Model and Observed Circulation Throughout the Annual Temperature Cycle of Lake Michigan

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(Manuscript received 28 December 1977, in final form 11 December 1978)

ABSTRACT

Monthly average currents and temperatures predicted by a three-dimensional, numerical model of Lake Michigan are compared with observations made in that lake during June–October 1976. The observed data are from 17 current meters with integral temperature recorders that were concentrated on a transverse section of the southern basin of the lake. A brief interpretation of the overall aspects of these data is given and the evolution of a deep temperature anomaly in the west-central basin is discussed. Model results are evaluated in terms of their comparability with the dominant features of the observed data. Lakewide-average temperatures in the model are reasonable and the signs of the computed and observed currents show some agreement. However, the model exaggerates upwelling along the upwind (western) shore, leading to temperature predictions that worsen progressively throughout the stratified season. The present study and other recent work suggest the need for improved mixed-layer physics in lake models.

1. Introduction

Previous numerical modeling of the Great Lakes has been focused primarily on simulating the dynamics of Lake Ontario, (Simons, 1974, 1975, 1976; Bennett, 1977). These modeling efforts have utilized the wealth of high-quality data from the International Field Year in the Great Lakes (IFYGL) for model input and to judge model results. Relatively little attention has been given to the modeling of Lake Michigan due to the paucity of recent long-term records of current and temperature in that lake. In addition, no one has yet reported on an attempt to simulate currents and temperatures over the annual cycle of one of the Great Lakes.

In the present study the numerical model developed by Bennett (1977) is applied to Lake Michigan in an effort to simulate currents and temperatures throughout the lake's cycle of stratification and destratification. The simulations are based on very limited wind and surface heat flux data, in contrast to those provided by IFYGL. It is of interest, however, to see how well the model can perform with such limited data. The observed currents and temperatures required to judge model performance are provided by new data collected during May–November 1976 by the Great Lakes Environmental Research Laboratory (GLERL), National Oceanic and Atmospheric Administration (NOAA). These new data are concentrated mainly on a latitudinal section across the southern basin of the lake, and represent the only recent lake measurements that could be utilized in the present study. The salient features of this comparatively large field effort are described in Section 3, and a brief interpretation of the data is offered in Section 4, although the full potential of these data is not explored here. All of the model simulations in this study encompass the NOAA data period. Model results are discussed in Section 5 in terms of their comparability with the dominant features of the observations on a monthly basis.

2. Numerical model and ancillary data

The numerical details of the circulation model, as applied to Lake Ontario, are given by Bennett (1977). He describes experiments with two somewhat different models. The version of the model applied here uses a non-uniform grid and computes vertical eddy exchange coefficients from a variation of the Munk-Anderson formulas. The only change to the numerical framework for the present study is that advection of momentum is included.

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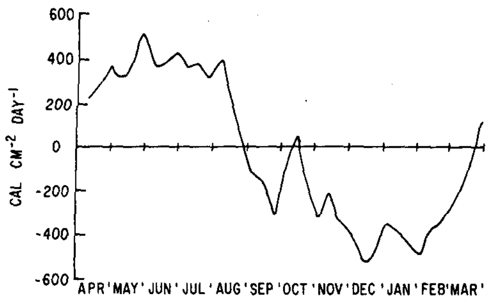

FIG. 1. Net surface heat flux (after Church, 1947).

The shape of Lake Michigan is approximated here by a rectangle 100 km × 500 km, with the longer dimension oriented north-south. These dimensions present about 95% of the actual surface area of the lake. The depth distribution is approximated by an analytic function that is latitudinally parabolic and is the sum of two Gaussian curves longitudinally. This function locates the northern and southern basins of the lake in a realistic manner, and gives the maximum depths of each basin, the average depth and the total lake volume to within 4% or less of their true values.

