Sediment resuspension in Lake St. Clair

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Abstract

Time-series measurements of water transparency, wave conditions, and current speed were made at several different sites in Lake St. Clair during five different 1-month periods in 1985 and 1986. Observed changes in suspended sediment concentration were modeled with a simple zero-dimensional, spatially averaged, mass balance model in which local bottom erosion was expressed as a linear function of the bottom shear stress. Estimates of the three parameters required by the model (particle settling velocity, resuspension concentration, and background suspended material concentration) are reasonably consistent for the various data sets, suggesting that the properties of the lake bottom do not change significantly through either space or time. The modeled settling velocities agree with the observed suspended particle size data and the erosion rates are comparable to laboratory results for freshwater sediments. The results show that a simple mass flux model can be used to model local sediment resuspension events in Lake St. Clair with reasonable accuracy.

Concern about environmental problems associated with contaminated sediments has led investigators to examine the resuspension of bottom material in open waters, both in lakes and on the continental shelf. Several of these investigations have included the development of mathematical models intend-

ed to predict changes in suspended sediment concentration with time as a function of the bottom shear stress due to either wave or current action (Aalderink et al. 1985; Lavelle et al. 1984; Luettich et al. 1990; Lyne et al. 1990). One of the problems encountered in developing and applying such models is that it has not yet been possible to relate the erosional behavior of many natural substrates to some easily measured sediment property (e.g. mean grain size). This problem is especially severe for cohesive sediments where laboratory experiments (e.g. Fukuda and Lick 1980; Lee et al. 1981; Maa and Mehta 1987) have shown that the shear strength of cohesive beds depends not only on the particle size but is also a function of composition, pore-water content, deposition history, and the degree of biological reworking. Given this degree of complexity, it is small wonder that resuspension of fine-grained material has been investigated on a case-by-case basis. Nevertheless, models relating the transport of cohesive material to flow conditions are needed, particularly for studies of the movement and fate of sediment-associated contaminants.

Recently, Luettich et al. (1990) successfully used a simple mass flux model to simulate a 2-week time series of suspended sediment concentrations measured in the

nearshore (depth, 2 m) of Lake Balaton, Hungary. Although concurrent measurements of the currents were not available for most of the study period, they were able to use windspeed and direction data with a wave model to estimate the bottom shear stress due to surface waves. These stresses were then used to drive the suspended sediment model. The same mass flux model used by Luettich et al. was used by Aalderink et al. (1985) in their study of sediment resuspension in Veluwe, a very shallow (avg depth, 1 m) lake in the Netherlands. Aalderink et al. also were forced to use wind data in the absence of direct current measurements and examined the success of alternative transfer functions relating windspeed to bottom shear stress. The best transfer function, determined by minimizing the sum of the squared differences between the modeled and observed values over a 15-d period, resulted in a fairly successful simulation of the observations. Application of the model to an independent 40-d time series was equally successful though Aalderink et al. (1985, p. 910) cautioned that spatial variations in sediment types may make it impossible to determine a "uniform and unique relationship between wind and suspended solids which is applicable to all times and at all locations."

If such a uniform and unique model could be found it would greatly simplify operational computations of sediment resuspension and transport. Few data are available, however, either to evaluate the applicability of a simple mass flux model in more open and deeper waters, or to examine the stability of the model parameters over extended periods of time. Our purpose here is to report on our application of the simple mass balance model (similar to those used by Aalderink et al. 1985 and Luettich et al. 1990) to an extensive set of time-series measurements of sediment concentration and flow velocity made in Lake St. Clair as part of the Upper Great Lakes Connecting Channels Study in 1985 and 1986.

Study area and methods

Lake St. Clair (Fig. 1) is a large (40 km wide), shallow (max depth, 7 m) lake between Lake Huron and Lake Erie. The en-
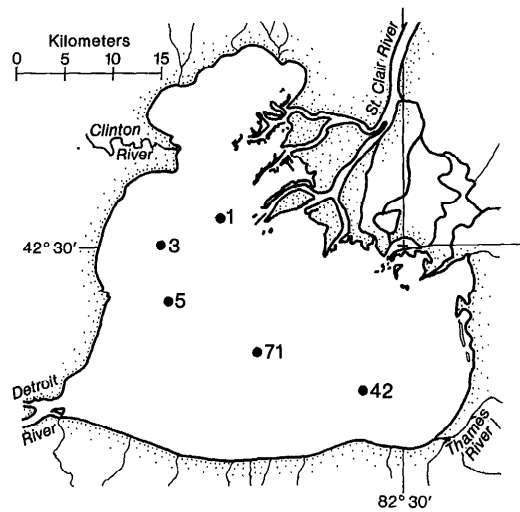

Fig. 1. Map of Lake St. Clair. Tripod deployment sites are indicated by their station numbers.

tire discharge of the upper Great Lakes enters Lake St. Clair via the St. Clair River, whose average flow is 5,100 m^3 s^{-1} (Derecki 1984). Much smaller volumes of water (~150 m^3 s^{-1}) empty into the lake from the Clinton and Thames Rivers. On average, water remains in the lake for only about 9 d before entering the Detroit River, which in turn empties into Lake Erie. Circulation in the lake is mainly hydraulically driven, although meteorological conditions affect the size and shape of the two gyres that are its predominant feature (Schwab et al. 1989). These two gyres correspond to the two major water masses identified by analysis of water quality parameters (Leach 1980); a larger one that occupies the northern and western part of the lake, and a smaller mass in the southeastern part. The larger water mass is dominated by the discharge of the St. Clair River and is relatively clear while the smaller, more turbid water mass is dominated by inflow from the Thames River and other small tributaries.

To study sediment transport in Lake St. Clair we deployed instrumented tripods at various locations in the lake during five separate, month-long periods in 1985 and 1986 (Table 1). Water depth at all stations (Fig. 1) was 6.5 m. Three of the stations (1, 3, and 5) were in the larger, clearer, water mass,

Table 1. Dates and locations of tripod deployments in Lake St. Clair.

	Sta. No.	Location	Tripod	Mud (%)
11 Jul–8 Aug 85	42	42°23′45″N, 82°42′03″W	ANL	25
10 Sep–9 Oct 85	71	42°24′55″N, 82°41′45″W	ANL	35
15–24 May 86	1	42°31′18″N, 82°44′48″W	GLERL	30
15 May–6 Jun 86	3	42°29′42″N, 82°47′42″W	ANL	39
15 May–6 Jun 86	5	42°28′06″N, 82°47′24″W	GLERL	53
8–27 Jul 86	1	42°31′18″N, 82°44′48″W	GLERL	30
8–31 Jul 86	5	42°28′00″N, 82°47′18″W	GLERL	53
8 Jul–8 Aug 86	71	42°25′00″N, 82°40′48″W	ANL	35
10–28 Oct 86	1	42°31′11″N, 82°44′49″W	GLERL	30
10 Oct–11 Nov 86	42	42°23′08″N, 82°44′49″W	ANL	25
10–28 Oct 86	71	42°24′58″N, 82°40′38″W	GLERL	35

while station 42 was in the smaller, more turbid one. Station 71 was near the boundary of the two water masses and, depending on wind conditions, could be located in either. Although in places the mud (silt plus clay-sized particles) content of the lake can exceed 50%, most of the lake floor is silty sand. The percentage of mud at our deployment sites is given in Table 1 (Great Lakes Inst. unpubl. rep.). With the exception of station 5 (53% mud), the bottom sediments at our stations were fairly uniform, ranging from 25 to 39% mud. The particle size spectra of the suspended material in the lake was measured with a model TA-II Coulter counter equipped with a 50-μm aperture. The resulting volume spectra showed a pronounced mode in the 8–10-μm size range. Particles with these diameters accounted for $\sim$30% of the total particle volume.

