J. Great Lakes Res. 18(4):571–589
Internat. Assoc. Great Lakes Res., 1992

THE PHYSICAL EFFECTS OF THE GREAT LAKES ON TRIBUTARIES AND WETLANDS

A SUMMARY

Keith W. Bedford
*Department of Civil Engineering
The Ohio State University
Columbus, Ohio 43210*

ABSTRACT. *Wetlands and tributary confluences are susceptible to physical influences imposed by the Great Lakes, particularly through the effects of short and long-term water level fluctuations and accompanying transport disruptions including flow and transport reversals. With there being few, if any, direct field observations of these disruptions based upon velocity measurements, the objective of this paper is to review the possible physical effects on these regions by first, reviewing the relevant contributing physics known about the Great Lakes; second, contrasting possible marine estuary transport mechanisms with what little is published about the Great Lakes circumstances; and third, summarizing modeled results exemplifying these behaviors from a study of Sandusky Bay, Lake Erie. Because it exhibits the strongest response to storms and the clearest measureable signals resulting from them, attention is centered on Lake Erie. In contrast to a typical research paper, the objective herein is to provide a summary of what is known and commonly accepted about these physics which can serve as a backdrop for the other papers in this special issue.*
INDEX WORDS: Wetlands, tributaries, estuaries, storm surges, seiches, mixing, transport.

INTRODUCTION

As clearly seen in the articles by Herdendorf and Mitsch (this Special Issue), a considerable number of the wetlands in the Great Lakes exist in regions which are associated with tributary confluences and relatively flat flood plain regions, both of which are susceptible to long and short-term water level fluctuations. Long-term water level fluctuation effects ($\sim$ years) on these regions are directly addressed in other papers in this issue. Short-term water level disruptions derive from storms and result in not only fluctuating volumes of water of different chemical and biological character being transported within the confluence and wetland regions but also a considerably altered rate at which the mixing and transformation of these constituents occurs. The objective of this paper is to review the physics of the Great Lakes with regard to how these episodic disturbances occur and how they might impact the confluence and wetland regions.

Several perspectives are adopted. First, this paper does not present the state-of-the-science of each aspect of each physical process. Rather, this article draws from existing published and commonly accepted information in a way which serves as a backdrop for the remaining papers in this Special Issue. Secondly, in the author's opinion, the physical analysis of tide-affected marine estuaries is far more advanced than the tributary confluence physics being addressed here. Consequently, the analogy to estuaries will be drawn whenever it is useful, especially in pointing the way to more useful future data collection. It is important to note that the confluence of these tributaries with Lake Erie is not an estuary as defined by Pritchard (1955), Cameron and Pritchard (1963), Hansen and Rattray (1966), Dyer (1973), Officer (1976), and McDowell and O'Conner (1977). There is no mixing of salt and no propagation of solar/lunar driven tide waves. Yet there are estuary transport analogies within Great Lakes tributaries and it will be the goal of this paper to indicate when analogous activity occurs and when it doesn't occur. The uniquely different transport mechanisms help to further frame the research issues. Finally, the most extreme examples of these physical interactions occur in Lake Erie, particularly in the western basin. Therefore, this paper concentrates its examples and discussions on Lake Erie as the field data provide the clearest and most unambiguous examples.

Much of the information in this paper is abstracted from specifically cited references; there are, however, a number of fundamental documents which by their comprehensive nature serve as source works. These include the "Project Hypo" report (Burns and Ross 1972), the special issue entitled "Lake Erie in the Early Seventies," (*J. Fish. Res. Board Can.* 33), and most recently and importantly, the Special Issue entitled "Lake Erie Binational Study," (*J. Great Lakes Res.* 13(4), edited by F. M. Boyce *et. al.*, 1987). Brandt and Herdendorf (1972) first suggested modern analogies to marine estuaries and this author, Herdendorf, and coworkers attempted a first effort (Bedford *et al.* 1983) at collating Lake Erie physical effects on tributary confluence transport as primarily summarized from the thesis work of D. Lindsay (1981). Finally, special mention should be made of the report by Herdendorf (1987) which was amongst the first comprehensive review of wetlands that made special mention of the possible impact of episodic Great Lakes physics.

THE LAKE ERIE/TRIBUTARY-WETLAND SETTING

Lake Erie itself, as noted in previous papers, contains three basins, the western, central, and eastern basins, (Fig. 1) separated by two sills. Also, from Figure 1, the islands separating the western and central basins also define two channels for hydraulic flow, the Pelee passage and the south passage. The lake is shallow, particularly in the western basin, which, along with the lake's southwest to northeast orientation, makes it particularly susceptible to storm activity. As will be seen, the lake response to storms is a controlling factor in confluence mixing and transport.

With the glaciation processes responsible for their creation being summarized in the Herdendorf article (this Special Issue), Figure 1 delineates the major tributaries entering Lake Erie. The numbers refer to the summary in Table 1 of tributary names and associated hydraulic data as extracted from Bedford *et al.* (1983), the United States Geological Survey Water Basin characteristic file, Brandt and Herdendorf (1970), and Herdendorf (1990). The last two columns in Table 1 are based upon the estimation procedures developed by Brandt and Herdendorf (1972) and Herdendorf (1990). Several items of note: 1.) Tributary No. 4 is the Sandusky Bay/River system which is completely anomalous in its structure compared to the other tributaries, 2.) Old Woman Creek, just east of the Huron River (No. 5), is an Estuarine Preserve but is not carried in the computerized USGS Water Basin file; and 3.) the Detroit River, also not listed in the USGS Water Basin file, is considered the major tributary to the lake.

The locations of the wetlands in Lake Erie have already been mapped, sized, and parameterized in the papers by Herdendorf (this Special Issue) and Mitsch (this Special Issue) and the reader is referred to them for additional details.

In general, the river slopes are quite small, especially in the western basin, and therefore, during non-storm periods, flows are quite small. Maximum tributary discharge occurs during and just after the spring "melt" season while a seasonal low flow occurs in the early fall. In general, the Detroit and Maumee rivers contribute the largest load of water volume and sediment mass to the lake; for example, Kemp *et al.* (1976) reported that almost three million metric tons of sediment are transported to lake Erie annually via these two rivers out of a total input (by mostly shore erosion) of 33 million metric tons.

Reversing flows have been frequently observed in all these rivers, even the Detroit River (Dereki and Quinn 1990), but much less frequently for those near the center of the lake, e.g., the Cuyahoga (Bedford *et al.* 1983). It is the delineation of this activity and its origins that provides the key to understanding the similarities and dissimilarities between Lake Erie tributaries and marine estuaries and their effects on wetlands.

LAKE ERIE PHYSICS—AN OVERVIEW

Scales of Activity and Variability

The singularly important feature of Lake Erie physics is its variability. It is extremely difficult, if not impossible, to identify physical characteristics of the Lake which are in some sense steady. The variability of Lake Erie extends over many decades of time and space with vertical spatial scales being quite compressed in contrast to the horizontal scales. Table 2 (from Boyce 1974, and Bedford and Adelrhman 1987) lists the physics and transport

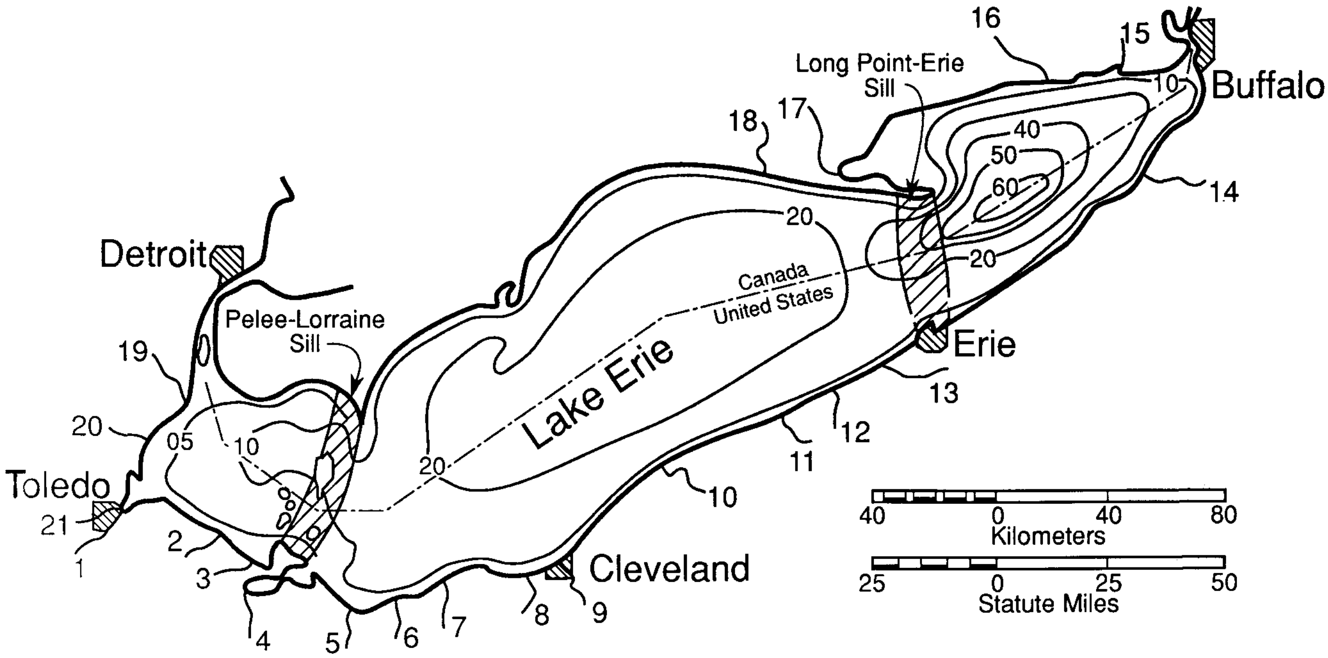

FIG. 1. Lake Erie bathymetry and tributary schematic.

mechanisms extant in the Great Lakes and Lake Erie. What makes Lake Erie such a difficult area to study is that many of the mechanisms over-lap in space/time structure and therefore it is very difficult to organize the controlled field programs necessary to validate hypotheses about the effect of one separate or individual transport agent in contrast to another.

