications have been chosen to illustrate interesting and significant processes in sediment transport and also to illustrate our present ability to model these processes. This work on the transport and fate of sediments is presently being extended to include the transport and fate of hydrophobic contaminants associated with these sediments. A brief discussion of contaminant fluxes between the bottom sediments and the overlying water is given in the section after the sediment transport modeling applications. A summary and concluding remarks are presented in the final section.

PHYSICAL PROCESSES AND PARAMETERS

A detailed knowledge of the hydrodynamics of a river is necessary in order to accurately predict the transport and fate of sediments and contaminants. Of fundamental importance are the flow rates as a function of time, especially during high flows where most of the sediment and contaminant transport occurs. The significance of high flows will be discussed further below.

The currents are of course dependent on the flow rates and can be determined once the flow rates are known. This can be done by means of numerical models. The simplest models are quasi-one-dimensional, either steady or time-dependent. However, the depth of most rivers changes rapidly in the direction across the river, and this significantly affects sediment resuspension and deposition. Because of this and because one-dimensional models do not take this variation into account, one-dimensional models are generally inadequate for the accurate prediction of sediment transport.

In our recent work (Ziegler and Lick, 1988; Gailani et al, 1991), we have mainly used a vertically integrated, two-dimensional, time-dependent hydrodynamic model. This model is more accurate than a one-dimensional model, but is simpler than a three-dimensional model, and is a valid approximation for shallow waters when the horizontal velocities and suspended sediment concentrations are approximately independent of depth,
i.e., when there is almost complete mixing of the water column in the vertical direction. For the Fox, Saginaw, and Buffalo Rivers (which are generally less than 10 m deep), this is almost always true, with the exceptions being during a few low flow, highly stratified events.

For the low-flow, stratified conditions, a three-dimensional, time-dependent model is necessary. These models have been applied to many rivers, lakes, and oceans (e.g., see Sheng and Lick, 1978; Blumberg and Mellor, 1980; Paul and Lick, 1985) but are inherently complex and consume large amounts of computer time. Because significant stratification generally occurs only during low flows, the effect of stratification on the net sediment transport over the entire year is generally not significant. However, during low-flow periods, stratification does influence the transport of sediments and contaminants and hence influences water quality during this time. For these latter problems, a three-dimensional, time-dependent model is necessary. Applications of both the two-dimensional and three-dimensional models will be illustrated in the following sections.

During storm conditions, effects due to changing lake levels (seiches) can be significant in modifying the currents and hence resuspension in the river. The main effect is on the currents and sediment resuspension in the river near its mouth. Due to the very nonlinear dependence of resuspension on flow (to be discussed below), the effect of seiches on resuspension is proportionately much larger than its effect on current magnitude. For example in the Fox, changes in resuspension by more than a factor of two due to seiches have been calculated, while changes in current magnitudes are only about 25 per cent.

Under certain conditions, the effects of waves on resuspension can also be significant. Generally, waves are not important since in a winding river the fetch is small and waves can not build up as they would on a lake. However, for a straight river with an appreciable fetch or if waves generated in the lake propagate in such a direction that they can enter the river without appreciable reflection or refraction, then the effects of waves on resuspension may be significant.
Once the hydrodynamics and especially the bottom shear stresses are known, the resuspension of the bottom sediments as a function of this bottom stress must be determined. The emphasis here is on fine-grained sediments because of their relatively large adsorptive capacity for contaminants compared to coarse-grained sediments. It has been demonstrated that resuspension rates of sediments are significantly affected by particle size variations and also by cohesion between particles. In particular, it has been shown that for fine-grained, cohesive sediments at any particular stress, only a finite and relatively small amount of sediment can be resuspended as opposed to noncohesive, uniform-size, coarse-grain sediments, which have a uniform rate of resuspension (Tsai and Lick, 1987; MacIntyre et al, 1989).