The finite-difference grid is 24 × 16 × 8 in the north-south, east-west and vertical directions, respectively, and is variable in all three dimensions. The minimum east-west resolution is 3.5 km at the boundaries, increasing by a factor of 1.25 for each successive grid box toward the center of the basin. The minimum north-south resolution is 6.2 km, increasing symmetrically by a factor of 1.25 toward the center. Some grid boxes are eliminated at each corner to imitate a curved boundary. Interface depths are at 6, 15, 27, 43, 67, 101 and 149 m, with the exception that the thickness of the bottom box varies according to the depth distribution. All simulations begin on 10 April 1976 when the lake temperature is taken as 2.0°C isothermal, and the initial velocity field is zero. A 1 h time step is used for most simulations.

Ancillary data required by the model are surface heat flux and wind stress. Church (1947) estimated the total heat lost or gained by the mid-lake section from Milwaukee, Wisconsin, to Muskegon, Michigan, during 1942–44. His estimates were based on a large number of bathythermograph surveys made from passenger car ferries. The estimated surface heat flux (change in total heat content), represented here by the mean of the two years of data, is shown in Fig. 1. The maximum heat input (positive) to the lake is about 500 cal cm^{-2} day^{-1} (231 W m^{-2}) and occurs in early June. The maximum heat loss (negative) is of similar magnitude and occurs in early December. Advective effects are obviously omitted in Church's results. However, these results provide a first approximation to use in the model, and small adjustments give maximum water temperatures that are in rough accord with observations (see Section 5).

The wind data for this study are the daily winds reported at Milwaukee (no overwater wind data were available). Speed and direction corrections can be used to yield a more realistic overwater wind. Resio and Vincent (1977) gave a cogent summary of the factors involved in estimating winds over the Great Lakes. Following this summary, a speed-dependent multiplier is applied to each overland speed, and a clockwise veering is added to each overland direction. The veering angle is a function of stability as well as speed. Monthly air-sea stabilities are taken as follows: unstable—latter half of November, December, January, February and March; stable—June, July, August and September; neutral—April, May, October and first half of November. Table 1 summarizes all of the correction factors. The speed correction makes a large difference in the estimated wind, while the direction corrections are comparatively small. Omission of the clockwise veering would be a systematic error, however. The sensitivity of the model to these corrections is discussed in Section 5.

The bulk transfer coefficient C_D relating wind stress to the square of the surface winds is based on the work of Hicks (1975) and is taken as

$$C_D = 10^{-3}\,[1 + 0.07(u_{10} - 5)],$$

where u_{10} is the corrected wind speed (m s^{-1}) at 10 m height. Fig. 2 shows the daily stress with all correction factors employed. Easterly stresses are predominant, as expected. Monthly averages of these data show that southerly stresses dominate over northerly stresses, except during June. Finally, stresses are typically 1.5–2 times greater during the winter months than the summer months.

3. Data collection

During May through November, 1976, an extensive survey of water currents and temperature structure was performed in the southern basin of Lake Michigan. GLERL planned and coordinated the field study and was the primary data collector. The objectives of the experiment were twofold. Transects

TABLE 1. Wind corrections.

Speed (m s^{-1})	Correction factor	Clockwise veering (deg)		
		Unstable	Neutral	Stable
2.6	1.9	19	15	5
5.1	1.6	18	14	5
7.7	1.5	17	13	4
10.3	1.3	15	11	3
12.9	1.2	13	9	2
15.4	1.2	11	7	1

perpendicular to the eastern coast of the southern basin were established to monitor flow characteristics within the coastal boundary layer and to detect the propagation of coastally trapped waves (Fig. 3). The University of Wisconsin at Madison participated in this part of the study and installed a dense array of current meters (with integral temperature recorders) in the vicinity of Holland, Michigan. A lake cross section extending from Holland, Michigan, to Racine, Wisconsin, also was intensely instrumented, primarily for the purpose of studying the characteristics of near-inertial period, internal Poincare waves. The University of Wisconsin at Milwaukee participated in this part of the experiment and installed a series of five current and temperature recording stations (not shown in figure) spaced along a longitudinal axis of the southern basin to determine the north-south structure of these wave forms. Argonne National Laboratory (ANL) added three current meters placed at 1 m above the lake bottom on the transverse cross section.

