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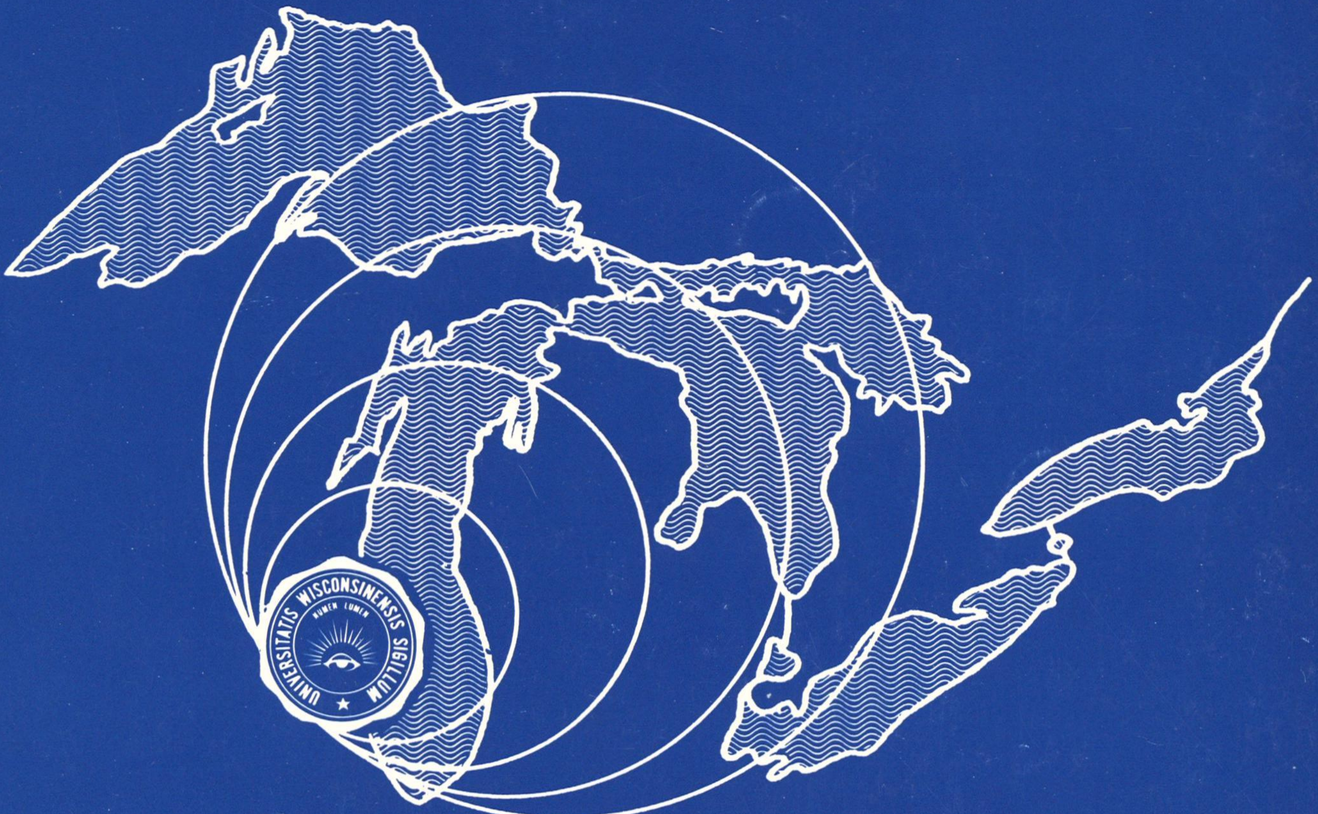

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LONG INTERNAL WAVES IN LAKES: REVIEW OF A CENTURY OF RESEARCH

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Long internal waves in lakes: review of a century of research.*

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Abstract

The century began with Murray's 1888 demonstration of isotherm-tilting forced by wind in stratified lakes in Scotland, where later studies by Wedderburn (1903-1913) identified the internal seiche as the free response to that forcing. Wedderburn's findings were confirmed and extended in Windermere, 1947-51, by deployment of thermistor chains and by Heaps' modelling of the forced and free responses. Transferred to Loch Ness in 1952, the thermistor chains revealed internal seiches of large amplitude with marked nonlinear features, subsequently explored and interpreted by Thorpe and here compared with internal surges in other lakes.

Modification of the internal seiche by the earth's rotation was demonstrated in Loch Ness and Lac Léman, in the latter case by means of 1950 surface-level "signatures" of postulated internal Kelvin waves. More recent findings in Léman and in lakes of similar size are reviewed. Temperature records from water intakes around a much larger basin (Lake Michigan) revealed (i) occasional cyclonic progression of nonlinear, shore-trapped interfacial waves traveling at Kelvin wave speeds. Those waves were set in motion after strong, sustained wind had forced extensive downwelling along one side and upwelling along the other side of the basin. Also revealed were (ii) persistent thermocline oscillations of inertial frequency or slightly blue-shifted therefrom. To interpret observations (i) and (ii) the writer invoked a well-known wide-channel model which incorporated shore-trapped Kelvin waves, offshore inertial (Sverdrup) waves, and channel-bound Poincaré waves. That model was tested by extensive current and temperature measurements in Lake Michigan (1963/4) and Lake Ontario (IFYGL 1972). The achievements and limitations of the model are discussed in the light of those findings. The responses to changes in wind stress were: locally-generated inertial motion offshore (the most common and immediate response of stratified Great Lakes to wind impulses); Ekman drift and consequent nearshore up- and downwelling; geostrophic readjustment near shore, which generates shore-modified inertial waves and associated offshore-migrating fronts, also shore-trapped Kelvin waves and surges; and relatively rare transverse internal seiches.

Each of those phenomena, here described for Lakes Michigan and Ontario and anticipated in other large basins, are to be viewed as components of the combined response of the basin to wind stress changes. At a particular time or place, one or the other component may dominate; and some may be seen more often than others; but they remain combined and interacting.