One tripod, operated by Argonne National Laboratory (ANL), was used during all five deployments. It was equipped with a Sea Tech 25-cm pathlength transmissometer located 0.9 m above the bottom (mab), temperature and current direction sensors at 1 mab, and a Marsh McBirney 512 current meter at 0.7 mab. These sensors were sampled in several different ways as experiment objectives changed. During all but the fall 1985 deployment each sensor was sampled at 3.4 Hz in 75-s bursts (total of 256 measurements per burst). These burst measurements were made every 45 min during summer 1985 and spring 1986 deployments, every hour during summer 1986, and every half hour during fall 1986. The individual current meter samples and the

burst-averaged water temperature and transparency values were recorded for each burst except during the fall 1986 deployment, when the individual (3.4 Hz) transparency measurements also were recorded. During the fall 1985 deployment, only average observations, consisting of continuous 5-min averages of current speed, transparency, and temperature, were recorded. Instrument performance and data recovery were generally very good, but one axis of the current meter failed during the summer 1985 deployment, and the power supply failed for about a week during fall 1986.

Two similarly equipped tripods were operated by the Great Lakes Environmental Research Laboratory (GLERL) during the 1986 deployments. Unfortunately, the current meters on these platforms never worked properly, so only measurements of water transparency (also measured with a Sea Tech 25-cm pathlength transmissometer at 0.9 mab) and temperature (measured at 1 mab) are available for these stations. The instruments were sampled at 1 Hz for 1 min every 15 min; the mean temperature and the mean and standard deviation of the transmissometer readings were recorded.

We converted the transmissometer readings to total suspended material (TSM) values with a calibration determined by Hawley and Zyrem (1990). They compared transparency measurements made with the Sea Tech transmissometer to TSM measurements obtained from filtered water samples and found that

$$c = 0.48 \text{ TSM} + 0.93 \qquad (1)$$

where c is the beam attenuation coefficient

Notation

C	Sediment concn (mg liter^{-1})
C_{bak}	Ambient sediment concn (mg liter^{-1})
c	Beam attenuation coefficient (m^{-1})
D	Water depth (m)
d	Index of agreement
E	Erosion flux (g cm^{-2} s^{-1})
H	Wave ht (m)
L	Wave length (m)
n	Exponent of shear stress
R_c	Resuspension concn (g cm^{-3})
RMS	Root-mean-squared velocity (cm s^{-1})
% RMSE$_u$	Percent unsystematic root-mean-squared error
S	Settling flux (g cm^{-2} s^{-1})
T	Wave period (s)
TSM	Observed total sediment concn (mg liter^{-1})
u_z	Measured current velocity at height z (cm s^{-1})
WOV	Wave orbital velocity calculated from the wave model (cm s^{-1})
w	Particle settling velocity (cm s^{-1})
z	Height above the bed at which u_z is measured (cm)
z_0	Bed roughness length (cm)
κ	von Karman's constant
ν	Kinematic viscosity of water (cm^2 s^{-1})
ρ	Density of water (g cm^{-3})
σ^2_u, σ^2_v	Variance of the velocity components (cm^2 s^{-2})
τ	Bottom shear stress (dyn cm^{-2})
τ_c	Current shear stress (dyn cm^{-2})
τ_r	Reference shear stress (dyn cm^{-2})
τ_t	Threshold shear stress (dyn cm^{-2})
τ_w	Wave shear stress (dyn cm^{-2})
τ_{wc}	Wave shear stress calculated from surface wave model (dyn cm^{-2})
τ_{wm}	Wave shear stress calculated from RMS velocities (dyn cm^{-2})

(units given in the list of symbols). The r^2 value for this equation is 0.91 based on 104 samples.

Although Eq. 1 accurately predicts the TSM for the measured transparency values, all of these measurements were made during fairly quiet conditions when little or no coarse material was in suspension. During storms, given the high percentage of sand in the bottom sediment, it is almost certain that significant quantities of sand were resuspended. Baker and Lavelle (1984) showed that finer material attenuates far more light than the same mass concentration of coarse material, so the TSM values calculated during episodes of resuspension are probably underestimates. Moody et al. (1987) addressed this problem in some detail and concluded that unless some way of collecting samples during these events is devised, such as the bag sampler of Sternberg et al. (1986), TSM estimates will be low.

Vertical profiles of water transparency with an instrument identical to those on the tripods were made whenever a tripod was deployed or retrieved. In addition, vertical profiles were recorded weekly at each station during the summer 1986 deployment. These readings were used to correct the transparency data for the signal degradation that occurred as the result of biological growth on the transmissometers. A simple linear correction between each pair of profiles was used for each site. A fouling correction was also made for the summer 1985 deployment, although it was based on fewer observations. Fouling was not a problem during spring and fall deployments.

Meteorological stations collected continuous above-lake wind data, water temperature, and air temperature during all but the spring 1986 deployment. Scientists from the Canadian Centre for Inland Waters established a meteorological tower near station 71 in 1985 and one near station 42 in fall 1986. During summer 1986, we deployed a meteorological tower near station 71. No above-lake weather data are available for the spring 1986 deployment.

Development of the model

Our primary objective is to implement a model that reproduces the observed changes in local suspended sediment concentration (TSM) given some knowledge of the hydrodynamic forcing. As described above, we made direct, detailed measurements of the current flow three times in 1986, partial observations of the flow during the 1985 deployments, and observations of the above-lake winds during all but the spring 1986 deployment. We describe the simple mass flux model that relates our TSM observations to our flow measurements. We then establish the relationship between winds above the lake and observed flow conditions. Next we use the mass flux model to demonstrate the relationship between both wave and current action and sediment re-

given location. To test whether these calculated wave parameters were reasonable, we used them (both wave height and period) and linear wave theory to calculate the wave orbital velocity (WOV) 1 m above the bottom. We then compared these values to those estimated from the measurements made in summer 1986 at station 71 and in fall 1986 at station 42. The variance (σ^2) for each velocity component was determined from the 256 samples recorded during each 75-s burst and the root-mean-square (RMS) velocity fluctuations calculated from

$$RMS = (\sigma^2_u + \sigma^2_v)^{0.5} \qquad (5)$$

where u and v denote the two horizontal components of the velocity. If there are no other causes of velocity fluctuation, then the RMS velocity fluctuation is due solely to wave action. A linear regression of the combined data gives

$$WOV = -1.645 + 1.652RMS. \qquad (6)$$

Equation 6 is based on 918 data points (Fig. 2A) and has an r^2 value of 0.78. The value of the multiplicative coefficient (1.652) is in good agreement with the theoretical relationship (1.416) between the maximum and RMS values, while the negative intercept reflects velocity fluctuations in the absence of wave action.