For fine-grained sediments, experimental work (Partheniades, 1972; Mehta, 1973; MacIntyre et al, 1989; Lick and Kang, 1987; Tsai and Lick, 1987; Xu, 1991) has determined the dependence of the resuspension rate and the total amount of sediment $e$ that can be resuspended at a particular stress as a function of (a) the turbulent stress at the sediment-water interface, and (b) the water content of the deposited sediments (or the time after deposition) for various sediments from both lakes and oceans. A formula for $e$ which approximates the experimental data can be written as

$$e = \frac{a}{t_d^m} [\tau - \tau_0]^m$$

for $\tau > \tau_0$

$$= 0$$

for $\tau < \tau_0$  \hspace{1cm} (1)

where $e$ is the net amount of sediment resuspended per unit surface area in gm/cm$^2$, $a$ is a parameter which depends on the sediment, $t_d$ is the time after deposition in days, $m$ is a constant approximately equal to three, $n$ is a constant usually between one and two, $\tau$ is the shear stress (dynes/cm$^2$) produced by wave action and currents, and $\tau_0$ is an effective critical stress which varies from approximately 0.1 dynes/cm$^2$ for freshly deposited sediments to approximately 1 dyne/cm$^2$ for $t_d$ greater than two days. Each of the parameters $\tau_0$, $a$, $m$, and $n$ is dependent on the particular sediment (and the effects of benthic organisms) and needs to be determined experimentally. Experimental results,
summarized by the above equation, have only been obtained for a very limited range of parameters and conditions. In particular, $\varepsilon$ has not been determined at the high shear stresses often encountered when there are strong currents due to large floods. Nevertheless, Eq. (1) is a valid description of existing experimental data and is useful as a first approximation to describe the variation of $\varepsilon$ in many realistic situations.

The above formula is for the net resuspension. The total amount of sediment is not resuspended instantaneously but over a period of time on the order of an hour. In numerical computations, a reasonable approximation to the resuspension rate is that it is constant and equal to its initial value until all available sediment is resuspended and is then zero until further sediment is deposited and is available for resuspension.

In our present modelling of resuspension and sediment bed dynamics, the sediment bed is assumed to consist of layers in the vertical direction; the properties of each layer depend on time after deposition and composition (relative fractions of medium and coarse sediments) and are allowed to vary in the horizontal direction. An arbitrary number of layers, their properties, and their thicknesses can be defined initially. Usually 11 layers are assumed. Typically, the top layer is assumed to be newly deposited sediment less than three hours old. It has a very high water content and hence a very low critical shear stress ($\tau_0 = 0.1 \text{ dynes/cm}^2$), and is easily resuspended. Below this fresh layer are three, six, and twelve hour old layers with critical shear stresses increasing with age. Below these layers are 7 layers one to seven days old with age increasing with depth. For these layers, it is assumed that $\tau_0 = 1.0 \text{ dyne/cm}^2$. All layers compact with time with $\varepsilon$ for each layer given by Eq. (1) above.

The program monitors changes in the sediment bed for each cell in the river model. Deposited material is added to and sediment is resuspended from the surface layer. The material in this surface layer is moved into the three hour old layer every three hours, the three hour old material is moved into the six hour old layer every six hours, and so forth. For a given cell in the model, any or all layers may be empty. Frequently the bottom layer (seven days old) is the only layer containing sediment, and therefore it is the surface layer.
The net sediment flux is then given by the difference between the resuspension rate and the deposition rate, where deposition is assumed to be due to settling of the suspended particles.

Once particles are resuspended, their transport is determined by the hydrodynamics and by their settling speeds. These settling speeds are dependent upon the sizes and densities of the particles. However, because of the cohesive nature of fine-grained sediments, the basic sedimentary particles (a large fraction of which are microns in size) aggregate together to form much larger particles or floes. These floes may be microns to more than a centimeter in size. In addition to being larger, these floes have a much lower density than the primary particles making up the floc. Because of this, the settling speeds of the floes are much different from the settling speeds of the primary particles. The aggregation and disaggregation of these floes occur continually and continually modify the sizes, densities, and settling speeds of the floes. In order to investigate the rates of aggregation and disaggregation and the parameters on which these processes depend, flocculation experiments have been done by us (Tsai et al, 1987; Burban et al, 1989). From these experiments, a quantitative description of flocculation is being obtained, and a reasonably general numerical model of the dynamics of flocculation is being developed (Lick and Lick, 1988; Lick et al, 1992). Measurements have also been made of the settling speeds of the floes produced in these experiments (Burban et al, 1990).