The review closes with an examination of some unanswered questions and a synopsis.

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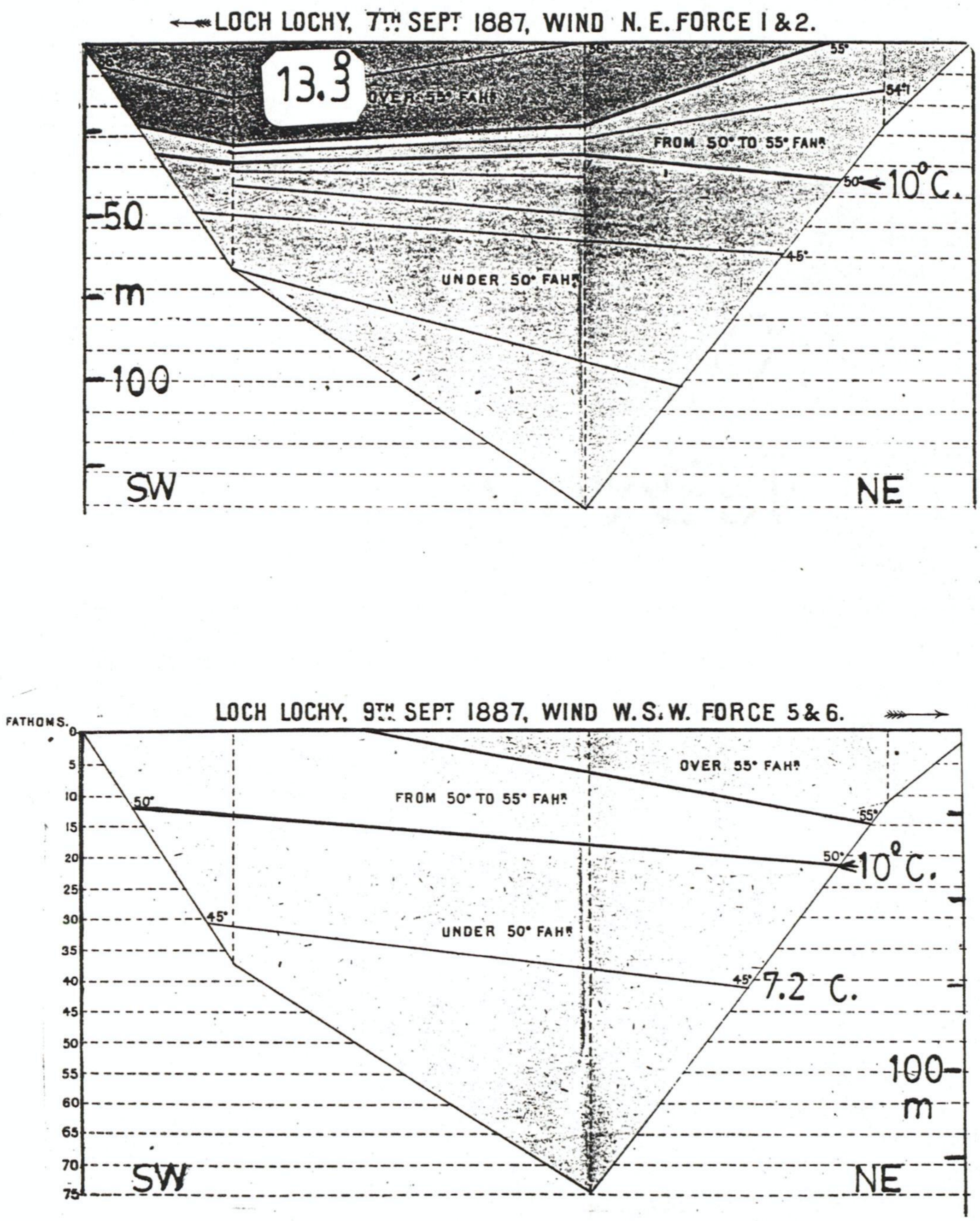

Fig. 1. Wind-induced isotherm displacements in Loch Lochy, Scotland, 1887 (retouched facsimile of Fig. 4 in Murray 1888).

Early lake thermometry and oceanographic connections

The word "century" in the title is appropriate because 1888 saw the publication of John Murray's first description of wind-forced tilting of isothermal surfaces in the coastal-marine and freshwater basins of Scotland (Fig. 1).* The types of oscillatory responses of water basins, in general, to such wind-forcing are the subject of this review. Although, in 1888, the occurrence of internal waves in lakes was unknown, the systematic exploration of lake temperature distributions had begun (Geistbeck 1885**, Buchanan 1886). With an improved tool -- the reversing thermometer -- such studies were continued in sub-alpine lakes by Richter (1891, 1897), Grissinger (1892) and, above all, Forel (1895) whose classic study of the surface oscillations (seiches) of Léman justified his definition of limnology as "l'oceanographie des lacs."

The reversing thermometer, however, did not lend itself to the simultaneous collection of long and detailed series of temperature records at various stations and depths. And, although Thoulet had drawn attention in 1894 to depth-oscillations of near-thermocline isotherms as "une sorte de seiche interieure", and Richter in 1897 was the first to invoke the wind as the generator of "leichte Schwankungen" (gentle fluctuations) in isotherm level from hour to hour at a single station, the extent and nature of internal