We used the wave model to produce time series of estimated wave conditions for all of the stations except those occupied during spring 1986 (for which we have no above the lake wind measurements). We then estimated the bottom wave shear stress (τ_{wc}) from the calculated wave parameters by using linear wave theory combined with the usual expression for bottom shear stress in a laminar boundary layer (Madsen 1976). This expression can be written in terms of wave height and period as

$$\tau_w = \frac{H\rho\nu^{0.5}(2\pi/T)^{1.5}}{2\sinh(D2\pi/L)} \qquad (7)$$

where H is the significant wave height, T the wave period, L the wave length (calculated from T), ρ the density of water, and ν the kinematic viscosity of water.

We also used Eq. 7 to estimate a "measured" wave shear stress (τ_{wm}) from the three
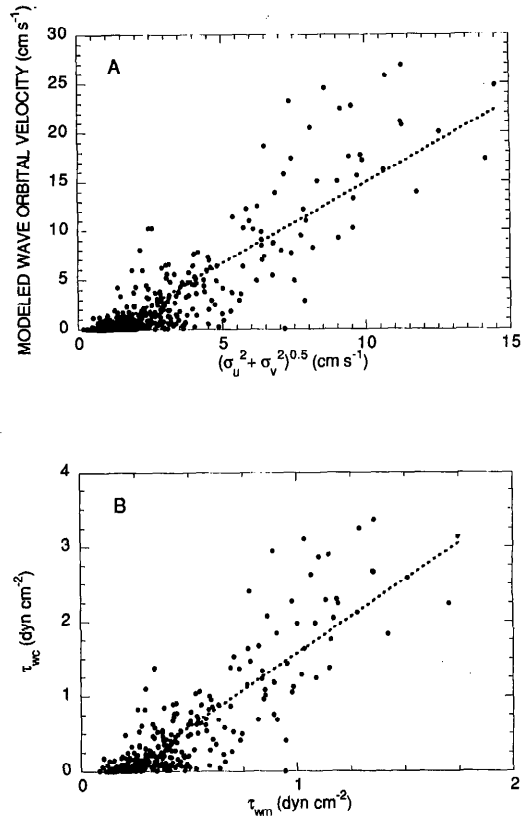
Fig. 2. Comparison of the results from the wave model and the current velocity measurements for the fall 1986 deployment at station 42 and the summer 1986 deployment at station 71. A. The RMS and the calculated wave orbital velocities; regression line is given in Eq. 6. B. The calculated and measured wave shear stress; regression line is given in Eq. 8.

sets of RMS values collected in 1986. An equivalent wave height was calculated from the measured RMS velocity fluctuations and the wave period determined from power spectrum analysis of the velocity measurements. Because the burst length was short (256 samples), resolution of the period was only 0.5 s (the wave model resolves the period to 0.1 s). Figure 2B shows the values of τ_{wm} and τ_{wc} for the same data sets used in Fig. 2A. The regression line is

$$\tau_{wc} = -0.39 + 1.97\tau_{wm}, \qquad (8)$$

with $r^2 = 0.74$. Schwab et al. (1984) showed that the wave model underestimates the wave period, which accounts for the increased slope in Eq. 8 relative to that in Eq.

6. The good correlations for Eq. 6 and 8, and the good agreement between the observed and predicted wave heights in Lake St. Clair (D. J. Schwab pers. comm.), give us confidence in using the wave model results as a surrogate for measured wave parameters.

To examine the relative influence of the mean flow, we also estimated the bottom current shear stress (τ_c) from the measured average velocities. These measurements were available for four deployments (those occupied by the Argonne tripod in 1986 and fall 1985). The stress was calculated from

$$\tau_c = \rho \left[\frac{\kappa u_z}{\ln(z/z_0)} \right]^2 \qquad (9)$$

where κ is von Karman's constant (0.4), z is the height above the bottom (0.7 m) at which the velocity (u_z) was measured, and z_0, the roughness length, was assumed to be 0.02 cm. Other values of the roughness length could be used, but serve only to scale the stress estimates. Lavelle et al. (1984) and Luettich et al. (1990) found that their models were relatively insensitive to the assumed value of z_0.

The wave shear stresses are usually considerably greater (by a factor of 3–10) than the corresponding current shear stresses. Although the maximum wave period in the lake is only ~4 s, wave heights may reach 1 m, leading to estimated shear stresses > 3.5 dyn cm^{-2} in some cases. In contrast, the highest current shear stress was 1 dyn cm^{-2}, suggesting that wave action is probably more important than current action in resuspending sediment in the lake. However, because the effects of waves with periods < 2.9 s do not reach the bottom (depth, 6.5 m), there are time intervals in all of the deployments when the wave shear stress is zero. There are no such intervals for the current shear stress which—although usually quite small—is never zero.

Implementation of the model—We used the method of Luettich et al. (1990) to integrate Eq. 4a,b and followed their example by setting the erosion coefficient, α, equal to the product of the settling velocity, w, and a resuspension concentration, R_c. Our preliminary tests also confirmed their find-

ing that the three model parameters n, R_c, and τ_t are highly correlated. Because this correlation makes two of these parameters redundant, nothing is gained by solving for all three in the model. Luettich et al. used regression equations to determine values of R_c and n from their estimates of the threshold value but since we obtained acceptable results in our preliminary model runs with n equal to one and τ_t to zero, we have used these values in all of our simulations. Other values could be used but would only serve to alter the values of R_c, not improve the fit of the model.

Setting τ_t to zero implies that there is no threshold for sediment resuspension in our model. Although this is contrary to more traditional views about the onset of sediment transport, Lavelle et al. (1984) and Lavelle and Mofjeld (1987) argued that sediment movement is a stochastic process and that a threshold value may be an artifact of the experiments done to determine criteria for sediment transport. They found that previous experimental results could be explained equally well without a threshold stress. Their findings (and those of Luettich et al.) suggest that the data collected to date are not adequate to definitively determine whether a non-zero threshold value is needed.

Because we are applying the model to data sets collected at different times and locations, we also let C_{bak}, the nonsettling background concentration, vary. Thus the solution to Eq. 4a,b (given by Luettich et al. 1990) is

$$C_i = R_c \frac{\tau}{\tau_r} + C_{bak}$$

$$+ \left(C_{i-1} - C_{bak} - R_c \frac{\tau}{\tau_r} \right)$$

$$\cdot \exp\left[\frac{-w\,\Delta t}{D} \right] \qquad (10)$$

where C_i is the predicted concentration at time i and Δt is the time step. In our formulation, Eq. 10 has three free parameters, C_{bak}, R_c, and w, that are independent of each other. Note that τ is divided by a unit reference stress to keep the units correct.

Examination of Eq. 10 reveals several features that may be expected of the simulation and of the parameter values. C_{bak} sets a constant, asymptotic (in the absence of forcing), lower bound on the predicted concentration (C_i) and thus should be close to the lowest observed concentration. The settling velocity (w) appears only in the exponential term and determines the rate of mass loss by settling during each time step. For our values of Δt (2,700 or 3,600 s) and D (650 cm), reasonable values of w (0.001–0.01 cm s^{-1}) give an exponential term of about one (0.946–0.996). Because the magnitude of the exponential term varies only slightly, we expect the model to be relatively insensitive to the precise value determined for w. The value of R_c, however, should greatly influence the model results because it is multiplied by the forcing term, which varies considerably. Indeed, our results show that the model is most sensitive to changes in R_c, followed by changes in w and then in C_{bak}.