In addition to the transport of suspended load, bed load can also be significant. Bed load transport is defined as the motion of particles that either roll or saltate along the river bottom without ever being brought into resuspension. In the present analyses, bed load equations developed by Rijn (1984a) have been used. Rijn bases his equation for the transport rate, \( Q_b \), on the particle sizes and densities and on particle trajectory equations. For grains in the size range from 200 to 2000 \( \mu \text{m} \), the resulting bed load equation can be written as

\[
Q_b = \frac{0.053(s-1)g^{0.5} D_5^{1.5} T^{2.1}}{D_s^{0.3}}
\]  
(2)
where

\[ T = \frac{(u^*)^2 - u_{cr}^2}{u_{cr}^2} \]

\[ D* = D_{50}[(s - 1) \frac{g}{v^2}]^{1/3} \]

\[ u^* = \frac{ug^{0.5}}{C'} \]

\[ C' = 1810 \log_{10} \frac{12R_b}{3D_{90}} \]

\( Q_b \rho_s \) (gm/m s) is the mass in grams of sediment crossing a one meter width of sediment bed per second, where \( \rho_s \) is the sediment density. In these equations, \( D_{50} \) is the median grain size, \( s \) is the ratio of sediment density to water density, \( u^* \) is the bed shear velocity related to grains, \( g \) is the acceleration of gravity, \( C' \) is the Chezy coefficient related to grain roughness, \( R_b \) is the hydraulic radius (ratio of the cross sectional area to wetted perimeter), \( v \) is the kinematic viscosity, \( D_{90} \) is the ninetieth percentile particle size, and \( u \) is the mean velocity of the overlying water. The critical bed shear velocity \( u_{cr} \), below which no bed load movement occurs, is determined from the Shields curve and is a function of the mean particle size and sediment density. The parameter \( T \) is the transport stage parameter; if the value is negative, no bed load movement occurs. The value of \( C' \) is dependent on the channel geometry and maximum grain size. For many rivers, the value of \( C' \) is approximately 50 to 75 (Chow, 1959; Rijn, 1984b). In our work, we have chosen \( C' = 60 \).

**FOX RIVER**

As a first application of the above ideas and models, consider the Fox River in Wisconsin. The Fox is 56 km long and runs from Lake Winnebago in the south to Green Bay in the north. The valley through which the Fox runs is heavily industrialized and
contains large concentrations of pulp and paper industries. Because of this, the waters and sediments in the Fox are heavily polluted. For this river, work on the modeling of suspended load (Gailani et al, 1991) as well as bed load (Gailani, 1991) has been done and will be briefly summarized here. In these studies, the particular concern was with the lower Fox River which extends to Green Bay from a dam at DePere, 11 km upstream from the mouth of the river.

A bathymetric map of the lower Fox River (Figure 1) shows that the upstream portion (defined as upstream of the Ft. Howard Paper Company) is wide with many shallow areas less than two meters in depth. This portion is no longer dredged, but previous dredging has established a channel up to five meters deep which is still present today. The bottom sediments in this upstream portion are essentially all fine-grained, cohesive sediments, predominantly silts and clays. The river narrows near its midpoint at Fort Howard and remains narrow down to its mouth at Green Bay. The U.S. Army Corp of Engineers dredges this narrow portion as needed to allow large ship passage. Sediments in the deeper channels of this downstream area are generally coarse-grained and sandy. In the shallow, near-shore areas, the sediments are again fine-grained and cohesive. The only significant tributary to the lower Fox River is the East River which joins the Fox approximately 2 km upstream from the mouth. The junction of the two rivers has been widened and dredged to form a turning basin for large ships. The flow in the East River has been estimated to be approximately ten percent of the flow in the Fox River (a median flow of 105 m$^3$/s). The flow in the Fox is controlled primarily by two dams upstream of the lower Fox. For a forty-five year period from 1940 to 1985, the maximum flow was 668 m$^3$/s while the low flows were about 20 m$^3$/s. This is a relatively small range of flows. Because of the control, extreme flow variations are now less than before control and less than in most other comparable rivers without control.

In our calculations for the Fox, we used laboratory and field measurements to determine the resuspension parameters; a quasi-equilibrium model of flocculation for simplicity; settling speeds as measured in the laboratory; a two-dimensional (vertically-integrated), time-dependent, hydrodynamic and sediment transport model; an SMB model
of wave action when waves were significant; a nine-layered sediment bed model with properties based on our experimental work; suspended solids concentrations measured at the DePere Dam (the upstream boundary) as input; and suspended solids concentrations at the river mouth at Green Bay (the downstream boundary) as verification.