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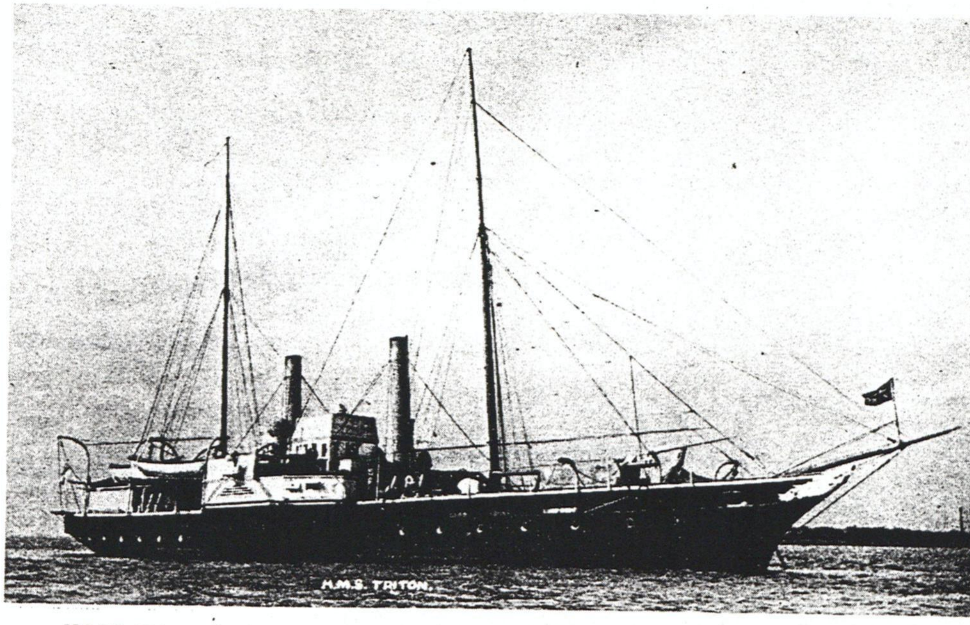
*(See opposite page). Commenting on pending fisheries legislation, Murray (1888) hoped that "a thorough knowledge of the physical and biological conditions of our coasts and lochs might, at least, tend to prevent pernicious and foolish legislation."

**In his review of the previous century's work with less satisfactory instruments, Geistbeck (1885) cited temperature soundings at 6-weekly intervals in 1847-49 on Lake Thun by the Bern physicists deFischer and Brunner (1849). They described, for the first time, the seasonal cycle of thermocline formation and destruction. They suspended a string of heavily-insulated thermometers from a float in 180 m of water. They touchingly acknowledge (translation). "the hospitality and obliging cooperation given to us by the noble inhabitants of the castle which softened the contrarities inevitably attached to our enterprise." Those hospitable interludes ashore, needed to allow the thermometers to equilibrate, must have lightened the routine!



H.M.S. *Triton*, about 1911. Reproduced by permission of the Trustees, National Maritime Museum, Greenwich.

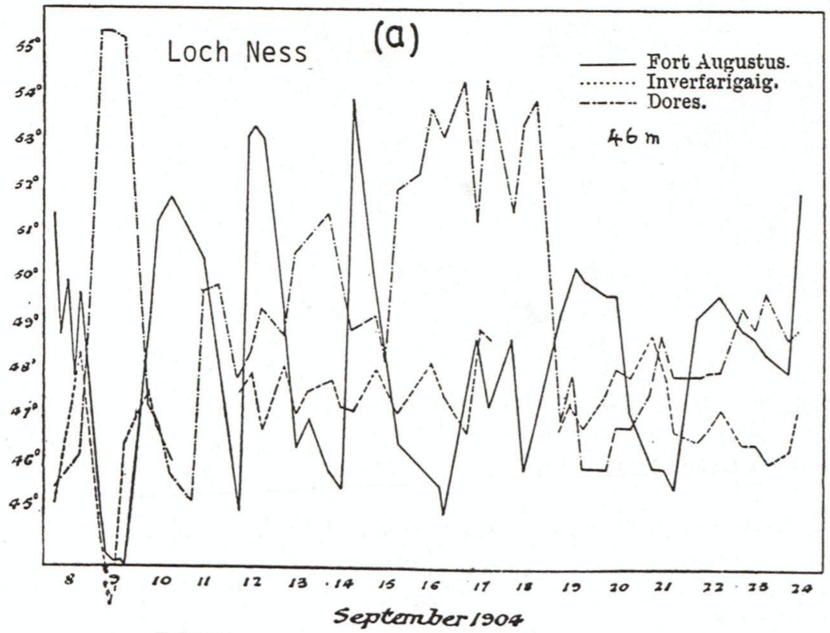
Fig. 2. H.M.S. "Triton" (from Deacon, 1977); Professor George Chrystal, 1883 etching by William Hole (source, Dept. Math., Univ. Edinburgh).

seiching was first revealed in Scottish lochs. There, as noted below, the debt to oceanography was a direct one.

In the decade which followed the pioneering world-ocean expeditions of H.M.S. "Challenger", the British Navy continued to place vessels at the disposal of scientists; but that support came to an end when H.M.S. "Triton" returned from the last of the post-Challenger cruises with the Scottish scientists J. Y. Buchanan, John Murray, and George Chrystal aboard (Deacon, 1977). The first two had been members, as physicist and naturalist respectively, of the Challenger Expedition; and the third was Professor of Mathematics at the University of Edinburgh (Fig. 2 and Black and Knott, 1911). Their task was to determine the pressure correction to be applied to the less-than-satisfactory maximum/minimum thermometers used on "Challenger." But the "Triton" cruise was to be the end, for the time being, of Navy support and government funding for oceanography. John (later Sir John) Murray used his own funds to set up and run the Scottish Loch Survey (of which the Scottish Lake Survey was a later offshoot) to study the coastal fjords and fresh-water basins of that country (Murray and Pullar 1910). There is no doubt, writes Margaret Deacon (1977), that Murray would

> "have preferred to continue the exploration of the deep oceans and felt that this was where the important results were to be won. This field, however, was to remain almost wholly closed to British scientists until the establishment of the Discovery Investigations 10 years after Murray's death."

But, that disappointment notwithstanding, the Lake Survey enabled Chrystal and his collaborators (notably E. M. Wedderburn) to contribute a brilliant chapter to lake physics, first with mathematical modelling and observational confirmation of surface seiches (Chrystal 1905, Chrystal and Wedderburn 1905), and then the adaptation of those models to internal seiches, coupled with and tested by remarkably thorough field work.