Model evaluation—Evaluation of the fit of the model to the data was done with Willmott's (1982) index of agreement (d) and percent unsystematic error (% RMSE$_u$). The index of agreement is defined by

$$d = 1 - \left[\sum_{i=1}^{N} (p_i - o_i)^2 \right.$$

$$\left. \div \sum_{i=1}^{N} (|p'_i| + |o'_i|)^2 \right] \quad (11)$$

where p_i is the model value at time i, $p'_i = p_i - \bar{o}$, $o'_i = o_i - \bar{o}$, and the overbar indicates the average value of the observations (o_i). Unlike the more familiar r^2, which is a measure of correlation, d is a direct measure of the total error in the model. The total error can be divided into its systematic and unsystematic parts, where the systematic part represents the bias in the model and the unsystematic part the random error. Willmott's unsystematic root-mean-squared error (RMSE$_u$) measures the unsystematic error by calculating the average deviation of the model predictions from the values ($\hat{p}_i$) that would be predicted by a linear regression of the modeled values on the observations. That is,

$$\mathrm{RMSE}_u = \left[N^{-1} \sum_{i=1}^{N} (p_i - \hat{p}_i)^2 \right]^{0.5}. \quad (12)$$

The % RMSE$_u$ is then obtained by dividing Eq. 12 by the total root-mean-squared error. Values of both d and the % RMSE$_u$ range from zero to one. In general, we selected as the best fit that set of parameters for which the index of agreement is a maximum. However, in several cases the index of agreement varied only slightly (within a percentage point or two) for several sets of parameters, while % RMSE$_u$ varied substantially (by up to 20%). In these cases we chose the parameter set that had the highest value of % RMSE$_u$.

Results

By systematically varying the three model parameters, we determined the best set of parameter values (C_{bak}, R_c, w) for each set of observations with each available forcing. Because we were interested in comparing spatial and temporal differences in the model results, these determinations were based on simulations of the entire data series—as opposed to simulation of a single resuspension event within a series. The results of these simulations are summarized by forcing variable, season, and station in Table 2. The ability of the model to reproduce the fall observations is excellent (d and % RMSE$_u$ approaching 1.0) at all stations for all of the forcing variables (except for τ_c in the fall 1985 deployment, discussed below) but not as good for spring and summer observations. Although the model appears to better simulate conditions in the southern part of the lake (stations 42 and 71) than in the northern part (stations 1, 3, and 5), that may be because all of the spring observations were made in the northern area and most of the fall observations in the south. The simulations based on stress estimates made from modeled wave conditions (τ_{wc}) are similar to those made from the observed wave velocity measurements (τ_{wm}), but because the latter have smaller maximum values, the values of R_c are slightly greater. This variation in R_c is more pronounced in the simulations using the current shear stress (τ_c) as the forcing. Illustrations of the rela-

Table 2. Summary of model optimization parameters and statistics for all deployments.

Sta., deployment	Shear stress* (dyn cm^{-2})	C_a (mg liter^{-1})	R_c (mg liter^{-1})	w (cm s^{-1})	d	% RMSE$_u$
42, Fall 1986	τ_{wc}	4.0	43.0	0.007	0.93	0.98
71, Fall 1986	τ_{wc}	3.0	18.0	0.007	0.97	0.98
71, Fall 1985	τ_{wc}	2.0	17.0	0.005	0.98	0.98
1, Fall 1986	τ_{wc}	4.0	15.0	0.011	0.92	0.96
71, Summer 1986†	τ_{wc}	3.0	36.0	0.006	0.46	0.50
1, Summer 1986	τ_{wc}	2.0	33.0	0.004	0.77	0.96
5, Summer 1986	τ_{wc}	4.0	16.0	0.010	0.76	0.90
42, Summer 1985†	τ_{wc}	1.0	59.0	0.004	0.57	0.73
42, Fall 1986	τ_{wm}	2.0	62.0	0.010	0.90	0.95
71, Summer 1986†	τ_{wm}	1.0	43.0	0.002	0.50	0.43
3, Spring 1986	τ_{wm}	3.0	16.0	0.007	0.70	0.77
5, Spring 1986	τ_{wm}	3.0	14.0	0.007	0.61	0.80
1, Spring 1986†	τ_{wm}	3.0	8.0	0.086	0.61	0.74
42, Fall 1986	τ_c	4.0	158.0	0.029	0.93	0.98
71, Fall 1985	τ_c	1.0	158.0	0.011	0.61	0.15
71, Summer 1986†	τ_c	1.0	104.0	0.039	0.56	0.71
3, Spring 1986	τ_c	1.0	62.0	0.032	0.39	0.39
5, Spring 1986	τ_c	1.0	57.0	0.073	0.33	0.45
1, Spring 1986†	τ_c	2.0	43.0	0.140	0.50	0.64

* τ_{wm}—stress is calculated from RMS velocities; τ_{wc}—stress is calculated from wave model parameters; τ_c—stress is calculated from measured mean velocities.
† Excluded from discussion.

tionships between the observed concentrations, the model predictions, and the forcing functions for the data collected in 1986 are shown in Figs. 3–5. Since in all cases, the values of the current shear stress are much less than the corresponding wave shear stresses (Figs. 3D, 4D, and 5D, note the changes in scale), we believe that wave action is the major cause of sediment resuspension in the lake.

Data collected from stations 1 and 42 during the fall 1986 deployment are shown in Fig. 3. Background concentrations are ~4 mg liter^{-1} at station 1 and 8 mg liter^{-1} at station 42 (due to their positions in different water masses), but the concentrations increase greatly during events with large shear stresses. The high correlation between the periods with high TSM values and those with high shear stresses strongly implies that resuspension is occurring and that the model should give good results. The highest observed TSM values were recorded at station 42 during a storm on 14–16 October, as were the highest shear stresses. Unfortunately the transmissometer became saturated during this period, so the values shown are only the maximum value that could be recorded; they are almost certainly under-

estimates. High TSM values were also recorded at stations 1 and 71 (not shown).