For purposes of quantitatively understanding some of the main features of the flows and the effects of flow rate, several calculations of steady-state flows were also made. One such calculation was for a 99.7 percentile flow of 280 m$^3$/s. The suspended solids concentration at the Dam was assumed to be 75 mg/l, a typical concentration for a flow of this magnitude. Currents are shown in Figure 2. Calculated results for erosion and deposition are shown in Figure 3. It can be seen that large parts of the river show erosion while deposition is more confined to the near-shore, shallow regions.

For purposes of calibration and verification, several time varying flow events were modeled; one was from May 22, 1989 to June 20, 1989. For this event, the flow rate and measured sediment concentrations at DePere Dam and at the mouth of the river at Green Bay are shown in Figure 4. Results of calculations for the suspended solids concentrations from May 22, 1989 to June 20, 1989 are shown and are compared with the observed concentrations in Figure 5. It can be seen that good agreement between the two has been obtained. Calculations for other storms show similar good agreement between calculations and observations.

Although these calculations served to verify the transport of sediment as suspended load, the problem of predicting changes in bathymetry still remained. By comparing measured changes in bathymetry with calculated changes due to resuspension/deposition, it became clear that resuspension/deposition contributed significantly to these changes but that bed load was also important. Because of this, bed load transport (as described in the previous section) was also included in the analysis.

For purposes of verification, depth measurements were made on the Fox at eleven transects on October 27, 1989 and again on September 20, 1990. The measurements were made by Jeffrey Steuer of the USGS. The eleven month time period between transects included one storm with flow rates greater than 400 m$^3$/s (a once in five year storm) similar
to the May 1989 storm, two events with flow rates over 300 m$^3$/s, as well as a more modest storm (maximum flow rate 187 m$^3$/s). The EPA stopped measuring daily averaged suspended solids concentrations on April 30, 1990, so this input data was not available for the latter part of the period. However, by using data from previous storms, the concentrations for storms during this latter part of the period were estimated.

Using available flow rate, lake level, and concentration data, the sediment transport during the eleven month period between transects was then modeled. The total erosion for this period is shown in Figure 6a and the total deposition in Figure 6b. Contour intervals are in units of gm/cm$^2$, where 1 gm/cm$^2$ is roughly equal to 2 cm of either erosion or deposition. As would be expected, most erosion occurs below Fort Howard with up to 50 gm/cm$^2$ being eroded. Erosion is less in the area just downstream of the East River turning basin than in areas where fine-grained sediments dominate because of the assumption of more coarse-grained sands in this area. This area displays about two thirds of the erosion that would have occurred if uniform grain size were assumed throughout the river with a maximum erosion of about 40 gm/cm$^2$. Most deposition is outside the channel with the few exceptions of pockets of bed load deposition as high as 30 gm/cm$^2$ in the channel; but it should be emphasized that these pockets are small and few in number. Net deposition is generally less than 10 gm/cm$^2$ except near the dam where it is as high as 20 gm/cm$^2$.

The actual erosion/deposition for the eleven month period, as determined from the transects, was then compared to the calculated changes. For each transect, a comparison of the maximum bed changes occurring in the channel is presented in Table 1. In general, it can be seen that both resuspension/deposition and bed load contribute significantly to changes in bathymetry. It can also be seen that generally good agreement between the calculated and observed changes has been obtained. Exceptions are at transect 5 (with -182 cm observed and only -35 cm predicted) and transect 2 (with -55 cm observed and only -1 cm predicted). At all other transects, the predicted and observed changes are in reasonable agreement.

Transect 5 is just below the Fort Howard turning basin. At this transect, the river is
330 m wide with a wide channel up to 8 m deep. The maximum measured erosion (182 cm) is three times greater than at any other transect, but is confined to a gash 20 m wide. The gash is inconsistent with the rest of the contour, as it has steep walls and is strongly suggestive of dredging. According to the Army Corp of Engineers, no dredging occurred in this area during the period between transect measurements. However, a large cargo ship ran aground in the vicinity, and maneuvering to free the ship caused a great deal of disturbance in the sediment bed probably causing this gash. It is difficult to envision erosion of this magnitude occurring naturally. Transect measurements made in April 1991 show this gash to be filling in rapidly as already 100 cm of sediment had deposited. If the gash is ignored, the remainder of the channel eroded about 8 cm. By comparison, the model calculates 35 cm erosion.

Transect 2 spans a wide portion of the river measuring 670 m across with a maximum depth of 4.5 m. The measured erosion was 55 cm during the period, a surprisingly large amount considering the large cross-sectional area of the segment. In the area near transect 2, the bathymetry is changing very rapidly with a narrow and deep channel upstream decreasing rapidly to a shallow and flat bathymetry downstream. Small errors in navigation during the bathymetric measurements could easily cause the differences shown in Table 1.