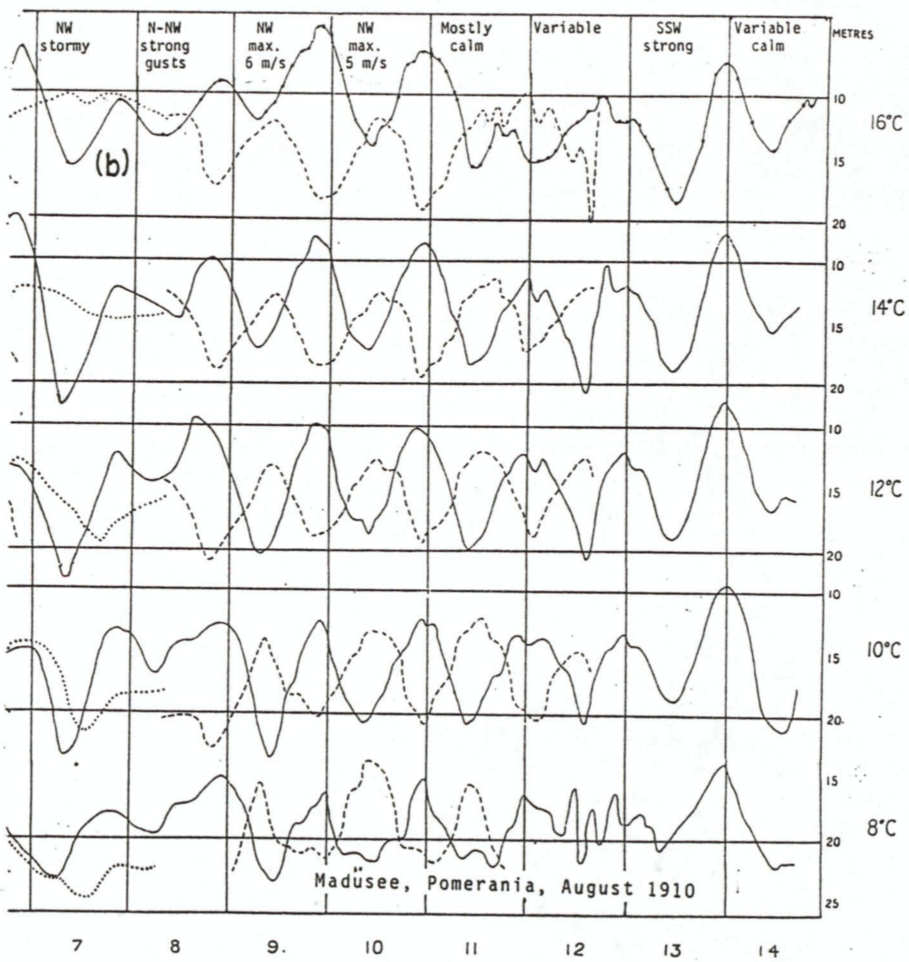


Fig. 3. Temperature fluctuations near the ends of: (a) Loch Ness, Scotland (Wedderburn, 1907); and (b) Madüsee, Pomerania (Wedderburn, 1911). The photograph of E. M. Wedderburn is from Anon (1947) taken some 30 years after he gave up lake research.

The Wedderburn Decade, 1905-1915

The Scottish Lake Survey's investigation of temperature fluctuations, led by Wedderburn in Loch Ness in 1903 and continued in 1904, used reversing thermometers and a continuously recording (but often defective) electrical thermometer, deployed from an anchored vessel and connected to shore near the SW end of the loch. "It soon became apparent [Wedderburn 1907] ... that movements of great magnitude were in progress ... [but] ... it was some time before the periodic nature of those changes was fully understood." E. R. Watson (1903, 1904), a member of the survey team, was the first to propose a 2-layered internal standing-wave model to explain the 1903 observations although (as Wedderburn 1907 pointed out) "many limnologists have expressed themselves skeptical of the soundness of his deductions". The periodic nature of the response was more evident in the 1904 observations (Fig. 3a), so also was an abrupt temperature rise as the boundary wave passed the recording station. Such rapid temperature fluctuations, to be discussed later with other examples, were regarded by Wedderburn (1907) as "quite natural ... where there are two layers of water of different temperatures in contact with one another, especially so in this case, as, owing to the presence of the temperature seiche [the first use of this term], there is relative motion between the layers".

Convinced of the widespread occurrence of temperature seiches as standing waves on the thermocline boundary, Wedderburn challenged one of the limnologist sceptics (Halbfass 1910) to a demonstration in a lake of the latter's choosing. The clear result (Fig. 3b) was a close fit in period and structure to a model which adapted Chrystal's equations and curve-fitting procedures to a 2-layered basin of irregular form (Wedderburn 1911). That result was also confirmed by a physical model (Wedderburn and Williams 1911). An expansion of