Although the TSM values usually decreased to background levels within a day after wave action ceased (as illustrated at station 1, Fig. 3A), at station 42 the concentration remained almost constant at a value somewhat above the background level during the 3-d period beginning on 17 October. A similar pattern was also seen at station 71 (although the period lasted only 1.5 d, from 17 to 18 October). We believe that these persistent high concentrations resulted from the movement of the more turbid water mass associated with the Thames River and variations in the TSM concentration within this water mass. Winds were from the southwest on 14–16 October, but changed to the north the next day. Schwab et al. (1989) showed that southwest winds will increase the area covered by the turbid water mass to include both stations 42 and 71, while northerly winds will confine the turbid water mass to the southern shore. We believe that the elevated TSM concentrations at station 71 on 17 and 18 October were caused by the presence of the more turbid water mass at this location and that the abrupt decrease in concentration marks
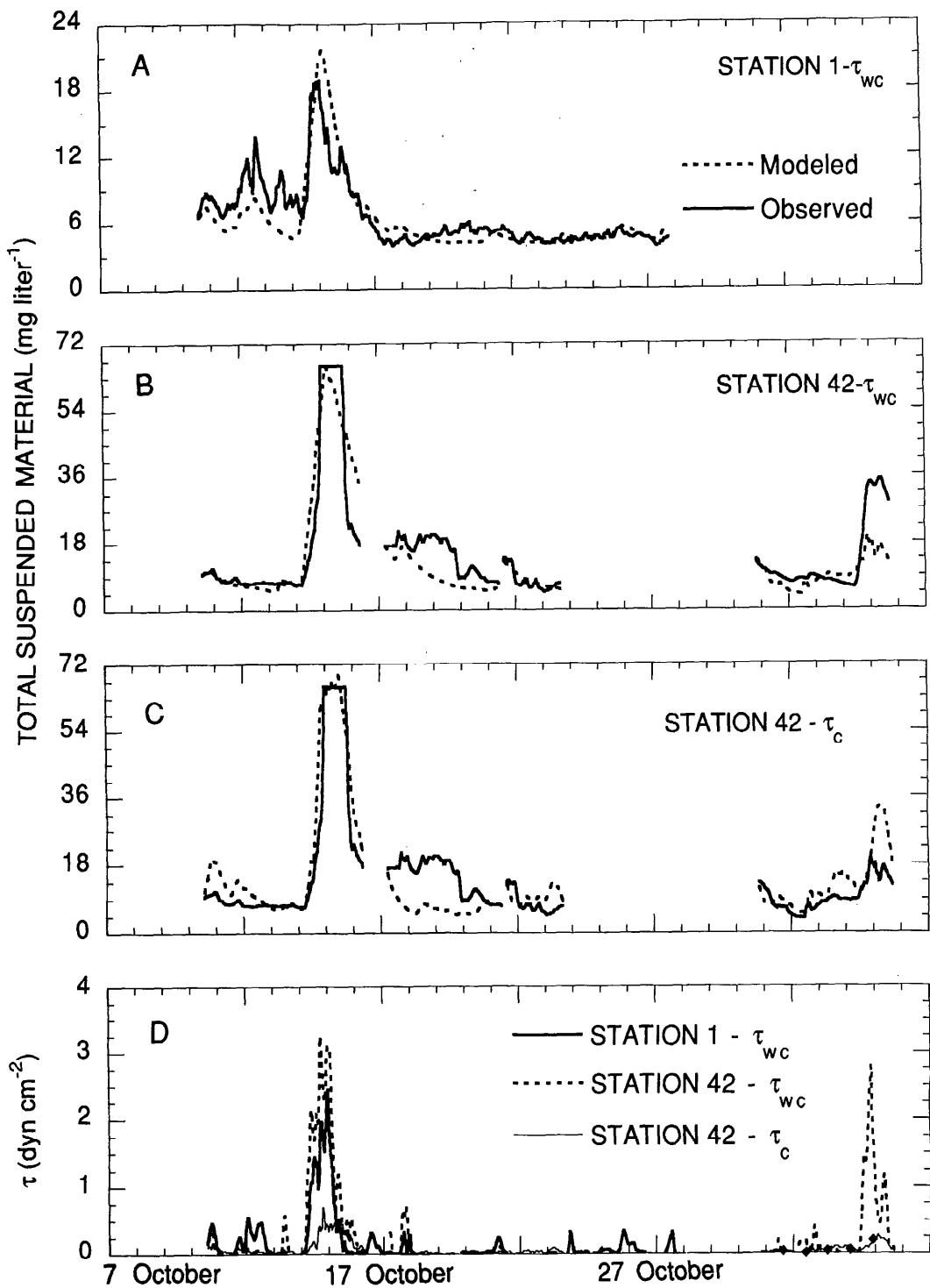
Fig. 3. Data from the fall 1986 deployments. A, B. Observations and model simulation with τ_{wc} as the forcing. C. Observations and simulation with τ_c as the forcing. D. Shear stresses. The gaps in records at station 42 are due to data tape reversal, power supply failure, and tripod redeployment respectively.

the movement of the turbid water mass back to the south.

Station 42 was always within the more turbid water mass, so variations in the TSM concentration within the mass may explain the observations at that station. Unpublished data collected by scientists at the Canadian Centre for Inland Waters show that the discharge of the Thames River forms a distinct highly turbid plume in the lake. The 20 mm of rain that fell during the storm (as recorded at Windsor airport, Fish. Environ. Can. 1986) caused an increase in the discharge of the Thames River during this period (data from the gauging station upstream at Thamesville show that the discharge on 14–18 October was double that of previous days, Inland Waters/Lands Directorate 1987). We suspect that the sediment load was also increased, although we have no measurements to confirm this. This combination of increased flow and sediment load produced a patch of highly turbid water within the water mass that was then advected past station 42 during the storm, thus increasing the already high TSM concentration due to local resuspension. Advection of turbid water would explain the TSM concentrations on 17–20 October and would also explain why the TSM levels during the storm on 3 and 4 November are much less than those on 14–16 October (although the winds were also from the southwest, there was no rain during this period), even though the wave shear stresses are about the same.

The model fits for the fall data at station 42 are very similar for all three forcing functions, although the values of the model parameters (particularly R_c) vary. The similarity of these fits accords with our observation that although the magnitudes of the various forcing functions varied, the patterns of the changes through time were similar (the pattern of τ_{wm} at station 42, although not shown in Fig. 3, is similar to that of τ_{wc}). The variation in R_c is simply a reflection of the differences in the magnitudes of the three stress estimates. Regardless of the forcing used, the model overestimates the observed sediment concentration on 3 and 4 November, which is not surprising since, as noted above, the TSM observations on 14 and 15 October

are underestimates (due to the transmissometer saturating) and the model attempts to match them. However, if one attempts to compensate by increasing R_c, then the model errors on 3 and 4 November become even greater. As noted above, we believe that the high TSM levels on 14–20 October are partly due to advection of highly turbid water discharged from the Thames River.

The model accurately simulates the observations at station 1 during the storm on 14–16 October and reproduces (though with less success) the smaller local resuspension events that occurred between 10 and 14 October. If the wave stress is used as the forcing function, the model also accurately simulates the observations at station 71 during both 1985 and 1986 (Table 2). However, if the current stress is used as the forcing, the model does not accurately simulate the 1985 observations. This poor fit is because the estimated current shear stresses have a very limited range (0–0.25 dyn cm^2) and do not have a peak during a storm which caused high TSM concentrations.

As might be expected, the magnitudes of the wave shear stresses in summer 1986 were lower than those observed during fall. Consequently, the fluctuations in observed TSM due to local sediment resuspension were generally small. However, model fits to the summer data collected in 1986 at station 1 (Fig. 4A) and station 5 (not shown) are fairly good. Despite the limited range of the TSM values, the model is sensitive enough to simulate the several small resuspension events that occur. The observations at station 5 are similar to those at station 1.