Based upon the previous references, what is quite clear, however, is that several basic attributes of Lake Erie do occur which are repetitively observed and fundamental to the understanding of the lake. First, being so shallow, Lake Erie responds robustly to the annual thermal heating and cooling cycle (Schertzer 1987, and Schertzer *et al.* 1987). Second, in combination with the shallowness, thermal regime, and orientation, the lake quickly responds to the passage of storms, setting off an extremely complex variety of water surface and thermocline oscillations and corresponding velocities (forced responses). Third, surges, when they occur, can often be dramatic (e.g., Libicki and Bedford 1990); and fourth, free responses exist for some time after the disturbance in the form of seiches and internal waves. Finally, during the interdisturbance or interevent time, transport is quite weak and exists against a backdrop of hydraulic flow established between the Detroit and Maumee river inflows and the Niagra River outflow.

Interevent times are often less than the time required for the previous event's disturbance responses to die away; therefore, the lake is in a some-what continuous cycle of impulsive storm forcing and response die away followed by another storm. Hence, a steady state description or explanation of the gross lake behavior based on averages is an often misleading but, nevertheless, common practice. Therefore, the behavior of Lake Erie is summarized below, first, from a long-term average or ensemble average point of view, then followed by a discussion of the episodic nature or climate of the lake response to wind forcing.

Persistant Thermal and Circulation Regimes

From Project Hypo and the more recent works of Schertzer (1987) and Scherzter *et al.* (1987) much is known about the average thermal structure of the lake. During the winter, ice cover exists over 90% of the lake (Schertzer 1987) which, however, does not necessarily suppress wind driven circulation (Campbell *et al.* 1987). After the ice break-up wind mixing aids in distributing incident heat uniformly

TABLE 1. Mean flow and size data for the rivers indicated in Figure 1.

ID#	Name and State	Mean* Annual Flow (m^3/s)	Q* 7,10 Low Flow (m^3/s)	Q* 7,10 High Flow (m^3/s)	Mean* Monthly Low Flow (m^3/s)	Mean* Monthly High Flow (m^3/s)	Bed* Slope (m/km)	Estuary* Length (km)	Estuary* Area (km^2)
1.	Maumee, OH	12.6	2.6	1,545.1	16.9 (Sept.)	286.9 (March)	0.25	23.8	11.6
2.	Toussiant, OH	N/A	N/A	N/A	N/A	N/A	N/A	16.1	3.9
3.	Portage, OH	8.2	0.03	133.3	0.75 (Aug.)	17.4 (March)	0.56	25.2	10.4
4.	Sandusky, OH	27.6	0.3	400.0	5.5 (Aug./ Sept.)	69.9 (March)	0.8	24.8	5.3
5.	Huron, OH	8.5	0.13	124.2	1.3 (Oct.)	21.7 (March)	1.8	7.4	1.6
6.	Old Woman Creek, OH	N/A	N/A	N/A	N/A	N/A	N/A	2.1	0.3
7.	Vermillion, OH	7.4	0.00	137.2	0.9 (Oct.)	22.0 (March)	1.4	2.4	0.3
8.	Black, OH	9.2	0.1	123.1	2.0 (Aug.)	24.3 (March)	1.34	6.6	0.9
9.	Cuyahoga, OH	22.8	1.8	171.8	3.6 (Sept.)	48.4 (March)	1.4	7.2	1.0
10.	Grand, OH	18.8	0.03	191.8	3.0 (Aug.)	46.1 (March)	0.3	5.3	0.8
11.	Ashtubula, OH	4.4	0.00	46.4	0.7 (Aug.)	9.8 (March)	2.6	2.8	0.3
12.	Conneaut, OH	7.4	0.05	80.6	1.8 (Aug.)	160.0 (March)	1.4	2.0	0.2
13.	Elk, PA	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A
14.	Cattaragus, NY	20.8	1.9	2.0	6.4 (Aug.)	48.0 (March)	4.5	N/A	N/A
15.	Welland Canal, Canada	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A
16.	Grand, Canada	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A
17.	Big River, Canada	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A
18.	Big Otter, Canada	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A
19.	Huron, MI	11.5	1.1	76.9	4.1 (Aug.)	23.8 (April)	0.5	N/A	N/A
20.	Raisin, MI	18.3	1.1	199.4	4.0 (Sept.)	45.2 (March)	0.64	N/A	N/A
21.	Ottawa, OH	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A

(*) data source: USGS Water Basin Characteristics File
(+) data source: Brandt and Herdendorf (1972)

throughout the water column. This heating is relatively slow compared to the tributary water heating (Schroeder and Collier 1966, Fraleigh *et al.* 1975, Herdendorf and Zapotosky 1977, and Bedford *et al.* 1983). Furthermore, the deeper (24 m avg. depth) eastern basin will lag the shallower central (18 m avg. depth) and western basins (7.3 m avg. depth) in the response to heat input in the offshore deep water zones.

The spring-summer-fall heating/cooling cycle is marked by both vertical and horizontal temperature variation. The thermocline generally reaches its maximum vertical depth at from 15–20 meters in the central basin and 30–35 meters in the eastern basin. These depths are determined by a particular year's total heat and wind energy (mixing) input. Hypolimnetic waters generally have summertime temperatures ranging from 5°-10°C. The thickness of the

TABLE 2. Summary of Lake Erie processes and scales (Boyce 1974 and Bedford and Abdelrhman 1987).

A partial list of motions and their associated time and space scales. Letters in parentheses refer to the nature of the scale used. The governing terms in the equations of motion and continuity are listed in column 6 according to the key below.

(1) Phenomenon	(2) Length Scale (3) Horizontal	Vertical	(4) Time Scale	(5) Vel. Scale	(6) Dynamics Major Components
a. Wind driven surface gravitational waves	10 m(S)	1 m(M)	IS(P)	10 m/s(C)	1,2,6,10
b. Surface gravitational waves-seiches	100 km(S)	10 cm(M)	2–14 h(P)	2 cm/s(H)	1,6,9,10
c. Short freely propagating internal waves	10 km(S)	2 m(M)	5 min(P)	2 cm/s(H)	11,2,7,10
d. Long propagating internal waves	100 m(S)	2 m(M)	1 day(T)	50 cm/s(C)	1,5,7,9,10
e. Internal gravitational standing waves or seiches	10 km(S)	2 m(M)	15 h(P)	10 cm/s(H)	1,5,7,10
f. Surface wind drift	–	10 cm(S)	–	2 cm/s(H)	10,12
g. Coastal currents	10 km(S)	–	1 day(T)	10 cm/s(H)	all
h. Upselling and downwelling	10 km(S)	10 m(M)	1 day(T)	<1cm/s(V)	all
i. Wind driven horizontal circulation	100 km(S)	100 m(S)	1 day(T)	10 cm/s(H)	all
j. Geostrophic current	–	–	1 day(T)	3 cm/s(H)	5,6
k. Langmuir circulations vertical mixing of epilimnion	–	10 m(S)	1 h(T)	1 cm/s(V)	1,2,3,4,10,12,14 and others
l. Formation and decay	–	10–100 m(S)	1 mo(T)	–	10,12,14
m. Tidal waves					
a) diurnal (k1,0.,P.)	Earth radius	–	1 day(T)	–	1,2,5,15
b) semidurnal (M_2,S_2,N_2)	Earth radius	–	1/2 day(T)	–	1,2,5,15

M amplitude of motion;
S distance over which phenomenon varies significantly;
P period;
T time interval over which phenomenon varies significantly;
C wave speed;
V vertical particle velocity;
H horizontal particle velocity.

----

1 time-dependent horizontal accelerations;
2 time-dependent vertical accelerations;
3 advective component of horizontal acceleration;
4 advective component of vertical acceleration;
5 Coriolis force;
6 pressure gradient force due to slope of free surface;
7 pressure gradient force due to slope of the thermocline;
8 pressure gradient force due to atmospheric pressure field;
9 variations in bottom topography;
10 wind energy/stress;
11 internal stresses arising from horizontal current shear;
12 internal stresses arising from vertical current shear;
13 friction against boundaries;
14 potential energy changes due to surface heating/cooling;
15 astronomical tidal-generating forces due to sun-earth-moon gravitational potential field;
16 potential energy changes due to temperature and salinity changes.

mesolimnion is a function of the severity of wind mixing, with the deepest possible water column depth being achieved by mid-August (Schertzer 1987). Vertical profiles during the fall return to homogeneity through surface cooling and storm induced destabilization. By mid-October homogeneous profiles are noted. In the fall it is noted that the lake retains heat much longer than the overlaying

TABLE 3. Circulation transport regimes of Lake Erie and corresponding dominant physical components referenced from Table 2.

Location Regime	Total Lake	Inshore Zone	Still Zone	Western Basin	Central Basin	Eastern Basin
1. Ice Covered	m,j	g	N.A.	j	m,j	m,j
2. Homogeneous, Non Storm Condition	N.A.	h,g	h,i	j,i	i,j	j,i
3. Stratified, Non Storm Conditions	N.A.	g,h.l h,i	c,d	c,d i,j	c,d,e i,j	c,d,e i,j
4. Homogeneous Storm Conditions	N.A.	a,b g,h	h,b,i	b,i	i,j,b	i,j,b
5. Stratified, Storm Conditions	N.A. N.A.	a,b g,h	c,d,e i,h	b,i	c,d,e i,j	c,d,e i,j

atmosphere, thus setting up unstable air/water temperature differences (Schwab 1978).

A considerable body of knowledge of measured and modeled currents is available in the previously cited literature, and Table 3 (Bedford and Abdelrhman 1987) is an attempt to define the structure and classification of persistent circulation features. Model results are quite prevalent in verifying these regimes and of early note are the initial numerical model studies of Gedney and Lick (1972). Full three-dimensional model results, assuming no vertical acceleration (Haq and Lick 1975) have been performed for nonstratified conditions, but a three-dimensional simulation of Lake Erie including full stratification effects has yet to be comprehensively performed. Lam and Simons' (1976) two layer approach has provided insight to flow dynamics during stratification.

Field observations of currents have not been as fully resolved as those observed in Lake Ontario during the International Field Year on the Great Lakes, however, good data do exist starting with Verber (1955). A review of the history of current data collection is found in the papers by Mortimer (1963) and Saylor and Miller (1987). The Saylor and Miller article, as well as the article by Boyce and Chiocchio (1987), provide a wealth of data about the persistent vertical and horizontal currents.