Data used in the present study are from 17 current meters (with integral temperature recorders) placed on the transverse cross section by GLERL. Though configured to study transverse internal waves, the dense array of current meters afforded an excellent opportunity to quantify flows into and out of Lake Michigan's southern basin. The currents were measured with AMF vector-averaging current meters, recording at 15 min intervals. The meters were suspended in a taut line below subsurface floats located just above the upper instrument on each mooring; current-meter depths are indicated by the data points in Figs. 4 and 5. Monthly averages of the north-south velocity component and of the water temperature are used in this study. Accuracies of the temperature measurements are within 0.1°C, while the currents are small residuals of usually quite larger speeds. In mid-lake regions the recordings were characterized by dominant oscillatory flows of near-inertial period associated with internal inertio-gravitational waves. Because of the presence of these oscillatory flows, we can report with some confidence mean flow speeds that otherwise would

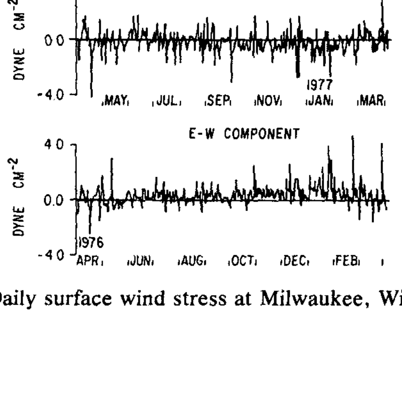

FIG. 2. Daily surface wind stress at Milwaukee, Wisconsin.

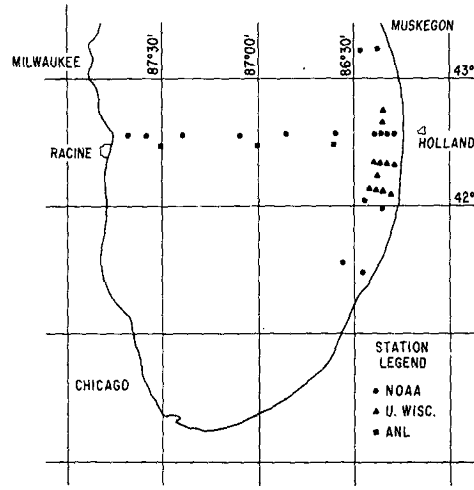

FIG. 3. Locations of current-temperature moorings, 1976.

be below the threshold velocities of the Savonious rotor current meters used.

4. Interpretation of data

A crude synopsis of the present observations can be given through monthly averages of the observed currents and temperatures measured across the southern basin of the lake. A month certainly does not represent an optimum averaging interval, because the degree of stratification of the lake can change markedly during a 30-day period. However, the aggregate effects of several interesting physical mechanisms can be summarized expediently in the present context. Fig. 4 shows the north (+) and south (−) components of the monthly-average observed (and computed) currents for June–October 1976. Fig. 5 shows the monthly-average observed (and computed) temperatures for the same period. The average northward (+) and eastward (+) wind stress components inferred from the Milwaukee data (cf. Section 2) are also given in each figure.

The overall evolution of observed currents and temperatures meets with expectations in several regards. Barotropic effects (downwind flow near the lateral boundaries with a weaker return flow in the interior) are dominant in June, when stratification is weak. July represents a transitionary month between the spring and summer regimes because of the large change in vertical thermal structure. Hence, the monthly-average currents are somewhat difficult to interpret in terms of a single dominant mechanism. The basin is well-stratified by August, when baroclinic effects are dominant and the average wind is at its minimum. Boundary currents strengthen in September in response to the average effect of stronger wind impulses, and average surface temperature begins to decrease as a result of net surface heat loss. The fall overturn begins in October and a