The poor fit of the model to the summer observations made at station 71 (Figs. 4B and C) and at station 42 in 1985 (not shown) is due to the presence of conditions that violate the assumptions of the model. In addition to the measurements of water temperature made at the tripod (1 mab), measurements were also made at 4 mab at the weather station located within 500 m of the tripod. These measurements, when combined with the weekly vertical profiles of transparency made at station 71, show that warmer, clearer water often overlaid a colder, more turbid bottom layer in summer. The summer 1985 deployment at station 42 shows similar features. Several of
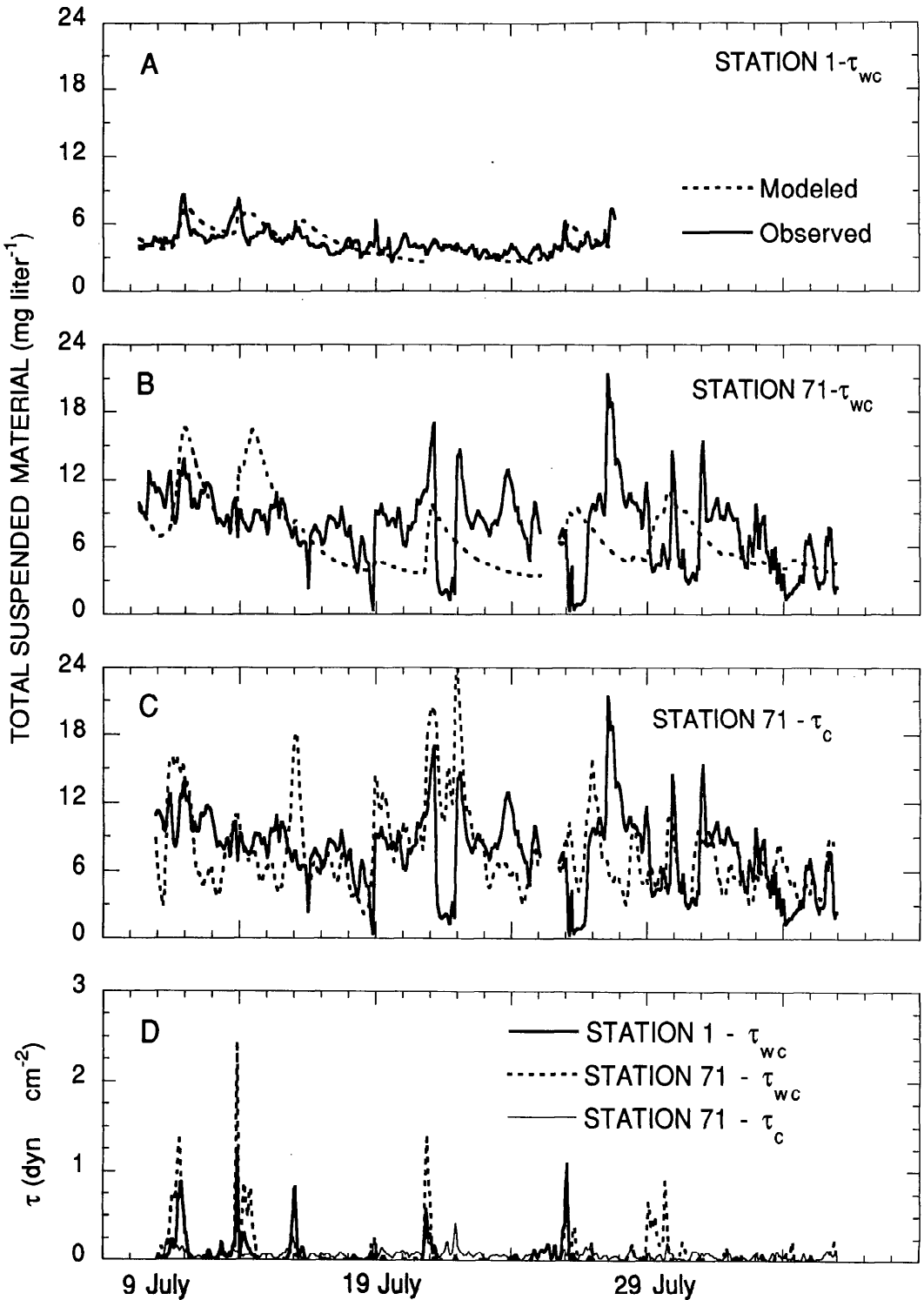
Fig. 4. As Fig. 3, but from the summer 1986 deployments. The gap in the record at station 71 occurred when the data tape reversed.

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the observed changes in suspended sedi-
ment concentration at station 71 (and to a
lesser extent at station 42) are due to shifting
of the boundary between these layers rela-
tive to the position of the transmissometer.
Because the model is based on the assump-
tion that resuspension and settling are the
dominant processes controlling the ob-
served suspended sediment concentration,
any change in concentration not caused by
resuspension (e.g. advection of the nephe-
loid layer) will result in a model error. The
presence of this layering also violates the
assumption of vertical homogeneity. Given
these violations of the model assumptions,
it is not surprising that the model simula-
tions are not very accurate for this deploy-
ment, regardless of which forcing function
is used (this is the only case where the sim-
ulation based on the current shear stress is
better than those based on the wave shear
stress). The lake was not stratified at our
other summer stations or during spring and
fall deployments.

Spring results are shown in Fig. 5. We
have no wind data for this period, so we
were unable to use the wave model to es-
timate wave stresses. The model results
shown in Table 2 were calculated by apply-
ing the current meter measurements made
at station 3 to all three stations. The TSM
values are considerably lower than during
fall and about equal to those observed dur-
ing summer. Six separate TSM peaks were
observed at station 3, but only three of them,
those beginning on 16 May, 1 June, and 3
June are associated with high wave shear
stresses. These three episodes also occur at
station 5 and the first at station 1 (note that
this is a short record), so they are almost
certainly wave-induced resuspension. The
two episodes that began on 22 May and 4
June are probably due to advection of a
turbid slug of water past the stations. Data
from six water-intake plants along the St.
Clair River indicate that turbid water did
pass down the river at these times. It is not
clear what caused the sixth event (the one
on 17 May), but it may be due to current-
induced resuspension. The relatively poor
fit of the model for these deployments is
due in part to the advective episodes in the

records and (for stations 1 and 5) to the use
of forcing functions recorded at a different
location. In addition, on 18–20 May there
are relatively high wave stresses but no in-
crease in sediment concentration. We have
no explanation for this. Because the record
at station 1 contains only one resuspension
event, that event is fit extremely well but
the settling velocity is quite large—approx-
imately an order of magnitude greater than
in the other deployments (when wave stress
is used as the forcing). With neither a second
resuspension episode to test the model pa-
rameters nor velocity measurements made
at the station, we are not confident that the
results are valid. The TSM observations at
station 5 (not shown) are fairly similar to
those at station 3, so the model results are
also similar.

The model does not do as well when the
current shear stress is used as the forcing.
Even at station 3, where the measurements
were made, the fit is quite poor. In part this
is because the current shear stress does not
correlate with the wave shear stress during
this deployment, so the wave-induced re-
suspension events are not well simulated.
At station 1, as with the wave stress simu-
lation, the model value for w is much larger
than it is for the other stations.