The essential nature of persistent circulation features is as a result of frequent wind impulses which relatively quickly (Haq and Lick 1975) spin the lake up to a new circulation state. Saylor and Miller (1987) stated that surface currents are a result of wind drift, while currents near the bottom are pressure gradient driven return flow. They went on to state that a quite complex suite of circulation gyres occurs whose simpler forms have been predicted by models (Saylor *et al.* 1980) and observationally confirmed by Saylor and Miller (1987). It is noted that the gyres are often unstable with one or the other becoming dominant in the central basin. Such dominance is possible as a result of spatial wind stress gradients. Of some note from the Saylor and Miller article is that stratification does not appear to be important in affecting current and gyre structures during episodes. Currents in the passages are a result of seiche activity; with western basin currents, in general, dominated by surface wind stress and the hydraulic influence of the Detroit and Maumee rivers. Finally, as noted by Saylor and Miller (1987), in the nearshore zone a number of investigators have found thin bands of eastward flow in both northern and southern shores, and have noted the familiar Coriolis affected wind driven surface drift currents. The new evidence in Saylor and Miller (1987) does not change this picture.

THE EPISODIC NATURE OF LAKE ERIE

As mentioned, Lake Erie physics are particularly susceptible or responsive to storms. Due primarily to prevailing storm tracks (Irish and Platzman 1962) and shallowness, Lake Erie responds to the wind stress by a combination of free and forced mode oscillatory responses in water level and thermocline position which give rise to periodic velocity and current structures.

Forced Mode Response

The response of Lake Erie to the imposed wind stress as it is in contact with the lake (i.e., before the passage of the storm) is called the forced mode. The most dramatic manifestations of this response are frequent and often dramatic storm surges. Water level increases occur downwind in the lake with commensurate water level drawn down at the upwind end. Therefore, for winds from the southwest, water level increases at Buffalo will occur with drawdowns at Toledo. The opposite will occur for winds from the northeast. The Buffalo minus Toledo water level is called the set-up, although this is a bit misleading in that the maximum positive surge amplitude above datum will occur some 3 hours or more before the corresponding maximum drawdown. Explanations for this vary, but have been offered by Rao (1967), Hamblin (1979), and Libicki and Bedford (1990) via the method of characteristics.

The statistics of surges are impressive and, following the probability analyses of Irish and Platzman (1962) and Pore *et al.* (1975), the following information is known (See Figs. 2a-2c). The water level at Buffalo exceeds the mean monthly water level by 3.5 feet at least once per year, while the same statistics hold for water drawdown at Toledo. Therefore, a maximum set-up of 7 feet is exceeded roughly once per year. A set-up in excess of 10 feet is equaled or exceeded once every 2 years. The worst positive surge at Buffalo was 9.0 ft. above the monthly average water level during the April 1979 storm (Hamblin 1979), while the worst positive surge in Toledo was 5.3 ft. above the monthly average water level during April, 1967. The worst negative surge at Toledo was 7.5 ft. in March, 1965, while the worst negative surge in Buffalo was 4.65 ft. during March, 1964. Extreme value/recurrence probabilities are presented in Figure 2a. Commercial navigation is prohibited during extreme drawdowns in Toledo and freshwater intakes cannot function; therefore, surge forecasts are issued as necessary by the National Weather Service as based upon the model of Schwab (1978).

The origins of such storm surges are atmospheric pressure drops and wind shear stresses acting at the lake's surface. Through the effect of large pressure drops and corresponding pressure gradients, a suction lift effect derived from hydrostatic pressure considerations is hypothesized to be one mechanism for creating surges. Harris (1957) examined such an occurrence on Lake Michigan.

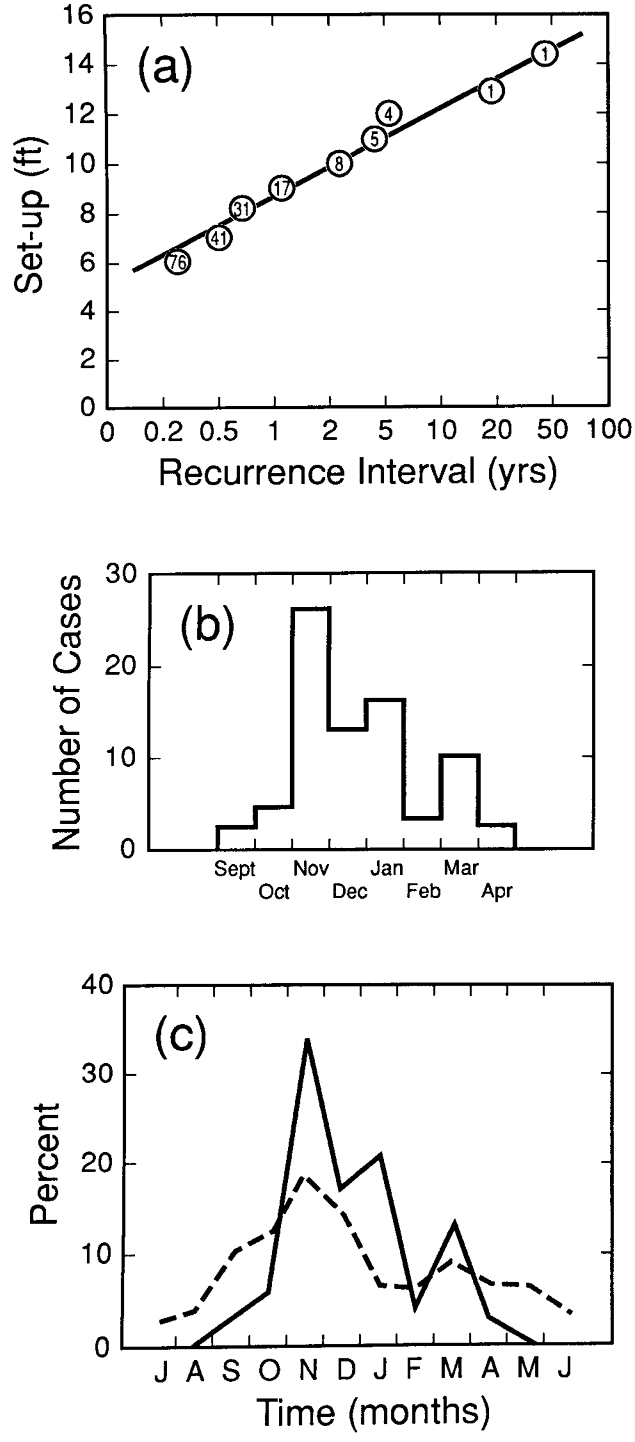

FIG. 2. Storm surge statistical features (redrawn from Pore et al. *1975) a. recurrence intervals (#'s of events in circle) for Buffalo-Toledo setups, b. histogram of monthly setups in excess of six feet, and c. frequency distribution of six foot setups or greater (-) and severe storms (---).*

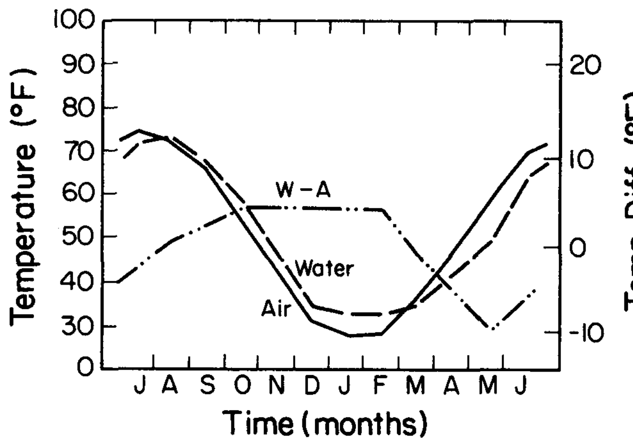

FIG. 3. Monthly air-water temperatures for Lake Erie (redrawn from Pore et al. *1975).*

However, the effect is small, for most large weather systems as the pressure gradients are small owing to the size of the storm over which the pressure drop occurs. For example, Dingman and Bedford (1984) analyzed the Lake Erie response to the January 1978 cyclone which had the lowest central pressure recorded for an extra-tropical cyclone observed over the lake. The water level elevation increase due to the suction lift was 30 cm at most.

Shear stress is by far the most dominant mechanism in the creation of a storm surge and two facets of the application of this shear stress to the lake surface are critical in the development of large surge events; an air-water temperature instability and possible stress-band resonant coupling.

Originally discussed by Hunt (1958), it has since been recognized that the shear stress at the water surface is not only a function of the wind speed, but the air-water temperature difference as well. While Hunt's analysis was primarily empirical, Schwab's (1978) stability dependant formulation was based on well founded boundary layer methodologies and is now a commonly used procedure for making shear stress estimates. It should be noted that the drag coefficients for thermally unstable conditions are nearly twice as high as those for stable thermal regimes.

This then explains to a large extent why the vast majority of very large surges occur during the late fall months (Figs. 2b and 3). The fall months are times when heat is retained in the lake due to its slower heat conductivity relative to the atmosphere, and the passage of cold fronts can cause air quite colder than the lake to pass over the surface. In the spring, the lakes take longer to acquire heat and so the air is usually warmer; therefore, instability plays less of a role in creating springtime surges.

Another factor in the creation of excessive surge heights is resonance. Resonance conditions have been defined in various ways and are hypothesized to result from bands of low pressure or high shear stress traveling across the lake with a speed equal to the free gravity wave speed for the lake. This idea was suggested early on for pressure disturbances by Hunt (1958) to explain an anomalous surge in Lake Michigan arising from a convective storm. Significant analytical attention has more correctly concentrated on the role of wind stress bands passing over the lake. Initial attempts by Irish and Platzman (1962) to address this problem were somewhat flawed as they concentrated on defining the coupling through the position of the front in the weather system. The front is an ambiguous tracking measure. It remained for Rao (1967) to identify the proper tracking measure as the wind stress band and, noting that large pulses of stress can occur through either large winds or an air water temperature instability, defined resonance for Lake Erie as follows. For wind systems whose stress band width is less than the lake's length, resonance occurs if the propagation speed of the stress band approximately equals the gravity wave speed. For Lake Erie this would be approximately 13.5 meters per second or 48.5 km/hr (30 miles per hour). For stress bands with a width longer than the lake, resonance occurs when the time it takes the stress band to pass a stationary point equals the time necessary for a gravity wave to pass the length of the lake. The resonance condition is frequently equaled because weather systems often travel with roughly that same speed.