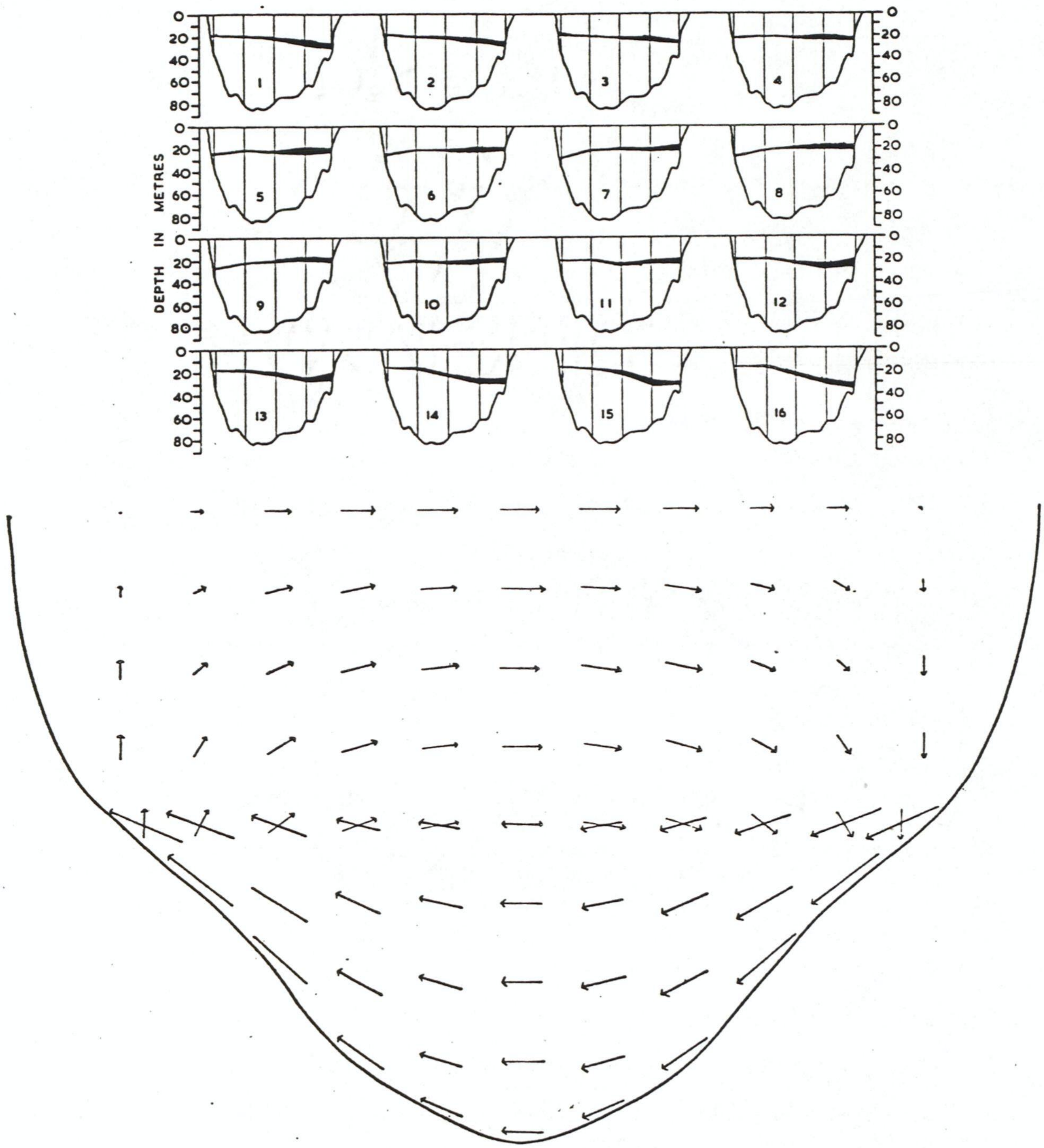


Fig. 4. Upper portion: hourly positions (0100 to 1600 hours, 9 August 1911) of the thermocline (black area bounded by the 9^{o} and $11^{o}C$ isotherms) on a longitudinal section of Loch Earn, redrawn from Wedderburn (1912). The vertical lines indicate station positions. Lower portion: Wedderburn's mathematical (1911) model of Loch Earn as a two-layered lake, the "interfacial normal curve" of which is parabolic. Motion of water particles is shown above and below the interface which oscillates about a central uninode.

the theory, to include the case in which density varies gradually in the vertical but which displays a layer of no horizontal motion, was applied (Wedderburn 1912) to a monumental series of observations from the anchored boats in Loch Earn. That basin was chosen because of its relatively simple shape and the prevalent along-lake directed wind stress, and because of the previous study of surface seiches there. There was satisfactorily close agreement between theory and the observed structures and periods of the principal modes (Fig. 4). This agreement justified the following conclusions:

> "It has been proved beyond all possible doubt that what we have called Temperature Seiches do exist, and that they are truly of the nature of a standing boundary wave.
>
> Examples of the effect of winds, both in starting and in damping oscillations already in progress, have been given, with an indication that even a wind of very moderate strength will start oscillations, and examples of oscillations forced by wind have also been obtained."

A final paper on the Loch Earn research (Wedderburn and Young 1915) will be discussed later in connection with the nonlinear dynamics of internal surges. That paper was Wedderburn's "swan song" as a lake physicist. After service in the 1st World War, he embarked on a brilliant legal career, as described by A.W.Y. (1959, presumably A. W. Young) in an obituary notice in the 1957-59 Yearbook of the Royal Society of Edinburgh and by (Anon 1947) in an article in the Scots Law Times.

<u>Early evidence of inertial oscillations (Scandinavian oceanography)</u>

During the decade of internal seiche investigations in Scotland, evidence for another type of oscillatory response to wind impulses began to emerge in the course of Scandinavian-based marine exploration. For example, Helland-Hansen and Nansen (1909) concluded that their observations of the wave-like fluctuations in currents and temperatures in the Norwegian Sea "indicate the probability that great, hitherto unknown, oscillatory movements may occur in the intermediate strata of the sea, and also at its surface. These movements may be