Discussion

Because we did not directly measure the
vertical flux of resuspended sediment, we
are actually only inferring when resuspen-
sion occurs, rather than measuring it di-
rectly. Thus our TSM observations could
be caused not only by resuspension, which
is what we are trying to model, but also by
other processes such as lateral advection.
Because these other processes may violate
the model assumptions of lateral and ver-
tical homogeneity, we need to distinguish
TSM peaks due to resuspension from TSM
peaks due to other causes. We have noted
several events which we believe were not
caused by resuspension. They were identi-
fied by carefully examining the time-series
records of the sediment concentration and
the forcing function (whatever it may be)
and evaluating them with two criteria. First,
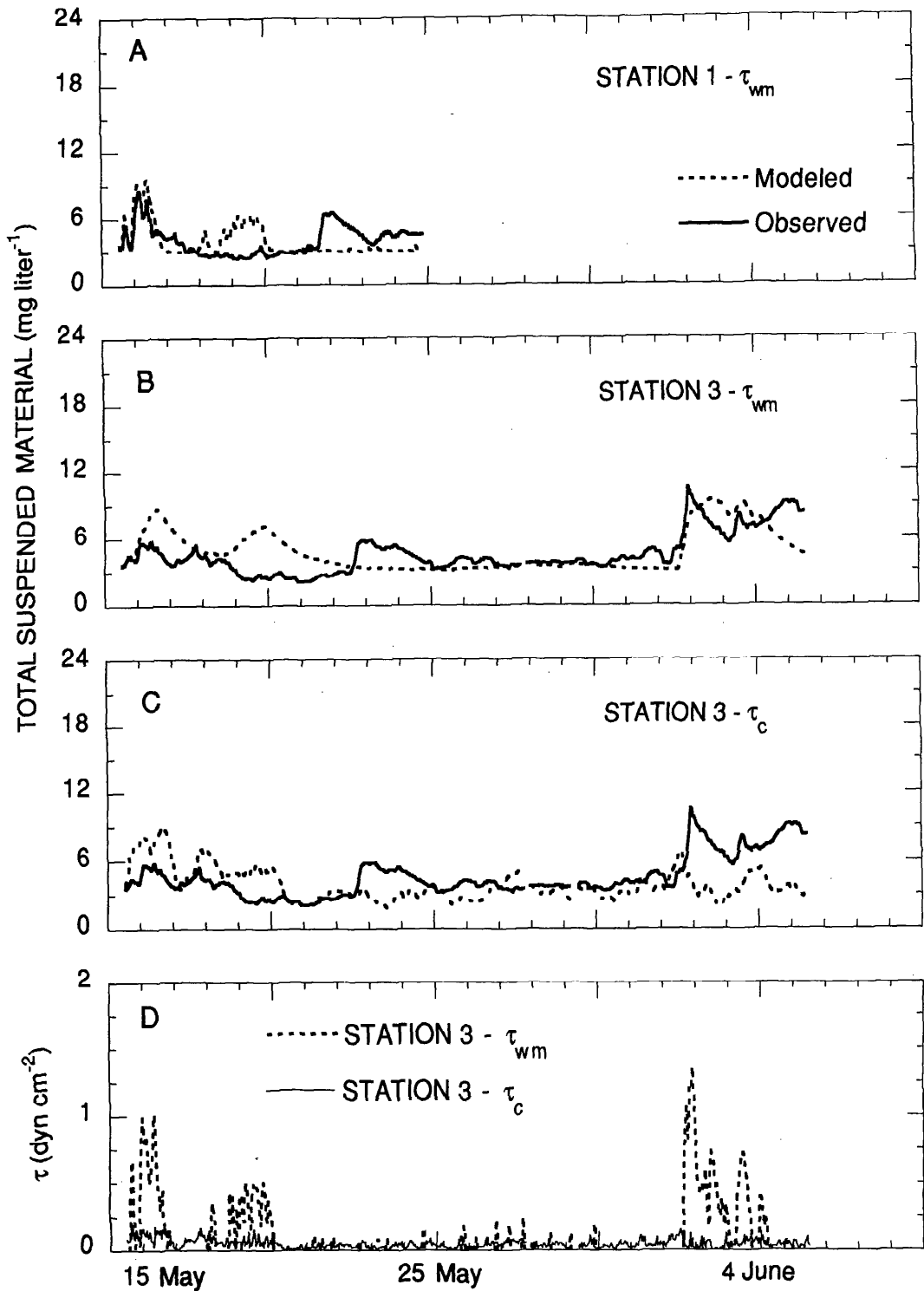an abrupt increase in sediment concentra-

Fig. 5. Data from the spring 1986 deployments. A, B. Observations and model simulation with τ_{wm} as the forcing. C. Observations and simulation with τ_c as the forcing. D. Shear stresses. The gap in the record at station 3 occurred when the data tape reversed.

tion should occur simultaneously with an increase in the forcing function. Second, after the forcing has ceased, the sediment concentration must decrease to background levels within a reasonable amount of time (the actual value will depend on water depth and the fall velocity of the suspended material). We would not expect high TSM values caused by advection, for instance, to be strongly correlated with changes in the forcing function, although they might be related to changes in other water mass properties, such as temperature or conductivity. Suspected episodes of lateral advection were confirmed with other data; the water intake records in spring, the temperature and profile measurements in summer, and the rainfall and discharge measurements in fall.

Resuspension events due to wave action do not violate the assumption of lateral homogeneity because wave action usually occurred almost simultaneously throughout the lake (although its intensity varied with location). The erosion of sediment and its subsequent deposition also occurred almost simultaneously throughout the lake, allowing us to treat the observations at each station as a record of purely local erosion and deposition, without worrying about lateral transport of material from one site to another. The assumption of vertical homogeneity is valid during spring and fall, but during summer it was not always so. If the water is not vertically homogeneous the model will incorrectly estimate R_c, because D is now no longer the appropriate model height. However, one can also argue that during resuspension episodes, the water was probably homogeneous. Because we measure TSM at only one point in the water column we cannot resolve this problem from our data.

The apparent seasonal difference in the ability of the model to reproduce the observations reflects the increased likelihood that the key model assumptions of vertical and horizontal homogeneity were violated during summer and spring deployments. During fall, changes in the observed suspended sediment concentration at all stations are usually dominated by local resuspension events associated with storms. Resuspension is not dominant in spring,

when analysis of water intake data shows that several of the observed concentration increases are due to advection of turbid water from the St. Clair River past our stations, nor during the summer at stations 42 and 71, when lateral advection also occurred. Because these advection events violate the model assumptions, the fit of the model to the records containing these events is poorer and the estimates of the model parameters less reliable. We do not believe that the results from the summer deployments at stations 71 and 42, or the spring deployment at station 1 (which was quite short), are reliable enough to be included in the discussion below.

The major differences between the current shear stresses and the wave shear stresses (either τ_{wc} or τ_{wm}) are that the wave stresses are usually larger by a factor of 3–10 (when they are not zero), and the current shear stress is never zero. These differences are reflected in the values of the model parameters. Values of R_c are larger when the current stress is used as the forcing because it is always multiplied by the stress in Eq. 10. Also, because the current stress is never zero, there is some resuspension every time step. This constant resuspension requires a larger value of w to redeposit the sediment. Our preliminary model runs showed that if a threshold forcing value is included in the model, the values of w decrease, but the model fits are no better. As stated above, we believe that since the wave stresses are considerably greater than the corresponding current stresses, wave action is the major cause of sediment resuspension in the lake. The model results tend to support this conclusion, since in only one case out of four is the simulation with the current shear stress as good as the one with the wave shear stress.