Other factors have recently been examined; extending Rao's (1967) work by use of more sophisticated computational algorithms, Libicki and Bedford (1990) demonstrated that the converging geometry of the shoreline near Buffalo was a far more important contributor in creating high surges than previously acknowledged. Pure resonance does occur and is marked by the presence of a high localized *flip* of the water surface near Buffalo. This flip is a brief localized pulse of water surface elevation which increases the water level 1 to 2 ft. more than the elevation due to non-resonant geometry focused surges. Of the five worst surges in

Lake Erie, this flip was observed only once (in the April 1979 storm) and resulted in the highest amplitude ever recorded by over 1 foot. This storm surge was a resonant response; the other four worst surges were not resonant responses.

Free Mode Response

After passage of the stress event, the potential energy stored in the surge is released and expressed as free oscillation gravity waves called seiches. The entire basin is activated by seiche activity. Following the lead of Csanady (1982), Csanady and Scott (1974), and Mortimer (1963) coastal jets and internal (baroclinic) and surface (barotropic) Kelvin waves can occur as well. Unlike the IFYGL Lake Ontario work, very little work has been done in Lake Erie in fully exploring this resulting nearshore activity. Since coastal jets and Kelvin waves might affect how tributary discharges develop, more research will be required in this area.

The structure of seiche activity is widely known from observations made at the nine water level gauges surrounding the lake and a number of one- and two-dimensional computational experiments. With the reviews by Mortimer (1963) and Hamblin (1987) providing the historical context, the following features are known. There are five persistantly observed longitudinal free modes in Lake Erie and, as the initial modeling and spectral analyses of Platzman and Rao (1964a,b) showed, the basin-wide average periods are 14.1, 8.9, 5.7, 4.1, and 3.7 hours. A transverse mode was numerically speculated upon by Platzman (1963) and observed for the first time in the data by Dingman and Bedford (1984). The structure of these modes, i.e., positioning of null oscillation lines, timing and phasing, volume transports etc., has been extensively researched. It has been initially shown by Platzman and Rao (1964a,b) that the seiches contain a counter-clockwise progression of high water which is introduced by Coriolis activity. This amphidromic structure then results in phase lags between high water occurring at different points around the lake. The first four amphidromic mode structures and phase relations are shown in Figure 4 as redrawn from Hamblin (1987). These structures are calculated from model results.

Platzman and Rao (1964b) used a one-dimensional analysis to define typical water surface elevation, volume transports, and cross section average current speeds for the longitudinal modes. The one-dimensional cross-section average approach is particularly effective in summarizing the gross behavior of the seiche modes and Figures 5a,b contain these behaviors as redrawn from the 1964b reference. For clarity of presentation, the authors have given each mode a unit elevation at Buffalo. Spectral analyses published by numerous authors clearly show that each successively higher mode has a smaller elevation contribution. For instance, at Toledo the mode elevations referenced to a unit elevation for the fundamental oscillation are in the ratio of 0.8, 0.72, 0.47, and 0.40 for the second through the fifth modes. Implicit in these diagrams are several simple features of seiche activity. First, the relative amplitude of each successively higher mode is less than the fundamental mode, however, different amplitude modes are dominant in different parts of the lake. For instance, the first, third and fifth modes are close to null east of Cleveland, while the maximum second mode amplitude excursion occurs at this point. By implication then, different tributaries along the lake will be subjected to water level fluctuations of different period and amplitude, i.e., the second mode dominates the spectra of Cuyahoga River (No. 9, Fig. 1) elevation while the first mode dominates at the Maumee River (No. 1, Fig. 1).

A second feature is that water volume transports are generally maximum at null oscillation lines. Water volume transports, therefore, will also be different at different positions on the lake. From Figure 5b it can be see that the first mode maximum transport occurs near Cleveland/Fairport, while the second mode maximum transport occurs near the Huron River (No. 5, Fig. 1) and Erie, Pennsylvania. It is important to remember these transports are, in general, oscillating/periodic as are the elevations but are in general 90° out of phase with the elevation. The concept of maximum seiche excursion length (Sorenson 1978) associated with each mass support mode is applicable but has yet to be performed in detail for the Great Lakes with regard to tributary influences.

From inspection of Figures 4 and 5a it is quite clear that, during and after storms, oscillatory positive and negative water levels relative to the still water level will occur at various wetland and tributary sites around the Basin. For example, Toledo and Buffalo sites will be dominated by the 14-hour fundamental water level oscillation. These oscillations are severe enough to cause not only flood waves upstream, but flow reversals in the Maumee River (Pinsak and Meyer 1976) and the Detroit

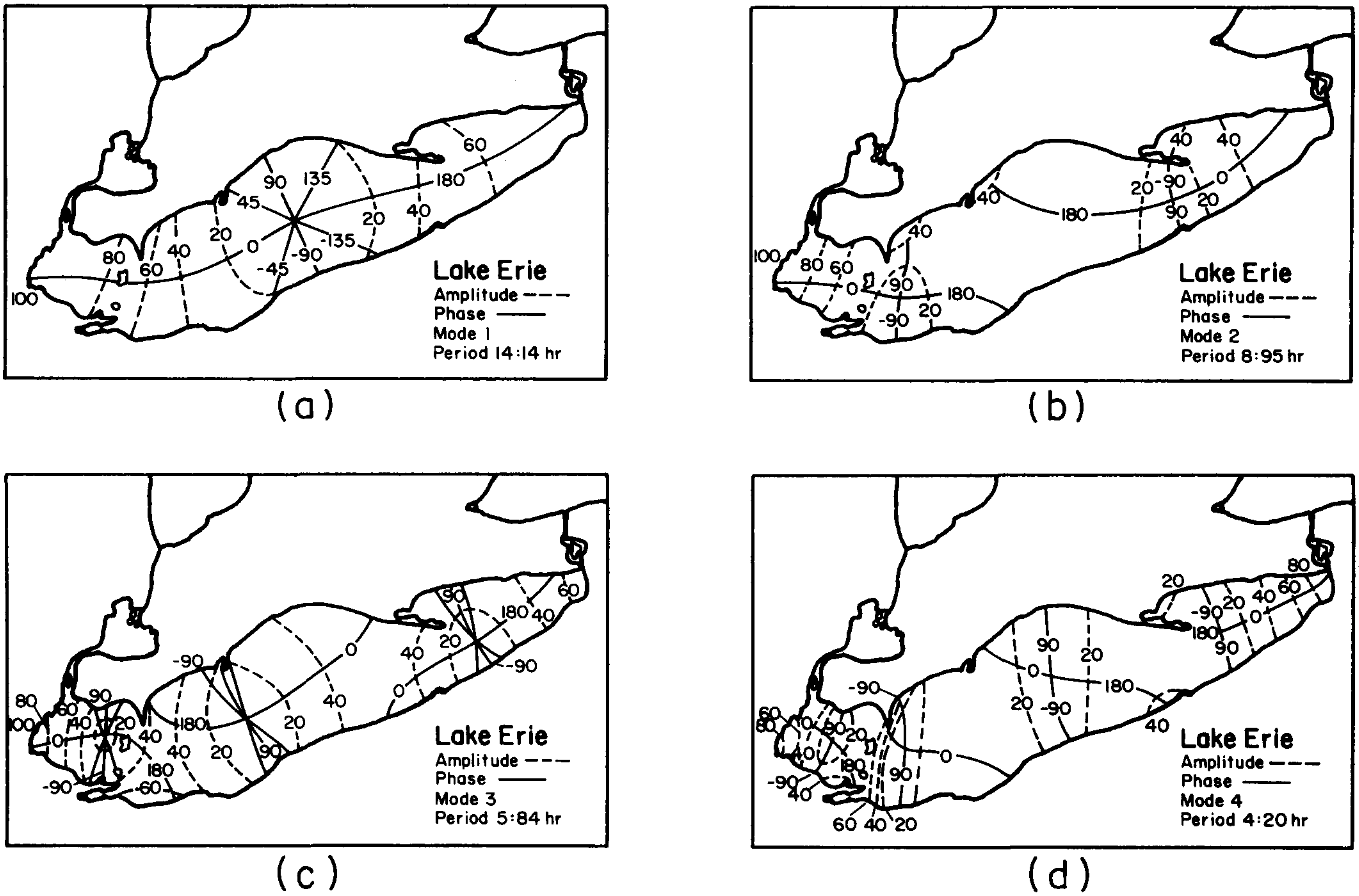

FIG. 4. First four Lake Erie free mode structures (redrawn from Hamblin 1987).

River (Derecki and Quinn 1990). The Cleveland/Cuyahoga area being near the maximum amplitude excursion of the second mode will see water level oscillations with 9-hour periods. In essence then, certain storms, by unleashing slow long-period gravity waves in Lake Erie, give rise to periodic flood wave phenomena in certain tributaries. These flood waves are often strong enough to cause partial or complete flow reversals. Since the decay modulus (Platzman and Rao 1964b) for these seiches indicates an exponential decay of amplitude maxima, it is estimated that over 3 days of fundamental mode oscillation are required for full diminishment. Often, in the spring and fall, storms arrive more frequently and therefore seiche activity persists. These seasons are expected to contain the highest probability for the episodic wetland and tributary transport disruption due to waterlevel fluctuations, flow reversals, or flood wave propagation.

Periodic waterlevel oscillations and flow reversals suggest that contrasts with estuaries are in order so as to learn of possible transport disruption mechanisms.