SAGINAW RIVER

Four tributaries form the Saginaw River. They are the Cass, Flint, Titabawassee, and Shiawassee Rivers. The Saginaw River winds through the cities of Saginaw and Bay City, Michigan for 35 km. There are no controls on the river or any of the tributaries. The river is used as a supply line for many heavy industries which line the shores. For this reason, the U.S. Army Corps of Engineers dredges the river channel to allow large ships to pass. The project depth limit is 25 feet (7.6 m), but the actual depth often exceeds this value, being as deep as 30 feet (9.1 m) or more in some areas at some times. There are shallow areas near shore, with minimum depths of 1 m or less, except where turning
basins are located. Here, the river is dredged nearly side-to-side. The average width of the river is approximately 250 m, with a maximum width of 600 m.

Since there are no controls on the Saginaw, the flow variance is large compared to a controlled river like the Fox. For a 48 year period from 1940 to 1989, the maximum flow on the Saginaw River was 1910 m$^3$/s, while the minimum flow was 7 m$^3$/s (maximum velocity less than 1 cm/s). These flows range over almost three orders of magnitude. In comparison, the range of flows on the Fox River is just a little over one order of magnitude. The 50th percentile flow on the Saginaw River is 57 m$^3$/s, creating a maximum velocity of 7 cm/s. The 90th percentile flow is only 250 m$^3$/s (maximum velocity of 30 cm/s). However, the 99.7th percentile flow is 982 m$^3$/s, creating a maximum velocity of 107 cm/s. It is clear that the river is slow flowing for most of the year, with large peaks in the flow rate occurring only a few times each year during storms.

In the present investigation, the area of concern is from Middle Ground Island to the mouth of the river at Saginaw Bay. The bathymetry for this part of the river is shown in Figure 7. Calculations of sediment resuspension, deposition, and transport are presently being made in a similar manner to that for the Fox. However, for the Saginaw River, a curvilinear grid is being used in the calculations. Of particular interest in the calculations is the period from May 20, 1991 to May 13, 1992 (during which time bathymetric measurements at 11 transects (see Figure 7) were made approximately every four months) and 1986 (during which year the largest flood on record for the Saginaw occurred). These calculations show that, as for the Fox, bed load and resuspension/deposition of suspended load are significant factors in changing the bathymetry. However, another physical process is also important and is as follows.

Consider Figure 8 where bathymetric measurements for station 21 for May 20, August 28, and December 10, 1991 are shown. The measurements for May 13, 1992 have been omitted for clarity. These transects clearly show slumping in the shallow, near-shore area, a phenomena which has been qualitatively mentioned by many observers and investigators but has never been quantified. By examining the bathymetric changes, it can
be seen that a large amount of sediment was lost in the shallow areas during the period from May 20 to August 28 while a smaller amount was lost during the period from August 28 to December 10. Other transects show similar behavior.

From these transects, from the flow rates during this time, and from the modeling, the description that is emerging of the sediment transport and the resulting bathymetric changes is as follows. From May 20 to August 28, large amounts of slumping occurred in the shallow areas of the river. Some (less than one-half) of this material was deposited in the channel while the rest was transported into Saginaw Bay. A moderately high flow occurred during this period and eroded some of the sediment in the channel. The net result of the slumping and erosion was a small amount of deposition in the channel. This is consistent with Figure 8.

From August 28 to December 10, a smaller amount of slumping occurred. A moderately high flow also occurred during this period. The flow caused resuspension which was only partially compensated by the slumping. The net result was moderate erosion at this transect, again consistent with Figure 8. These calculations are continuing, and a more complete and quantitative description of our modeling efforts will soon be available.

BUFFALO RIVER

Three tributaries form the Buffalo River. They are the Cayuga, Buffalo, and Cazenovia Creeks. The river winds through the south side of Buffalo for 8.8 km. The drainage basin of these three creeks, before they form the Buffalo River, is 1060 km$^2$. There are no controls on the river or any of the creeks. The river is used as a supply line for heavy industries which line the shores. For this reason, the U.S. Army Corps of Engineers frequently dredges the lower 7.7 km of the river to allow large ships to pass. The project depth limit is 22 feet (6.7 m), but the actual depth often exceeds this value, being as deep as 30 feet (9.1 m) or more in some areas at some times. The river is narrow. In most areas it is only 60 to 100 m wide, although it can be up to 250 m wide at some of
the bends. Because of the narrow width, coupled with the need to dock ships along the banks, the river is dredged almost from side to side in most areas. There are still shallow areas around some of the bends. From studying the natural depth of the river above the limit of dredging, it would appear that the natural depth of the river would be 1 to 3 m.