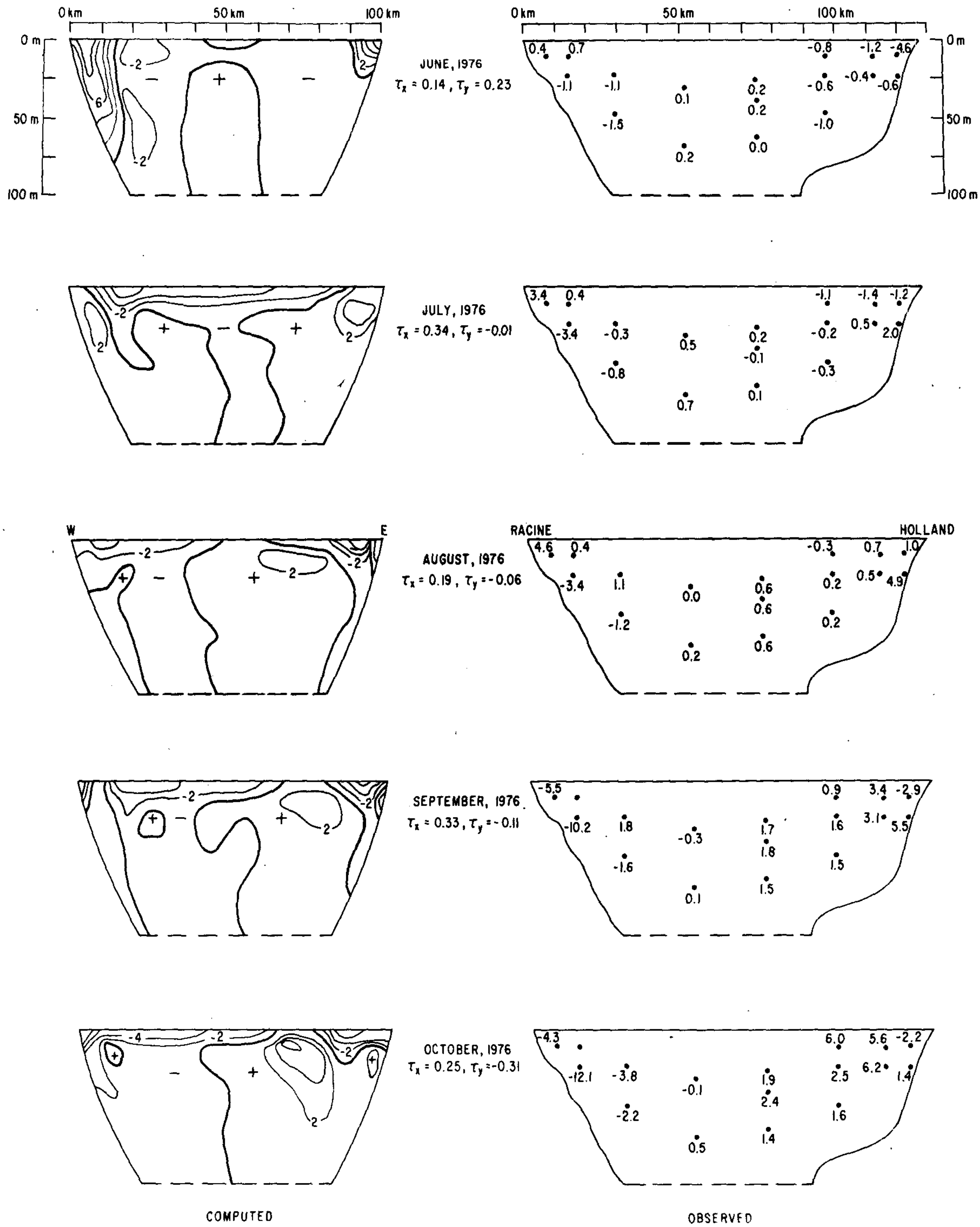

FIG. 4. North (+) and south (−) components of monthly-average computed and observed currents. All speeds in cm s^{-1}, stresses in dyn cm^{-2}.

rather unified cyclonic circulation becomes apparent.