A SUMMARY OF ESTUARY TRANSPORT ATTRIBUTES AND A FIRST COMPARISON WITH LAKE ERIE TRIBUTARIES

From the early estuary overview works of Cameron and Pritchard (1963), Hansen and Rattray (1966), Dyer (1973, 1986), Officer (1976), and McDowell and O'Conner (1977) several features of estuary transport physics serves as a starting point for selecting similarities and dissimilarities. First, estuarine density variations derive from salt with

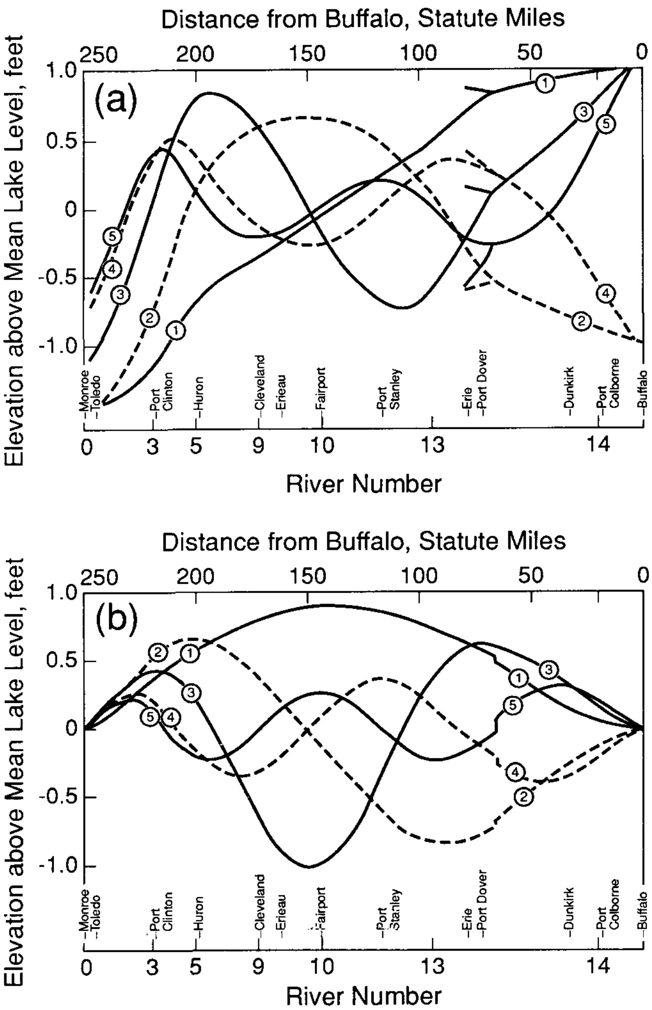

FIG. 5. Normalized Lake Erie water levels (a) and water volume transports (b) (redrawn from Platzman and Rao 1964b).

concentrations varying along the estuary from the freshwater head (zero concentration) to the ocean receiving water (concentration is maximum). The longitudinal variation can have a low gradient such as when freshwater inflow and tidal velocity might be weak or can have a gradient so high as to give rise to a density interface. Dyer (1988, 1986) and Pritchard (1955) presented schematics of the various types of density structures that can exist, and although one may argue the merits of the details, it certainly is the case that such density gradients exist and are permanent features of estuaries. The sharpness of the vertical and/or horizontal density gradient determines to what extent internal waves exist on the interface. Finally, depending on the severity of the freshwater discharge contrasted to the tidal flood or ebb momentum, the saltwater density interface may move up or down the estuary and attain a variety of positions. Figures 6a-6d (after Dyer 1973) are schematics of the possible density structure of estuaries where dense water results from salinity and fresh water is assumed lighter than salt water.

The velocity structure of estuaries has also been schematized in the early works of Pritchard (1955) and Dyer (1973). Dyer's ideas also appear in Figure 6 and even in these highly schematized portrayals, the simple interactions between freshwater discharge, tidal velocity, density structure, and to some extent bathymetry reveal quite complicated flow structures. For the most part, the most intuitive structure is the flow reversal introduced periodically by the tide. It is persistent for marine estuaries and results in considerable longitudinal and vertical differences in velocity. Indeed, as the schematics suggest, it is quite possible to have flows going in opposite directions at the same estuary position. Such intense flow gradients result in shear, a primary source of energy, mixing, entrainment, and sediment resuspension. Field examples have been in the literature for some time (Wright and Coleman 1974) and clearly document these schematics. Field studies of even the most elementary nature such as these early marine velocity observations have yet to take place in the Great Lakes tributary confluences.

Dyer (1973) also presented a schematic for estuarine flow over a sill or bar. The field evidence in Wright and Coleman (1974) concentrates on this feature and shows the cross-sectional density profile as the tributary enters the Gulf of Mexico during flood and ebb tides. Internal waves are clearly observed when the Froude Number is greater than one. Four regions of distinct hydrodynamic activity are noticed in the transition from channel to receiving water. Since sills and bars exist in wetlands and river mouths along the lakes, these features clearly suggest as yet unresearched processes affecting transport and exchange.

With these simple schematic outlines of estuary physics the question becomes whether there are any analogs to the Lake Erie tributary physics and if so, is there observational evidence upon which to confirm these analogies.

With regard to the density structure, Lake Erie tributaries do not contain salt in sufficient concentration to affect the flow structure. However, given the freshwater density versus temperature curve,

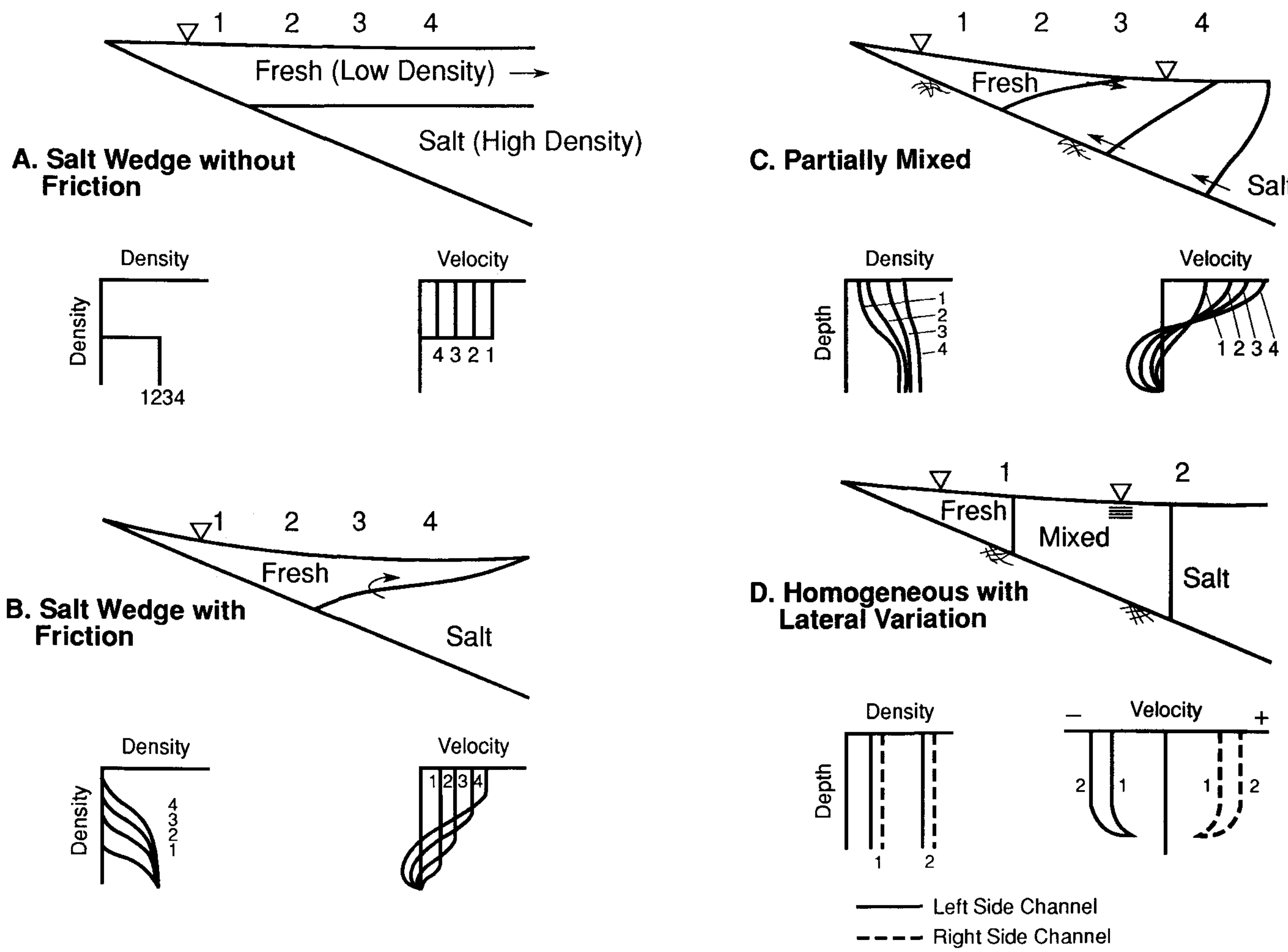

FIG. 6. Schematics of four marine estuary density and velocity structures defined from Dyer (1973).

and its maximum at 4°C, the effect of thermally induced density differences is quite possibly pronounced. This is particularly the case when tributary waters, being generally shallower, heat and cool more rapidly than Lake Erie water during spring and fall seasons, respectively. The possible consequences are as follows. Starting with water at 0°C during ice cover, the tributaries would warm more quickly and would reach density maximum more quickly. Therefore, for a brief time in the spring, tributary waters would sink below lake waters at the confluence only to have this reversed with further heating of the tributary waters and lake's waters having reached their density maximum later on. It is hypothesized that tributary waters would then be less dense and more readily flow over the lake water. This effect would obviously diminish as heating reached equilibrium during early summer.

Somewhat the reverse situation could occur during the fall; tributary waters losing heat more rapidly would be more dense and flow under the lake's water at the confluence. Again, a brief interchange could occur if the tributary waters go through density maxima before the lake's waters. Indeed, such an annual cycle of density structures could be considerably more complex than the estuary case and far more ephemeral as it is driven by meteorological and climatic circumstances. In contrast, the salt content in the ocean water is fixed and doesn't vary seasonally according to the weather. Both the estuary and confluence density structure are similar, in that freshwater inflow at the head plays a strong role in determining the position of the interface longitudinally in the channel and the degree of tilt in the interface. Observational evidence is summarized in the next section.

Insofar as the water surface elevation changes are concerned, it is clear that long waves propagate up and down the Lake Erie rivers; the period and mode structure is quite complex in that up to five long wave modes with periods as small as 3 hours may be propagated within the tributary. These wave events result from surges and seiches which are weather dependant and not persistent. Tide waves have a clear frequency and amplitude structure which is modulated by solar and lunar activity with periods of longer than one-half day, therefore, aperiodic pulses are less frequent or complex in estuaries. Unlike Great Lakes seiche waves, however, tide waves are persistent.

In the lake's tributaries, it is often assumed that seiche-induced flood waves propagate upstream without disrupting the lakeward tributary flow. This would especially be the case for non-stratified conditions where the tributary discharge momentum was more substantial than the weak channel flow induced by a weak flow floodwave. In contrast, strong flood waves resulting from storm surges could cause complete flow reversals in the tributary. Again, these occurrences are climate dependant, episodic, and are not persistent.