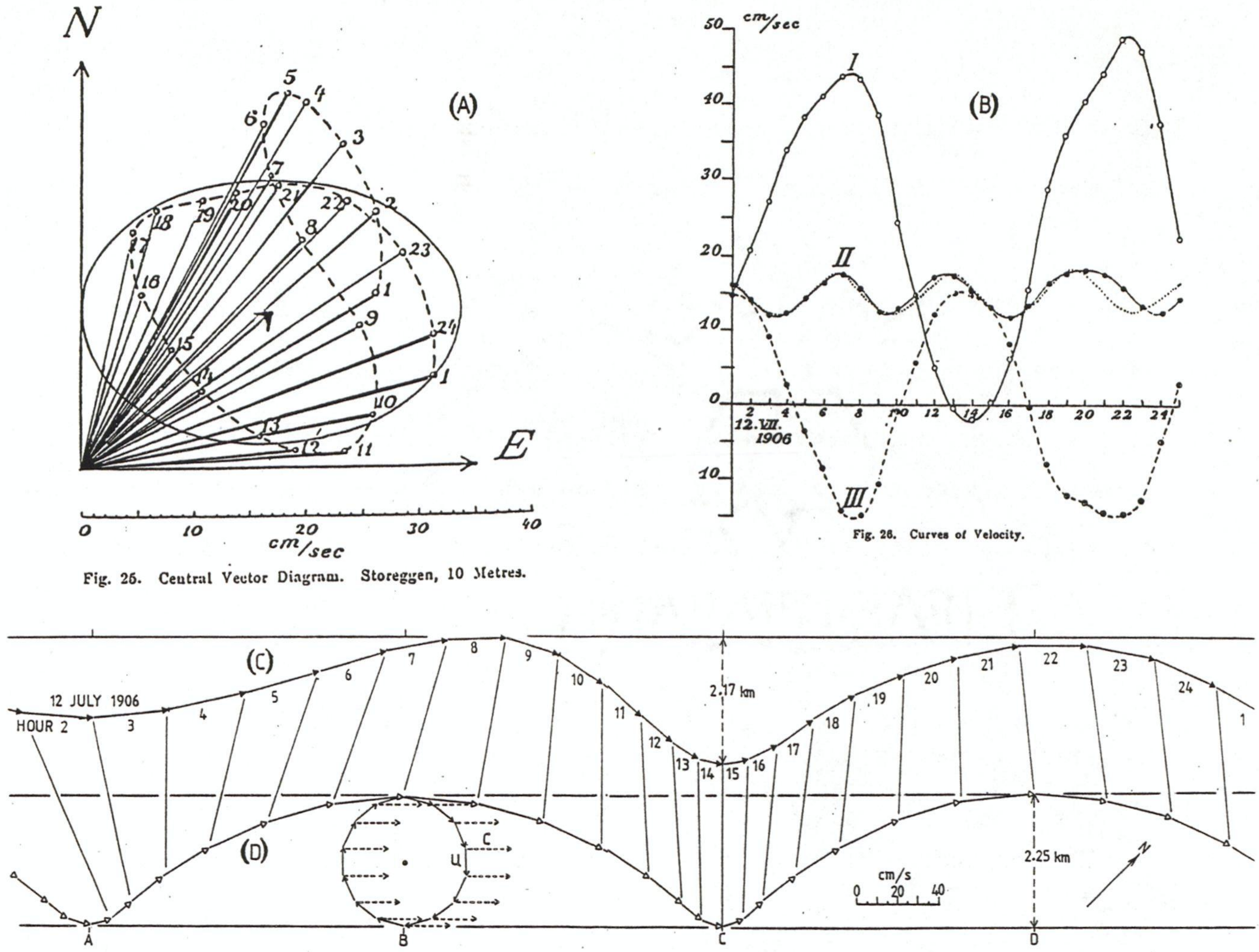


Fig. 5. A, hourly current vectors; and B, a model (details in text) of the current observed (Helland-Hansen and Nansen, 1909) at 10 m depth. C displays the same current episode in a progressive vector diagram (pvd), which may be compared with D, the pvd for an alternative model, combining a rotating inertial and a unidirectional current, as explained in the text.

waves of some kind on the boundary between water strata with different densities ... [or] ... may possibly also to some extent be due to some kind of pulsation in the velocity of the sea-currents."

To illustrate their conjecture, Helland-Hansen and Nansen presented an example (Fig. 5A) in which hourly-averaged vectors of current at 10 m depth were plotted from a common origin. The authors interpreted the observed current as a combination (Fig. 5B) of an anticlockwise rotating tidal current (defined by the ellipse in Fig. 5A) and a current steady in direction (50° N of E) but "pulsating" in speed. A present-day reader, with the benefit of hindsight, would suggest a more likely combination (in Fig. 5C) of: (i) a steady current of about 22 cm/s toward NE; and (ii) a current of about 14 cm/s rotating clockwise every 14 h. This fits the model of inertial motion, published four years earlier by V. W. Ekman (1905) to account for the initial transient response of horizontal surface currents to a change in wind stress. Ekman's model displayed clockwise rotation (in the N hemisphere) at a frequency near a "local inertial frequency", f, defined as 2 sin$\Omega\phi$, where ϕ denotes local latitude, and Ω the angular frequency of the earth's rotation. The corresponding "inertial period" in Fig. 5 was 14 h. The comparison is made, in the form of meandering, head-to-tail vector tracks, between the observed current track (C) and the theoretical inertial current track (D) below it, both lined up with the characteristic inflection point R (bottom line). During the forced phase of the response before R, agreement between the two "progressive vector" tracks is poor; but after R, during the free oscillatory response, the agreement is closer.

The inference from Fig. 5 is clear: the authors had observed for the first time a short episode, in which an inertial response was combined with a more steady current to produce the looping or meandering track, later shown to be a common "waltzing" response to impulsive wind* in all oceans, seas, and (as I shall show) in large lakes. In a footnote to a 1931 essay** on internal waves, Ekman proposed that a particular observed example had been generated by "free inertial motions, started by wind or some other disturbing agency and then continuing by virtue of inertia alone". The Scandinavians' early engagement with this type of motion continued with the classic demonstrations of inertial responses in the Baltic Sea (Gustafson and Kullenberg 1936, Kullenberg and Hela 1942); and subsequent research has also characterized inertial motion as a propagating internal wave when the water is stratified. Hence one may speak of inertial waves or inertial oscillations.

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*, **, see footnotes on next page.

Post-Wedderburn developments

Because of the thoroughness with which Wedderburn delineated and interpreted the features of the internal seiche, it is not surprising that Ekman (1931, in the essay cited below**) could write:

> "On the whole it seems to me that there is at present no fundamental question with regard to internal seiches in lakes which is unsolved in its essentials, unless perhaps the influence of the earth's rotation."