Two noteworthy features of the observations deserve further comment. First, observed flow in the coastal boundary layers during August can be explained quite well in terms of the conceptual model given by Csanady (1968). That model describes the flow in a two-layer circular basin resulting from wind-induced deformation of the thermocline (i.e., the impermeable surface separating the two model

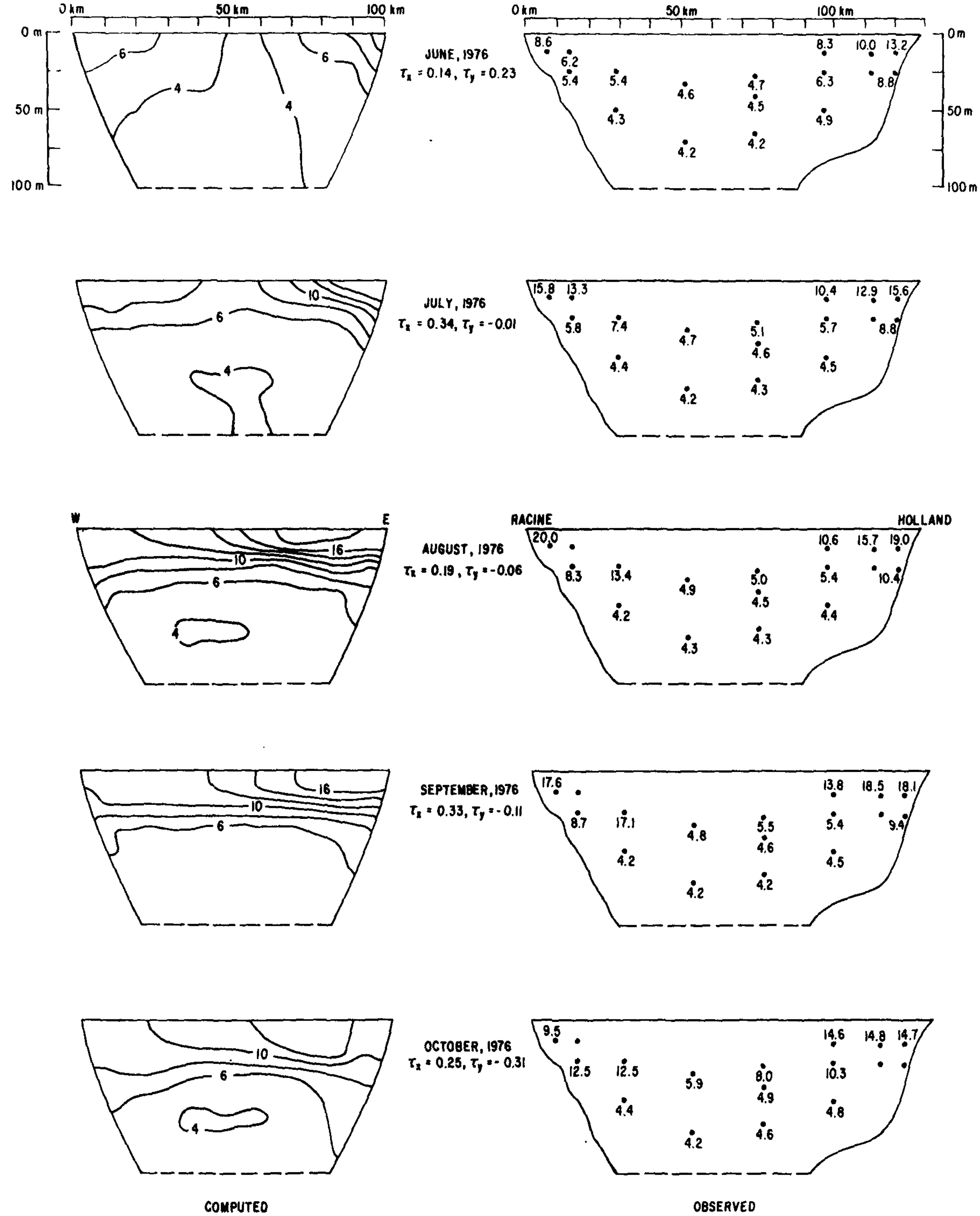

FIG. 5. Monthly-average computed and observed temperatures (°C).