With these simple contrasts in mind, can any observational evidence be cited to substantiate the hypothesized tributary-Lake Erie interactions?

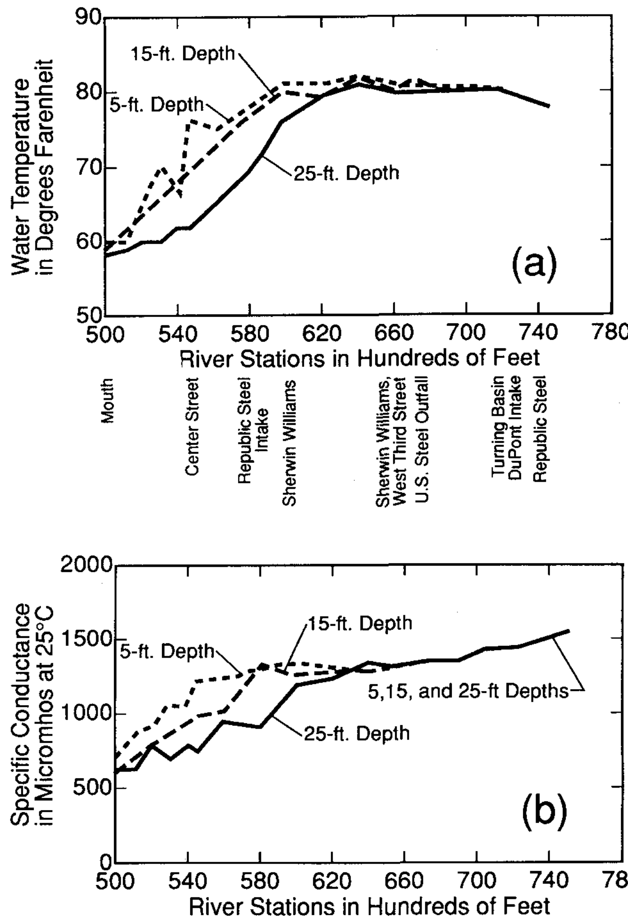

FIG. 7. Water temperature (a) and specific conductance (b) versus station in the Cuyahoga River in three depths, 21 October 1964 (redrawn from Schroeder and Collier 1966).

OBSERVATIONAL EVIDENCE

Observational evidence of these activities is neither coherent, comprehensive, nor thorough; particularly as regards circulation and velocity features. Temperature data, which are easily measured, exist and indicate that certain of the suggested density features do exist. Schroeder and Collier (1966) presented temperature data for the Cuyohoga River (Figs. 7a,b) which show the downstream variation of temperature for a fall condition (21 October 1964). The graphs do suggest compliance with the hypothesized variation. The Maumee River bay and lake interaction has been thoroughly investigated for dredging activities (Fraleigh *et al.* 1975) and from an International Joint Commission initiated study (Pollution from Land Use Reference Group, PLUARG) on how the winter melt affected the Lake Erie nearshore region (Pinsak and Meyer 1976, Herdendorf and Zapotosky 1977, for example). Figure 8 is a plot from Fraleigh *et al.* (1975) showing the seasonal variation in water temperature difference. Clearly, the Ottawa River/Lake Erie temperature difference behavior is as suggested earlier, but the Maumee River behavior is not consistent with the hypothesis. It should be noted, however, that in the Maumee River the lake effect intrudes to over 20 km upstream and since the Maumee River sampling point was well within this Lake Erie/confluence mixing zone, the Maumee River data are not representative of the freshwater tributary inflow temperature.

Detailed evidence abounds on the tributary water level disturbances in Lake Erie, most of it being anecdotal until the early work of Brandt and Her-

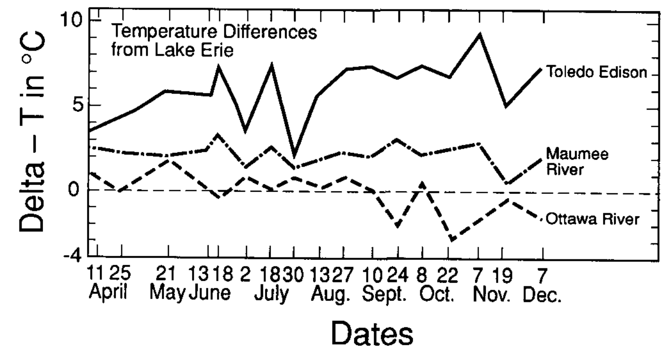

FIG. 8. Maumee Bay (Ottawa River) versus Lake Erie temperature difference for the year 1974 (redrawn from Fraleigh et al. *1975).*

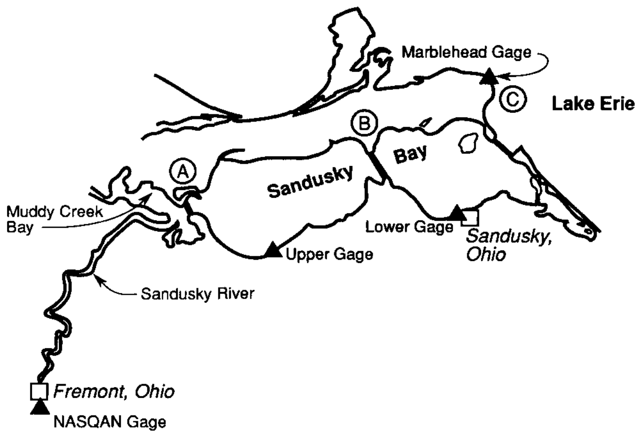

FIG. 9. Sandusky Bay site and transect location schematic.

dendorf (1972). Detailed probabilities of each tributary's water level structure and response to surges have not been done. Yet, the fact remains that the tributary water level disturbances were well enough known that USGS put their operational discharge and water quality gauges enough upstream to be away from the flood wave/flow reversing behavior. In the case of the Maumee and Sandusky rivers, these gauges are over 30 kilometers away from the lake's edge.

Velocity/current data in these confluence and mixing zones do not exist. Anecdotal information is all that is available, however, Pinsak and Meyer (1976) presented a flow persistence diagram for the Maumee River. It is indeed disappointing that velocity data are not to this day routinely collected simultaneously with the ubiquitously collected water quality data.

SANDUSKY RIVER/BAY – A COMPLEX TRIBUTARY

The Sandusky River/Bay tributary system, a site of considerable controlled and uncontrolled wetland acreage, has been extensively sampled from a water quality point of view and recently enough precision was obtained in the data to allow a coupled hydrodynamic and transport model of velocity, chlorides, and sediments to be verified. As a result, several of the hypothesized transport issues can, at least for limited time periods of data, be substantively addressed. Some of these results are presented here as examples of the possible tributary complexities. The work referred to here is principally based on the author's modeling as found in Lee and Bedford (1987a,b), Bedford *et al.* (1988), Bedford and Mark (1988), and Lyons *et al.* (1988). These models are predicated on the data collected by P. Richards and D. Baker at Heidelberg College (Richards and Baker 1982). It is unfortunate that the Sandusky Bay confluence is somewhat anomalous in contrast to all of the other tributary confluences in Lake Erie (or most of the rest of the Great Lakes for that matter). With reference to Figure 9, the bay is shallow, on the average being 2–7 meters in depth; only the presence of the shipping channel at the confluence allows depths to attain 7 meters or more. The bay is subjected to wind waves and wind-driven circulation. It is located at the null oscillation line of the second Lake Erie seiche mode (Figs. 4, 5a) and itself contains two free oscillation modes (Prater and Bedford 1982) with the first being 3.0 hours and the second being 1.7 hours. It is noted in Figure 9 that the USGS loading data gauge is in Fremont, Ohio, some 29 km upstream of the confluence on the perceived Lake Erie effect.

The model detailed in the above references was a combined two-dimensional, free surface nonlinear hydrodynamic and transport model which incorporated the effects of erosion and deposition from a multi-layer bed, wind waves including air-water temperature effects, time varying flow from the Sandusky River, and Lake Erie water levels. The time and space variations of currents, water levels, and chloride, sand, salt, and clay concentrations were predicted.

The conditions simulated (Figs. 10a-c) included two storm induced river discharge events and a 6-foot Buffalo/Toledo set-up Lake Erie surge event. As can be seen, the water levels varied continu-

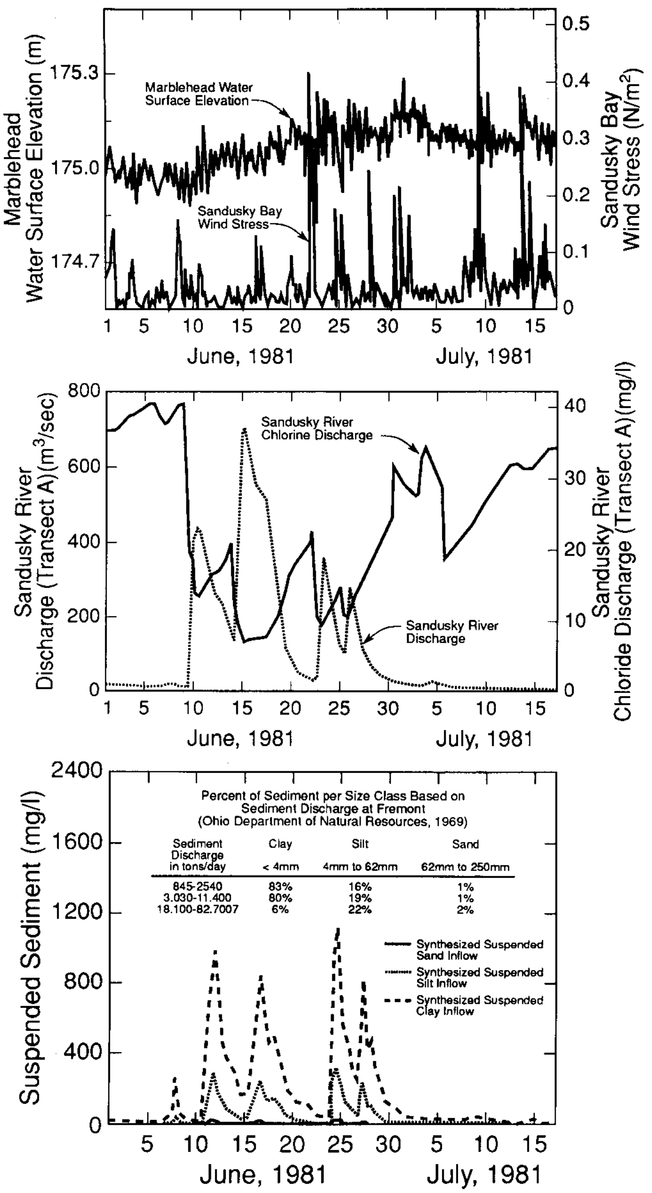

FIG. 10. Sandusky Bay simulation input data 1 June–15 July 1981; a. water level and wind shear time series; b. Sandusky River discharge and specific conductance inflow time series; and c. Sandusky River silt, sand, and clay mass inflow time series.