Examination of Table 2 shows that when the wave stress is used as the forcing (calculated from either the velocity measurements or the wave model parameters), the estimates of the model parameters for most of the deployments are remarkably consistent. Values of w vary between 0.004 and 0.011 cm s^{-1}, C_{bak} between 2 and 4 mg liter^{-1}, and R_c (except for three values) between 14 and 18 mg liter^{-1}. The exceptions have much higher values of R_c. Two of these

exceptions are the deployments at station 42. Because this station is the only one consistently located in the small, highly turbid water mass dominated by the Thames River, it is reasonable to expect the erosional properties of the sediment to be different from those in the rest of the lake. In addition, as noted above, the high TSM values are sometimes partly due to advection, which leads to higher values of R_c. However, the other exception—the summer deployment at station 1—is located in the main water mass. We reran the model for these three deployments with τ_{wc} as the forcing and a set of "generic" parameter values (C_{bak} equal to 3 mg liter^{-1}, R_c equal to 16 mg liter^{-1}, and w equal 0.007 cm s^{-1}).

The results (Table 3) show that the reduced value of R_c gives much poorer results at station 42, but that at station 1 the value of d is almost identical to that listed in Table 2. The value of % RMSE$_u$, however, is substantially lower. A decrease in % RMSE$_u$ indicates an increase in the bias of the model parameters. The effect of this bias is a reduction in the TSM values produced by the model. The new modeled peak TSM levels are lower than those shown in Fig. 4A, but they decay more quickly to the ambient level, so the overall fit of the model is almost unchanged. Although these generic parameters are more biased than those listed in Table 2, the small change in d leads us to believe that the model parameters for all of the stations except station 42 are about the same. Even at station 42, the model results with the generic parameters fit the concentrations observed on 3 and 4 November (fall 1986 deployment) fairly well, so it may be that the same model parameters could be used throughout the lake.

Given that our measurement errors are ~0.4 mg liter^{-1} and 0.5 cm s^{-1}, the consistency of the model parameters is remarkably good. Unfortunately, we know of no way to assign error values to our parameter estimates, so we cannot test to see whether the R_c values at station 42 are statistically different from those at the other stations. We can, however, look at the effects of varying the values of the parameters. We did a sensitivity analysis by computing the model results for all combinations of the three pa-

Table 3. Change in model-fit statistics for selected deployments with parameter values typical of all deployments ($C_a = 3$ mg liter^{-1}, $R_c = 16$ mg liter^{-1}, $w = 0.007$ cm s^{-1}) when model is driven by wave shear stress (τ_{wc}).

Sta., deployment	d	% RMSE$_u$	Δd	Δ % RMSE$_u$
42, Fall 1986	0.67	0.07	−0.26	−0.91
42, Summer 1985	0.46	0.10	−0.11	−0.63
1, Summer 1986	0.76	0.60	−0.01	−0.36

rameters between 0.5 and 10 mg liter^{-1} for C_{bak}, 1 and 300 mg liter^{-1} for R_c, and 0.001−0.1 cm s^{-1} for w. The values of d and % RMSE$_u$ were then computed for each set of parameters and compared to those listed in Table 2. We found that the model was most sensitive to changes in R_c; changes in w and C_{bak} had much less effect. For values of d and % RMSE$_u$ within 0.1 of the values listed in Table 2, for most of the model runs C_{bak} could range between 0.5 and 5 mg liter^{-1}, and w could vary between 0.003 and 0.1 cm s^{-1}. Acceptable values of R_c ranged from 6 to >100 mg liter^{-1} at station 42 and in most cases between ~3 and 35 mg liter^{-1} at the other stations. Note that this does not mean that all combinations of the parameters give acceptable results, but that some combinations within these ranges do. Although there may be two distinct sets of parameters, one for station 42 and the other for all other stations, the variability in the acceptable parameter values does not allow us to definitively distinguish one set of parameters from another. Thus, for our model and data, we cannot say that the erosional and depositional properties of the sediment varied over either space or time.

If the model values of R_c are multiplied by the corresponding settling velocities, we can compare our results to the values of α tabulated by Lavelle et al. (1984). Our values vary between ~10^{-6} and 10^{-8}, within the range they cited for experimental work done with unidirectional flow. Maa and Mehta (1987) reported the results of a laboratory study on the resuspension of cohesive sediments by waves, but reported only a normalized erosion rate that cannot be compared to other results without more information. Their results suggest, however, that wave action is considerably more effi-

cient at resuspending sediment than steady flow. However, we have no way of comparing our erosion rates to theirs.

For our simulations with wave shear stress as the forcing, the average value of w is 0.0076 cm s^{-1} or ~6.6 m d^{-1}. Thus—in the absence of resuspension—the model predicts that the water column (depth, 6.5 m) would clear in about a day. If a particle density of 2.65 g cm^{-3} is used, Stokes' law predicts that a particle ~9 mm in diameter will have this velocity—in excellent agreement with our particle size observations. The runs using current shear stress have somewhat higher values of w—the average is 0.032 cm s^{-1}, or 31 m d^{-1}—the Stokes' velocity for a particle with a diameter of 20 μm.

The model values for C_{bak} vary over a relatively small range and are consistent with the observed TSM concentrations in the lake during quiet weather. Thus the values for all three of our model parameters are consistent with other, independent measurements.

Conclusions

As far as we know, this investigation is the first to quantitatively determine the fit of a sediment resuspension model to several data sets collected at different times and locations. Lavelle et al. (1984) and Luettich et al. (1990) analyzed results from a single deployment. Lyne et al. (1990) presented observations from several stations over long time periods but did not quantitatively compare their model results to their observations. We found that although similar values of the index of agreement could be obtained for many combinations of the three model parameters, the use of % RMSE$_u$ helped considerably in determining the "best" model fit. Our model parameters are physically realistic and in the same range as previous results.

Our results show that when the wave shear stress is used as the forcing function, the model closely predicts the occurrence and rate of resuspension. Simulations done with current shear stress as the forcing function usually gave poorer results. Since the wave stress estimates are also considerably larger than the corresponding current stresses, we believe that wave action is the major cause of sediment resuspension in the lake. To a large degree, the apparent differences in model results with time and location are due to the complications introduced by advective episodes. The relative consistency of the model parameters through both space and time (Table 2) suggests the model is relatively robust and that small differences in substrate composition do not affect the results. However, the wide range of acceptable parameter values (based on a sensitivity analysis) prohibit us from concluding that the parameter values were the same for all of the deployments. The results also suggest that this relatively simple model may be useful in modeling sediment resuspension in a wide variety of settings, even though, as Luettich et al. (1990) noted, it does not allow for the complications introduced by cohesive sediments.

The dynamics of sediment movement in Lake St. Clair are complex, and processes other than resuspension may also cause changes in TSM concentrations. The most intriguing of these is the apparent movement of a bottom nepheloid layer observed in summer. Further study of this problem is needed. Future studies should also emphasize collecting data that will allow the importance of river discharges to be quantified and lateral advection episodes to be better identified.

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