layers). The calculated response of the thermocline along a diameter parallel to a steady and uniform wind stress is particularly à propos. In terms of the conceptual model, the August thermocline at the eastern shore is depressed so that the instrument at the 25 m depth is in the upper layer of the two-layer analogue; the observed flow is northward (to the left of the wind), in agreement with the conceptual model. Conversely, the thermocline at the western shore is raised so that the instrument at the 25 m depth is in the lower layer; the observed flow is southward (to the right of the wind), again in

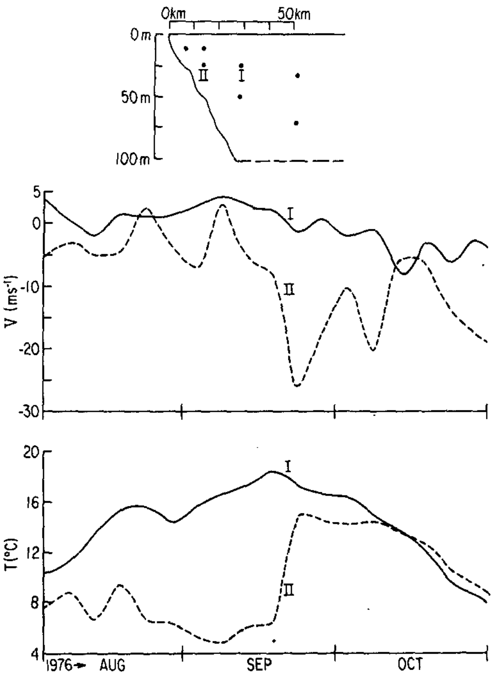

FIG. 6. Five-day average temperatures and northward flow at 25 m depth from stations I and II (see inset).

agreement with Csanady's model. Two factors appear to contribute to the applicability of the conceptual model in the present situation. The southern basin has a relatively circular configuration and is somewhat separated from the rest of the lake by a broad sill. The average wind is directed primarily along the transverse axis of the basin, i.e., the diameter in terms of the two-layer analogue.

The second outstanding feature of the observations is the deep, warm lens that develops in the west-central part of the basin during late August and through mid-September. A remarkably high average temperature of 17.1°C is found during September at the 25 m station located some 30 km from the western shore. This deep temperature anomaly is accompanied by an equally remarkable southward flow of 10.1 cm s^{-1} at the 25 m station to the west of the warm lens where the average temperature is only 8.7°C. The temperature difference between these two stations vanishes in October, although a strong southward flow of 12.5 cm s^{-1} persists at the shoreward station. A more detailed history of these rather surprising current and temperature measurements is given in Fig. 6, which shows 5-day averages of the currents and temperatures at the pertinent stations for August–October. Note that the temperature difference between station I (30 km from the western shore, 25 m depth) and station II (15 km from shore, 25 m depth) grows rather steadily from August through mid-September to about 12°C. Then, between the 16th and 25th of September, this temperature difference decreases to about 2°C, and a 5-day average flow of −26.1 cm s^{-1} is observed at station II. The temperature difference continues to decrease, as does the average temperature, while the average current remains strong but varies widely. A clear explanation for the generation and maintenance of this temperature anomaly cannot be given based on the data that are available. The destruction of this feature, however, is likely to be related to the weakening of the coastal thermal structure in late September. Examination of the wind history shows that two storms with relatively strong southward components occurred during the 10-day period cited above, followed by a period of light winds. The more rapid heat loss from the shore zone than from the central basin, coupled with these wind impulses, may provide the mechanism for the observed leveling of isotherms and the contemporary conversion from potential to kinetic energy. These data seem to capture the onset of the lakewide cyclonic circulation, a ubiquitous feature of motion in the Great Lakes that apparently persists into the unstratified season.