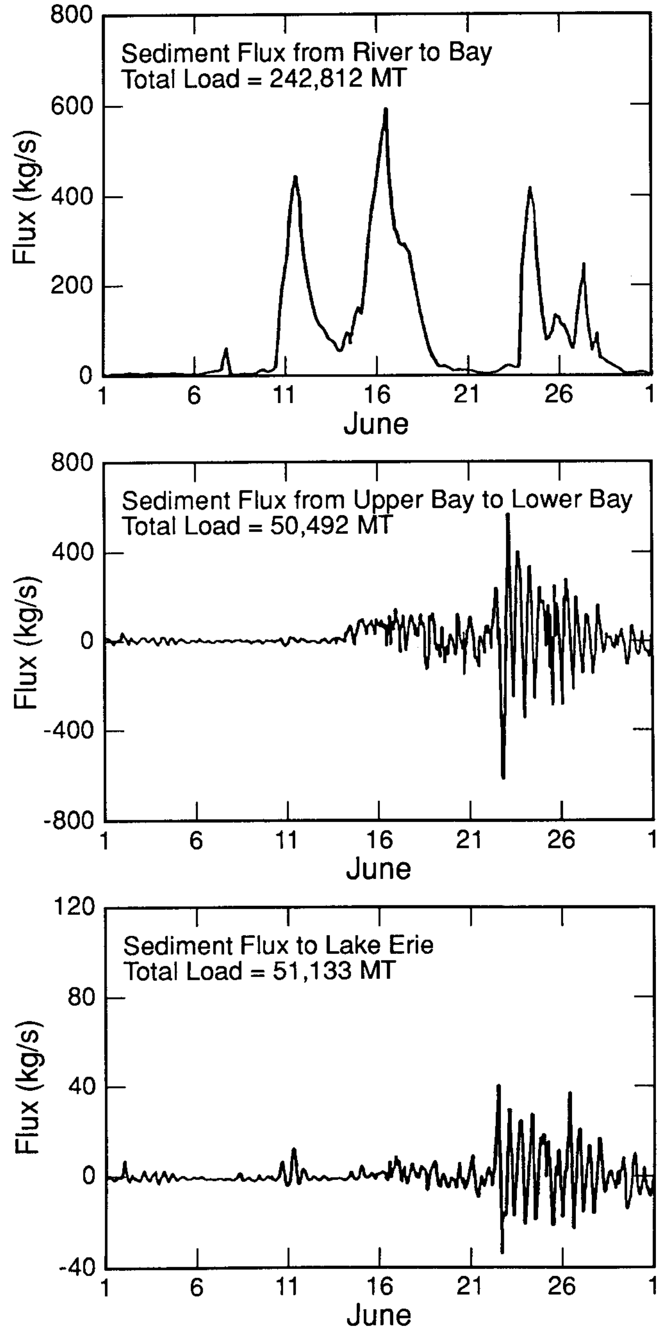

FIG. 11. Sandiusky Bay transect mass flux time series.

ously at Marblehead throughout the entire simulation period, 1 June-16 July 1981.

From the results presented in the above papers, a few directly speak to the hypotheses suggested here. The most important aspect of all is that Lake Erie water level fluctuations occurring at the confluence never die away and propagate disturbances into the bay. Figure 11 is a plot of sediment fluxes (mass/area/time) at the three transects and it is quite clear that the tributary confluence, Transect C, is subject to oscillating fluxes where material alternately enters and leaves the bay. Indeed, some of these events are weak flow reversals which result in daily average fluxes being entirely into the bay

TABLE 4. Total sediment load passing each transect in Sandusky Bay for period 1 June–30 June 1981.

Transect (Fig. 9)	Total (Metric Tons, MT)	Sand (MT)	Silt (MT)	Clay (MT)
C	51,133	13,659 (26.7%)	6,047 (11.8%)	31,427 (61.5%)
B	50,494	-573 (-1.1%)	1,616 (3.2%)	49,450 (97.9%)
A	242,812	8,200 (3.4%)	57,213 (23.6%)	177,399 (23.0%)
Fremont	372,194			

for the lake (Bedford and Mark 1988). It is also noted that the fluxes vary periodically at Transect B as well, while Transect A suffers no such lake-induced reversal.

A second general area of similarity is in the sorting, deposition, and scouring behavior of the sediments. Based upon the model simulated results in Lee and Bedford (1987a,b), Table 4 summarizes the total deposition of sediments in each size class in the upper and lower basins as well as the total flux past each transect during the simulation. It is noticed that the upper bay, which is not affected by Lake Erie seiches to any extent, is a net deposition zone for clays with extrapolated annual accumulation rates on the order of 4 centimeters per year. The western portion of the lower basin continues to be a net clay deposition zone, but in general the entire lower basin just transmits material through from Transect B to Transect C. The seiching effect of the lake, coupled with the bay's natural oscillations, keeps the material generally in suspension. A significant sand deposit occurs in the lower basin east of the Johnson Island-Sandusky second mode null oscillation (Fig. 12) and is hypothesized to exist because of sand in the Lake Erie littoral drift near the confluence being brought into the bay during flow reversals and surges as well as the seiching which prohibits the settling of clay and silt particles in the confluence. This behavior is quite consistent with the recent summaries of estuaries' sediment transport and hysteresis effects by Dyer (1988, 1986).

Finally, estuaries have turbidity maxima or interfaces that are arranged across the channel which are high concentration bands of sediment and it appears that there is one in Sandusky Bay as well (Bedford 1989). A Landsat image, developed from June 1981 data by Lyons *et al.* (1988), clearly shows this interface and its close correspondence with the null oscillation line. It should be noted that the available data permit no further hypothesis testing about this interface but it is more than coincidence that it occurs where it does, not only because the null oscillation line for the Sandusky mode is the point of maximum velocity energy which is required for entrainment, but perhaps more importantly, it is the maximum in-bay extent of the horizontal transport of the second Lake Erie mode, a 9-hour mode with a 2.5 km half excursion distance which represents a primary source of entrainment and mixing energy for the sediments.

FINAL REMARKS

While much of the information in this paper is well known, there are several points to be noted. First, the susceptibility of the shallower Great Lakes to storms introduces not only the intuitive widely observed water level changes but also possibly important changes in the flux of material being carried in the water column. It is anticipated that not only the flux magnitudes will possibly change over these transient disruptions, but also the flux direction. Second, as attention is increasingly focused on shorter term disruptions, the corresponding spatial reorganization of water masses of quite different character will involve the location, measurement, and tracking of interfacial conditions. Third, the analogy with marine estuaries is correct in suggesting that the flood and ebb conditions are asymmetric in that the concentration and fluxes during flood are quite different than those during the subsequent ebb. Fourth, the analogy is not correct in that the differential heating enhances springtime interfacial development which potentially reverses during the fall cooling season. Finally, the lack of

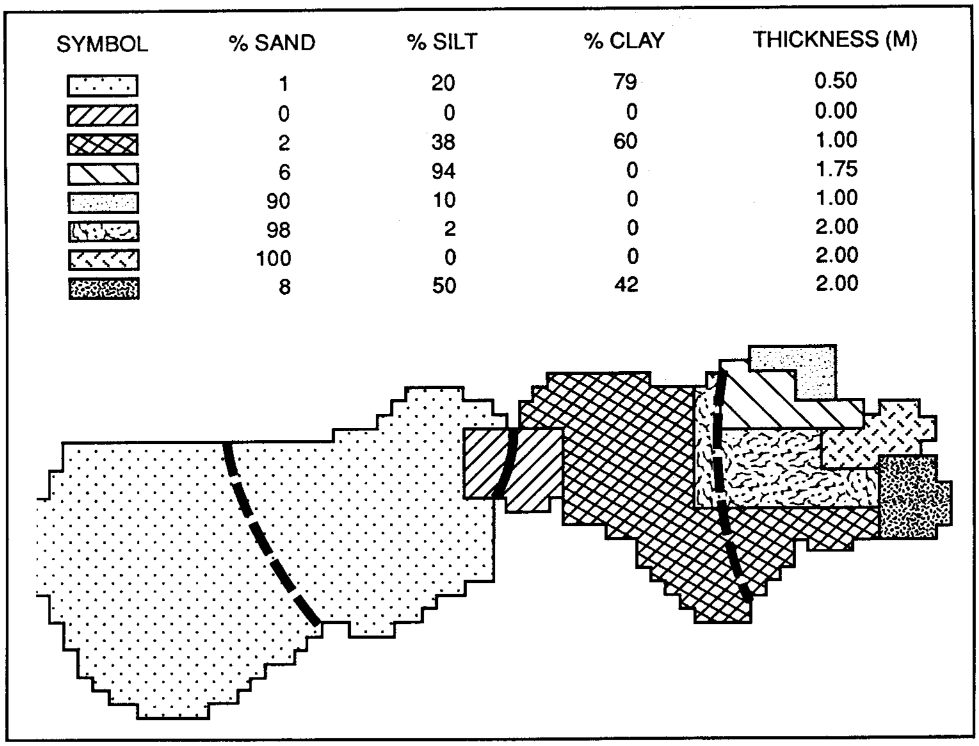

FIG. 12. Bottom sediment distribution and null oscillation location schematic (-,3.5 hr, S1 mode; --- 1.7 hr, S2 mode).

velocity field measurements in these regions prohibits any flux estimates from being robustly quantified and adaptations of useful estuary concepts from being attained. In this author's opinion, remedying the lack of velocity and corresponding flux data in these regions should be a primary management goal.

ACKNOWLEDGMENTS

The author very much appreciates the work of past graduate students D. Lindsay, J. Yen, D. Mark, and D. Lee, whose theses have formed the basis for much of this article. The author sincerely appreciates the support of Ohio Sea Grant for original preparation of this document. Final preparation was supported by the Great Lakes Forecasting System through grants from CRAY Res. Inc., USGS Water Res. Center, NOAA Coastal Ocean Program, and NOS.