Because the river has no controls, the flows vary greatly when compared with controlled rivers like the Fox. For a 45 year period from October 1, 1940 to September 30, 1985, the maximum flow on the Buffalo River was 565 m$^3$/s, while the minimum flows were less than 0.5 m$^3$/s. These flows range over three orders of magnitude. The 50th percentile flow on the Buffalo River is only 6.4 m$^3$/s, creating a maximum velocity of 2.1 cm/s. The 90th percentile flow is only 38 m$^3$/s. However, the 99.7th percentile flow is 258 m$^3$/s, creating a maximum velocity of 88 cm/s. It is clear that the river is very slow flowing for most of the year, with large peaks in the flow rate occurring only a few times each year during storms.

Bathymetric measurements at four month intervals as well as flow rates and sediment concentrations during high flows are presently being made for the Buffalo River. By use of these measurements as well as field and laboratory measurements of sediment properties, we are making calculations of sediment resuspension, deposition, and transport similar to those for the Fox and Saginaw Rivers. Details of these calculations will be reported in the near future.

The Buffalo River is much more slow flowing than the Fox or Saginaw Rivers during much of the year, especially during the summer. Flows less than 1 m$^3$/s with flow velocities much less than 1 cm/s are quite common. Because of these slow flows, the river is often thermally stratified. During periods of stratification, the currents and the sediment and contaminant transport are significantly affected by this stratification.

In order to illustrate the effects of thermal stratification on sediment transport, we have made a series of calculations using a three-dimensional, time-dependent model. In these calculations, a simplified geometry as shown in Figure 9 was used, i.e., an incoming river with a constant depth of 1.2 m flowing into a dredged region of the river (8 km long)
with a constant depth of 8.4 m. This river then empties into the lake, also assumed to be 8.4 m deep locally. For the calculations presented here, an average flow velocity of 1 cm/s in the dredged region was assumed. In the incoming flow in the shallow part of the river, the velocities are about 7 cm/s. It was also assumed that the temperature of the incoming river water was 22°C while the temperature of the lake water was cooler at 14°C. Other parameters and the calculation procedure are the same as in our previous work on sediment transport in a stratified river and estuary (Pickens et al., 1992).

For these conditions, the calculated velocities are shown in Figure 9a while the temperature stratification is shown in Figure 9b. It can be seen that the dredged part of the river is strongly stratified over its entire length. Because of this stratification, the warmer river water is confined to a layer near the surface of the river, while the colder lake water is present throughout the river at depth, flows slowly in the upstream direction, and is gradually entrained into the opposite flowing river water.

This stratification also modifies the sediment transport. Because of the low flow and low turbulence, coarse-grained material settles out far upstream or at least near where it is introduced into the river. Only fine-grained material can stay in suspension for any appreciable length of time and distance down river. Figure 10 shows the calculated concentrations for a fine-grained sediment with a settling speed of $5 \times 10^{-4}$ cm/s. For the slowest settling particles, the effects of thermal stratification are most apparent, i.e., the particle concentration is highest in the surface waters and lowest in the deeper waters. In the absence of thermal stratification, the concentrations would be much more uniform in the vertical direction while decreasing much more rapidly in the downstream direction.

It can be seen that thermal stratification affects sediment and contaminant transport during low flow periods. However, sediment and contaminant transport during low flow periods is relatively small compared to these quantities during high flow periods where, because of the nonlinear increase of sediment loading and resuspension with the flow rate, most of the sediment and contaminant transport occurs. Because of this, the net sediment and contaminant transport during the year is probably not significantly affected by thermal stratification.
CONTAMINANT FLUXES

Contaminant fluxes between the bottom sediments and the overlying water occur primarily by a combination of three processes: resuspension/deposition, bioturbation, and diffusion. Each of these processes is quite complex and also is distinctly different from the others. In general, they occur simultaneously, and there are interactions between them. However, in many realistic situations, one of the processes is dominant over the others and so, to a good approximation, can be considered independently. Chemical reactions can significantly affect these sediment-water fluxes. For hydrophobic organic chemicals, the adsorption/desorption reaction (especially its non-equilibrium nature) is particularly significant.