In arriving at that conclusion Ekman did not foresee three aspects: the influence of the vertical modal structure (to be introduced briefly in this section); and nonlinear dynamics (the subject of the next section). The question of rotation did, in fact, turn out to be important in large lakes, as illustrated in later sections of this review.

What is surprising is that it took 25 years for limnologists to appreciate, at least in print, the significance of Wedderburn's findings for the chemical and biological economy of lakes. Perhaps the pronouncements of eminent professors (Birge 1910, Halbfass 1909), that the observed phenomena were forced responses to wind action, explains that neglect. But such was the view generally accepted when I was charged, as my first job, with the task of routine measurement and interpretation of seasonal changes in biologically-important chemical species throughout the water column of Windermere in the English Lake District. It soon became apparent that turbulent diffusion was

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*Rossby (1938) pointed out that inertial waves could also be generated as transients when a current, previously in geostrophic balance, is disturbed and seeks a new equilibrium.

**Ekman was one of a notable gathering of experts, convened in Copenhagen by the International Council for the Exploration of the Sea to consider "The Mixing Question, viewed theoretically and practically together with a consideration of internal waves", Rapp. Proc.-Verb. Reunions, 76:1-62(1931). In a short but very influential contribution to that Report ("Internal waves and turbulence in a fluid of variable density") G. I. Taylor analyzed J. P. Jacobsen's concurrently observed profiles of current and density. Taylor explained, for the first time, why turbulent vertical diffusion of momentum in a stratified fluid is more rapid than the vertical diffusion of salinity or heat.

in progress, even in the isolated sub-thermocline water strata, not only during storms, but also during the intervening calm intervals (Mortimer 1941, 1942). The only plausible explanation for flow capable of generating the near-bottom turbulence was to be sought in Wedderburn's temperature seiche.

Wedderburn's findings were later strikingly reproduced by a basin-wide series of observations in 1947 in Windermere, which displayed not only a typical standing wave oscillations of the thermocline (period 19 h in that example), but also fragmentary evidence of a new feature -- a slower sub-thermocline oscillation with greater amplitude and a period near 50 h (Fig. 6). Those findings called for an interpretation which went beyond the response of a two-layered model and its single discontinuous interface.

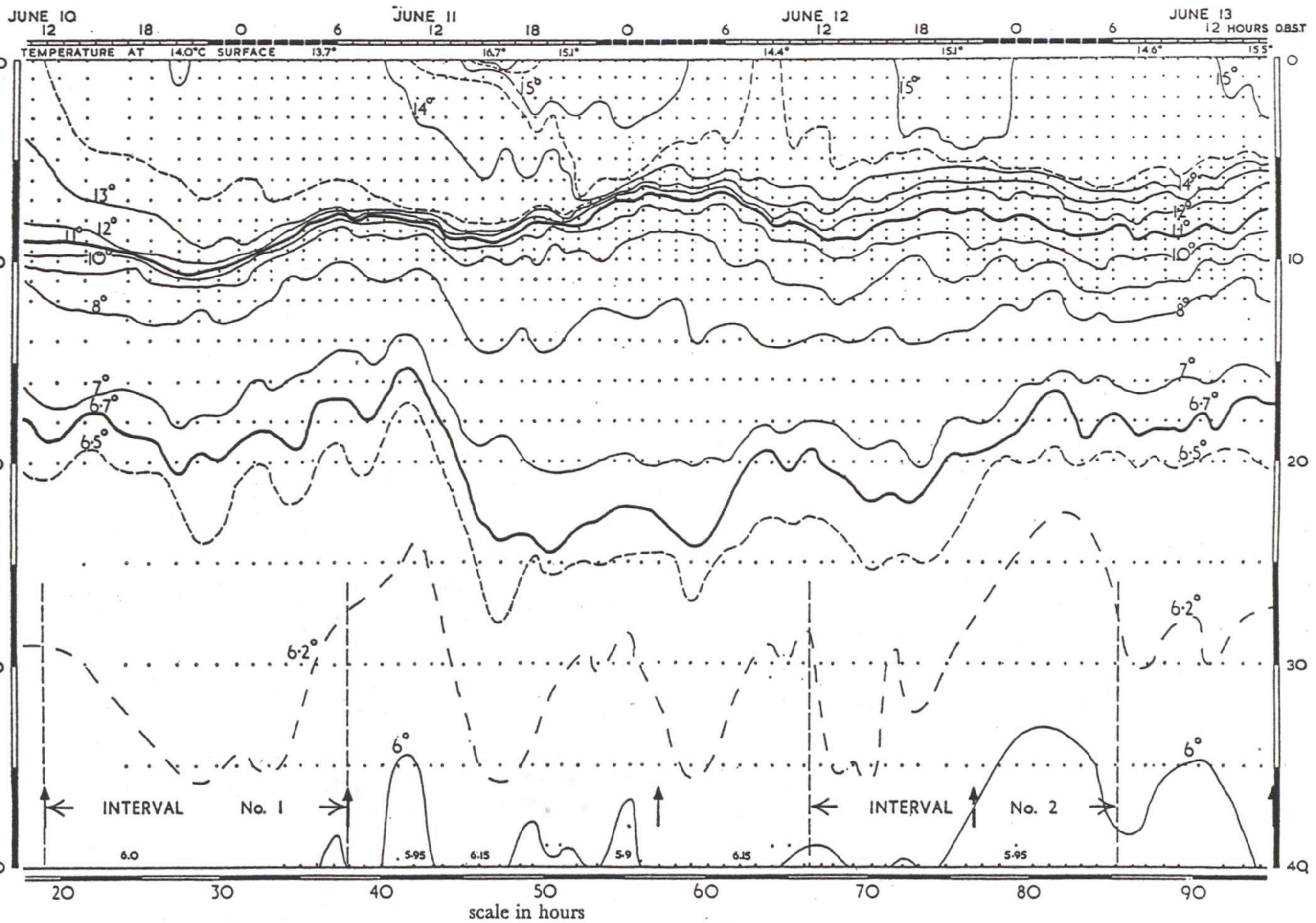
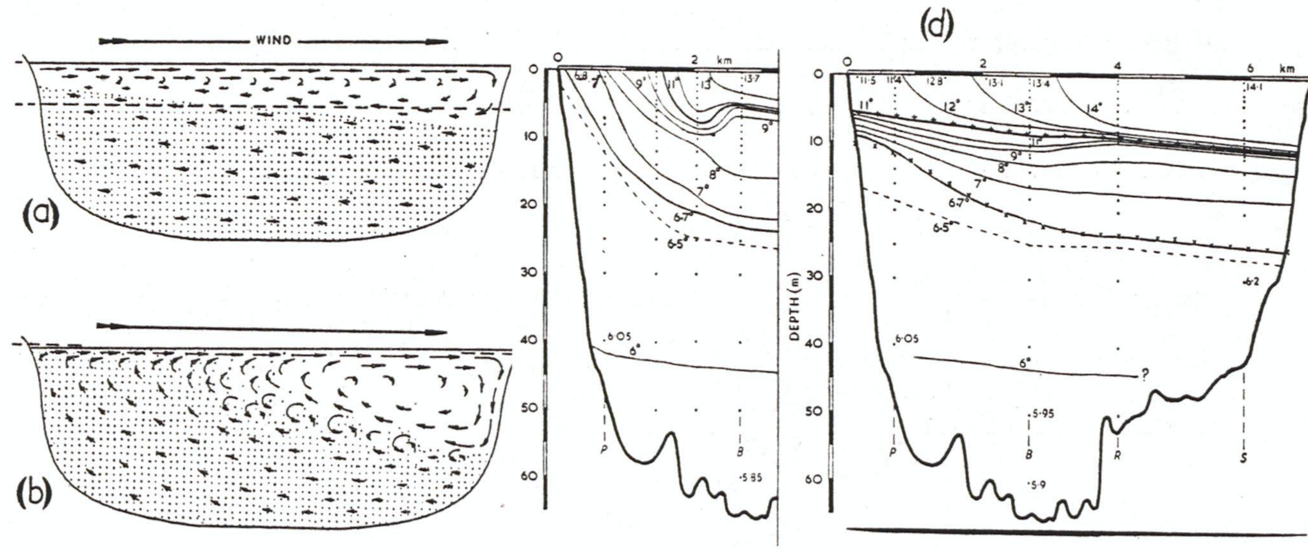