REFERENCES

Bedford, K. 1989. Storms and the occurrence of a turbidity interface/maxima in a freshwater estuary. In *Sediment Transport Modeling,* ed. S. Wang, New York: pp. 106–111. Amer. Soc. of Civil Engrs.

________, and Abdelrhman, M. 1987. Analytical and experimental studies of the bottom boundary layer and their applicability to near-bottom transport in Lake Erie. *J. Great Lakes Res.*, 13:628–648.

________, and Mark, D. 1988. Effect of storms and gage location on tributary load estimate. *J. Environmental Engrg.* 114:352–367.

________, and Lindsay, D., Mattox, W., and Herdendorf, C. 1983. *A review of estuary hydraulics and transport as applied to rivers tributary to Lake Erie.* USGS-OWRT Proj. Comp. Rept. Water Resource Center, The Ohio State University, Columbus, Ohio.

Boyce, F. M. 1974. Some aspects of Great Lakes physics of importance to biological and chemical processes. *J. Fish Res. Board Can.* 31:689–730.

________, and Chiocchio, F. 1987. Water movements at a mid-central basin site: time and space scales, relation to wind, and internal pressure gradients. *J. Great Lakes Res.* 13:530–541.

Brandt, R., and Herdendorf, C. 1972. Delineation of Great Lakes estuaries. In *Proc. 15th Conference Great Lakes Research*, pp. 710–718. Internat. Assoc. Great Lakes Res.

Burns, N. M., and Ross, C. 1972. *Project Hypo—discussion of findings.* Canada Centre for Inland Waters, Burlington, Ontario, Canada. US EPA Technical Report. TS-05-071-208-24.

Cameron,W., and Pritchard, D. 1963. Estuaries. In *The Sea*, ed. M. Hill, pp. 306-324. New York: Wiley Interscience.

Campbell, J., Clites, A. and Greene, G. 1987. *Measurements of ice motion in Lake Erie using satellite-tracked drifter buoys.* NOAA-Great Lakes Env. Res. Laboratory Data Rept. No. ERL GLERL-30, Ann Arbor, MI.

Csanady, G., 1982. *Circulation in the Coastal Ocean.* Dordrecht, Holland: D. Reidel Pub. Co.

________, and Scott, J. 1974. Baroclinic coastal jets in Lake Ontario during IFYGL. *J. Physical Ocean.* 4:524–541.

Derecki, J., and Quinn, F. 1990. Comparison of measured and simulated flows during the 15 December 1987 Detroit River flow reversal. *J. Great Lakes Res.,* 16:426–435.

Dingman, J., and Bedford, K. 1984. The Lake Erie response to the January 26, 1978 cyclone. *J. Geophysical Res.* 89(C4):6427-6445.

Dyer,K., 1973. *Estuaries: A Physical Introduction.* London: J. Wiley and Sons.

________. 1986. *Coastal and Estuarine Sediment Dynamics.* Chichester UK: J. Wiley and Sons.

________. 1988. The role of hysteresis in the estuarine turbidity maximum. In Ocean Sciences 1988, EOS 68(50):1783.

Fraleigh, P., Burnham, J., and Gronau, C. *et al.*, 1975. *Maumee Bay Environmental Quality Study.* Final report, Toledo Lucas County Port Authority, Toledo, Ohio.

Gedney, R., and Lick, W. 1972. Wind-driven currents in Lake Erie. *J. Geophysical Res.* 77:2714–2723.

Hamblin, P. 1979. Great Lakes storm surge of April 6, 1979. *J. Great Lakes Res.* 5:312–315.

________. 1987. Meteorological forcing and water level fluctuations on Lake Erie. *J. Great Lakes Res.* 13:436–453.

Hansen, D., and Rattray Jr., M. 1966. New dimensions in estuary classification. *Limnol. and Oceanog.*, 11:319-326.

Harris, L. 1957. The effect of a moving pressure disturbance on the water level in a lake. *Meteorological Monographs* 2(10):46–57.

Haq, A., and Lick, W. 1975. On the time-dependant flow in a lake. *J. Geophysical Res.* 80:431–437.

Herdendorf, C. 1987. *The Ecology of the Coastal Marshes of Western Lake Erie: A Community Profile.* Rept. 85(7.a) National Wetlands Research Center, Fish and Wildlife Service, US Dept. of the Interior, Washington, D.C.

________. 1990. Great Lakes estuaries. *Estuaries* 13:493-503.

________, and Zapotosky, J. 1977. *Effects of Tributary Loading to Western Lake Erie During Spring Runoff Events.* Center for Lake Erie Area Research, Report 65, The Ohio State University, Columbus, OH.

Hunt, I. J. Jr., 1958. Winds, wind set-ups and seiches on Lake Erie. Second National Conference on Applied Meteorology in Engineering.

Irish, S., and Platzman, G. 1962. An investigation of the meteorological conditions associated with extreme wind tides on Lake Erie. *Mon. Weather Review* 90:39–47.

Kemp, A., Thomas, R., Dell, C., and Jaquet, J. 1976. Cultural impact on geochemistry of sediments in Lake Erie. *J. Fish Res. Board Can.* 33:440–462.

Lam, D., and Simons, T. J. 1976. Numerical computations of advective and diffusive transports of chloride in Lake Erie during 1970. *J. Fish Res. Board Can.* 33:537–549.

Lee, D., and Bedford, K. 1987a. Resuspension and sorting effects on tributary sediments loads to Lake Erie. In *Topics in Lake and Reservoir Hydraulics*, ed. W. Graf and U. Lemmin, pp. 56-61. Proc. XXII IAHR Conf. Ecole Polytechnique Federale, Lausanne Switzerland.

________, and Bedford, K. 1987b. *Modelling Storm and Entrainment Effects on Tributary Sediment Loads.* Rept. Coastal Engrg. Laboratory, Dept. Civil Engrg., Ohio State U., Columbus, OH 43210.

Libicki, C., and Bedford, K. 1990. Sudden, extreme Lake Erie storm surges and the interaction of wind stress, resonance, and geometry. *J. Great Lakes Res.* 16:380–395.

Lindsay, D. 1981. A Review of Estuary Hydraulics and Related Coastal Processes; Application to the Rivers Entering Lake Erie. Thesis, Master of Science in Civil Engineering, Ohio State U., Columbus, Ohio.

Lyons, J., Bedford, K., Yen, J. C. C., Lee, D., and Mark, D. 1988. Determination of suspended sediment concentrations from multiple day Landsat and AVHRR data. *Journal of Remote Sensing of the Environment* 24:364-373.

McDowell, D., and O'Conner, B. 1977. *Hydraulic Behavior of Estuaries.* New York: John Wiley and Sons.

Mortimer, C. 1963. Frontiers in physical limnology with particular reference to long waves in rotating basins. In Proc. 6th Conference Great Lakes Res., pp. 9–42. Internat. Assoc. Great Lakes Res.

________. 1987. Fifty years of physical investigation and related limnological studies on Lake Erie, 1928–1977. *J. Great Lakes Res.* 13:407–435.

Officer, C. 1976. *Physical Oceanography of Estuaries and Associated Coastal Waters.* New York: John Wiley and Sons.

Pinsak, A., and Meyer, T. 1976. *Environmental Baseline for Maumee Bay.* Great Lakes Basin Commission Rept. NOAA Great Lakes Environmental Research Lab., Ann Arbor, MI.

Platzman, G. 1963. The dynamical prediction of wind tides on Lake Erie. *Meterological Monographs* 4(26): 1-41.

________, and Rao, D. B. 1964a. Spectra of Lake Erie water levels. *J. Geophysical Res.* 69:2525-2533.

________, and Rao, D. B. 1964b. The free oscillations of Lake Erie. In *Studies on Oceanography*, ed. K. Yoshida, pp. 359-382. University of Washington Press, Seattle, Washington.

Pore, N., Perrotti, H., and Richardson, W. 1975. *Climatology of Lake Erie Storm Surges at Buffalo and Toledo*. NOAA Tech. Memo NWS TDL-54, National Weather Service, Silver Spring, Maryland.

Prater, M., and Bedford, K. 1982. Element type comparison in basin oscillation analysis. In *Proc. 1982 Army Numerical Analysis and Computer Conference*, pp. 415-428. ARO Report 82-83, Waterways Exp. Sta., Vicksburg, VA.

Pritchard, D. 1955. Estuarine circulation patterns. *Proc. American Society of Civil Engineers*. 81(9), No. 717.

Rao, D. B. 1967. Response of a lake to a time-dependant wind stress. *J. Geophysical Res.* 72:1691-1708.

Richards, P., and Baker, D. 1982. *Assimilation and flux of sediments and pollutants in the Sandusky River estuary, Sandusky River and the adjacent nearshore zone of Lake Erie*. Rept. NOAA-OMPA#NA80RA D00038. Heidelberg College, Tiffin, Ohio.

Saylor, J., and Miller, G. 1987. Studies of large scale currents in Lake Erie, 1979-1980. *J. Great Lakes Res.* 13:487-514.

________, Huang, J. C., and Reid, R. O. 1980. Vortex modes in southern Lake Michigan. *J. Physical Oceanography* 10:1814-1823.

Schertzer, W. 1987. Heat balance and heat storage estimates for Lake Erie. *J. Great Lakes Res.* 13: 454-467.

________, Saylor, J., Boyce, F., Robertson, D., and Rosa, F. 1987. Seasonal thermal cycle of Lake Erie. *J. Great Lakes Res.* 13:468-486.

Schroeder, M., and Collier, C. 1966. *Water quality variations in the Cuyahoga River at Cleveland, Ohio*. U.S. Geological Survey paper No. 550-C: 251-C 255.

Schwab, D. 1978. Simulation and forecasting of Lake Erie storm surges. *Mon. Weather Review*. 106: 1476-1487.

Sorenson, R. 1978. *Basic Coastal Engineering*. New York: J. Wiley and Sons.

Verber, J. 1955. Rotational water movements in western Lake Erie. *Proc. Int. Assoc. of Theoretical and Applied Limnology* 12:97-104.

Wright, L., and Coleman, J. 1974. Mississippi River mouth processes, effluent dynamics and morphlogic development. *J. Geology* 82:751-778.

Submitted: 18 July 1991
Accepted: 29 July 1992