5. Model results

Monthly-average currents and temperatures predicted by the numerical model are given in Figs. 4 and 5, respectively. Note that the model section width is 100 km, by choice, while the actual section width is about 130 km. Model predictions are included in the same figures as the observations to facilitate discussion, rather than for direct comparison. Indeed, previous work (Allender, 1977) suggests that there are inherently serious restrictions on pointwise comparisons for currents. The average temperature varies more smoothly in space than do the currents, and consequently model-data comparisons can be made with somewhat greater scrutiny.

It is rather surprising that the model performs as well as it does, considering the rather poor model-input data that are used. The model basin appears to stratify and destratify at realistic times (late July and mid-December, respectively). Lakewide average temperatures are reasonable, and the signs of the computed and observed currents are generally in agreement away from the lateral boundaries. It should be noted that model current speeds were usually much too low unless the speed corrections (Table 1) were applied to the overland winds. Model results were not very sensitive to wind direction corrections, however. The overwhelming failure of the model is that it greatly exaggerates upwelling at the upwind (western) shore. A similar failure was noted by Simons (1976) in his modeling of long-term heat transports in Lake Ontario during IFYGL. The spurious cooling predicted by the model becomes progressively worse from June through August. The excessive transport of warm water toward the eastern shore in the model apparently precludes the formation of the observed, deep, temperature anomaly in September. The strong currents associated with this anomaly are, of course, lacking in the model results. A general cyclonic pattern develops in the October results in very rough accord with the observations. Predicted currents near the lateral boundaries are generally poor, however, due to the wrongly predicted shape of the thermocline.

Several attempts to improve model results were made. It was found that moderate changes in vertical diffusion did not improve the simulated temperature field. Most numerical experiments used 32 cm s^{-2} as the constant factor in Bennett's variation of the Munk-Anderson formulas for vertical exchange of momentum and heat. This factor was varied from 24 to 40 cm s^{-2} in different experiments with no noticeable improvement in results. (Drastic reduction of this factor to, say, 10 cm s^{-2} or less caused excessively high surface temperatures.) Small changes ($\pm 15\%$) in net surface heat flux (see Fig. 1) did not lead to model temperatures that were in better accord with the observations. Finally, the effects of a fetch-dependent wind were considered. Wind stress was increased from zero to full strength over a distance of 30 km from the upwind shore, according to the empirical fetch dependence for the Great Lakes suggested by Resio and Vincent (1977). This spatial variation in stress, although probably existent in nature, did not change model predictions appreciably.

Two conjectures for improving model results under stratified conditions were made by Simons (1976). He suggested 1) increased vertical resolution to allow return flow to be concentrated near the thermocline, thus minimizing advective heat transport; and 2) quasi-impermeable model interfaces to allow a more reversible model response. The present study (which was not designed to test these conjectures), and other recent work, suggest two additional areas for model improvement. First, Jacobs (1978) tested various forms for vertical eddy exchange coefficients and found several forms that were superior to the Munk-Anderson formulas. In fact, the latter were found to underestimate the sharpness of the temperature profile under steady advective effects, which is roughly similar to the present findings. Second, the present model omits the potential important effects of a shallow mixed layer during the development of the thermocline. Heat advection might be confined to a shallow slab, well above the thermocline, except during storm conditions. Lack of such confinement in the present model could account in part for the overprediction of upwelling. Ellsburg and Garwood (1978) have commented on the observation and effect of shallow mixed layers regarding the generation of sea surface temperature anomalies. Such layers also were observed during IFYGL and appeared in daily lakewide averages of vertical temperature structure (see e.g., Tucker and Green, 1977). The necessary data to resolve shallow mixed layers were not available in the present study. Further observational studies are needed to document the effects of these layers on motion in the Great Lakes.

Acknowledgments. The numerical experiments were supported by the U.S. Energy Research and Development Administration. The authors are grateful to J. R. Bennett for his helpful suggestions during this study and to G. S. Miller for his assistance in data processing. The U.S. Naval Ocean Research and Development Activity is thanked for providing technical aid in the completion of this work.

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