These processes have been discussed and compared previously (Lick, 1992) and so will not be discussed in detail here. As a general summary, the following can be stated. The effects of resuspension/deposition are highly variable in space and time and depend on water depth, topography, and meteorological conditions. During calm periods and average currents and winds, the effects of resuspension/deposition are relatively small and are probably comparable with the effects of bioturbation and diffusion. However, major floods and storms can cause mixing of sediments to depths much greater than that possible by benthic organisms or chemical diffusion. The release of contaminants from the bottom sediments due to this resuspension/deposition and subsequent desorption of these contaminants would also then be much greater than that due to bioturbation or diffusion. The effects of contaminant sorption on resuspension/deposition, bioturbation, and diffusion depend on the particular transport process as well as the rates of adsorption and desorption, but have not yet been quantified. However, consideration of the differences in the fluxes for the two limiting cases of fast sorption (equilibrium partitioning) and slow sorption (frozen partitioning) shows that the effects of sorption rates can be significant and must be considered in determining the flux of contaminants.
SUMMARY AND CONCLUDING REMARKS

In the present paper, recent work on understanding, modeling, and predicting the transport and fate of sediments and contaminants in rivers has been discussed. The numerical models of sediment transport and fate described above are in principle capable of accurately predicting sediment transport and fate. However, they are limited by insufficient laboratory and field data. The major limitations are (a) our inadequate knowledge of the resuspension properties of the bottom sediments, not only near the sediment-water interface but also deeper in the sediments at depths to which the sediments can be eroded in major floods, (b) insufficient data on sediment loading to the river including particle size distribution, and (c) insufficient bathymetric measurements for model input and especially verification. All of this data is necessary for the accurate, predictive modeling of sediment transport and fate.

From the modeling work reviewed here, the following can be stated. The convection of suspended sediments is the most significant sediment and contaminant transport mechanism. Because currents in a river are highly variable and because of the very nonlinear relations between resuspension, bottom shear stress, and flow velocity, the resuspension and transport of sediments is highly variable throughout the year. The transport of hydrophobic contaminants associated with these sediments is therefore also highly variable.

Bed load is important in locally modifying the bathymetry and hence in the flux of contaminants between the bottom sediments and the overlying water. However, it is generally not a major factor in the net transport of sediments and contaminants in rivers. Slumping of sediments is quite evident in the Saginaw River, is significant in modifying the bathymetry and local transport, and is probably important in other rivers as well.

Over long periods of time, resuspension/deposition is probably much more important than either bioturbation or diffusion in the flux of contaminants between the sediments and the overlying water. In determining the contaminant flux, the effects of finite reaction rates
(in particular, the sorption reaction for hydrophobic chemicals) must be considered. Much more work on contaminant fluxes is needed before these fluxes can be accurately predicted.

In all of the experimental work and modeling, the importance of high flow rates and the accompanying high shear stresses and fluxes needs to be emphasized. It is these high flows which, despite their infrequent occurrence, are responsible for most of the sediment and contaminant transport in rivers.

ACKNOWLEDGEMENT

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REFERENCES


Table 1: Maximum Actual and Calculated Bed Changes in the Channel

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<th>TRANSECT</th>
<th>SUSPENDED (cm)</th>
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<th>FIELD MEASUREMENTS (cm)</th>
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FIGURES

1. Fox River Bathymetry, Contour Interval is 1 m.

2. Calculated Velocities for a 99.7 Percentile Flow Rate, 280 m$^3$/s.

3. Calculated Deposition/Resuspension Areas for a 99.7 Percentile Flow Rate.
   - Deposition rate contour interval is 5 gm/cm$^2$ - years.
   - Zero net deposition areas are shaded as ///.
   - Net resuspension areas are shaded as . . . . .


6. Calculated Changes in Bathymetry from October 27, 1989 to September 20, 1990. Contour Interval is 5 gm/cm$^2$. Also Shown are Locations of Transects. (a) Deposition. (b) Erosion.

7. Saginaw River Bathymetry. Contour Interval is 2 m. Also Shown are Locations of Transects.

8. Bathymetric Measurements at Station 21 on the Saginaw River.


10. Sediment Concentrations (mg/L) in a Thermally Stratified River. Settling Speed of $5 \times 10^{-4}$ cm/s.