Fig. 6. Internal seiching on the thermocline and in the lower layers of a thermally stratified lake, Windermere, over a 77h interval in June 1947 (Mortimer, 1952).

The presence of a continuously-stratified thermocline, modelled as an intermediate "third" layer of finite thickness, has a dramatic effect on both the wind-forced and the free-seiche phases of the observed response. The forced phase may be represented qualitatively by the sequence in Fig. 7, in which the initial horizontal position of the thermocline is shown by a dashed line. Wind stress, applied at the air-water interface, transports near-surface water downwind, thereby raising the surface level at the downwind end. This generates a return flow directed upwind along the top of the thermocline -- which now begins to tilt -- accompanied by a bodily upwind drift of sub-thermocline layers, shown stippled in Fig. 7. The thermocline tilt increases until near-balance with the applied stress is attained. In



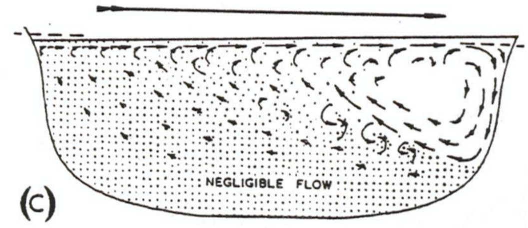


Fig. 7: (a) to (c), qualitative representations of the forcing sequence imposed by a steady wind on motions in a stratified lake (details explained in the text); (d) isotherm distribution in the N basin of Windermere on 9 June 1947 -- left, near noon at the north end at the peak of a gale from NW (force 7 to 8) and, right, along the basin (3 to 5 p.m.) as the storm subsided. Subsequent oscillations of the isotherms are illustrated in Fig. 6. Portions (a) to (c) are extracted from Mortimer (1954) and (d) is from Mortimer (1952).

long basins the thermocline may intersect the surface at the upwind end. Sub-thermocline water is thus brought to the surface by upwelling and becomes incorporated into the downwind surface drift and thereby adds to the volume of the upper mixed layer (Fig. 7b). At the downwind end of the basin that layer forms a mixing wedge with an upwind return current running above the thermo-ocline. At the same time, the upwind drift of sub-thermocline layers comes to a halt (Fig. 7c and in section c of the righthand panel of Fig. 8). Thus a shear zone is formed in the thermocline and, if the wind stress is strong and long-lasting enough, the flow there becomes unstable (low local Richardson number) with the creation of vortices which mix sub-thermocline water into the wedge. Moving upwind, the products of that enhanced mixing produce a fan-shaped distribution of the isotherms similar to that observed in Fig. 7d, which is the forced prelude to the free oscillations illustrated in Fig. 6.

A general conclusion suggested by the Windermere findings was that the course and extent of wind-induced mixing in stratified lakes and the amplitude of internal seiche response, illustrated qualitatively in Fig. 7 and in physical models (Fig. 8), were dependent on three main factors. These were: first, the kinetic energy injected by the applied stress; second, the potential energy incorporated in the pre-existing stratificaiton structure; and third, the ratio of thermocline depth to basin length. Those factors govern the type of mixing which ensues, by determining whether or not the thermocline intersects the lake surface. A modern treatment quantitatively combines those factors in the non-dimensional Wedderburn Number (Thompson and Imberger, 1980). The dynamics of the turbulent mixing process including upwelling, another major chapter in physical limnology not further considered in this review, is treated by Spigel and Imberger (1980) and by Monismith (1986).

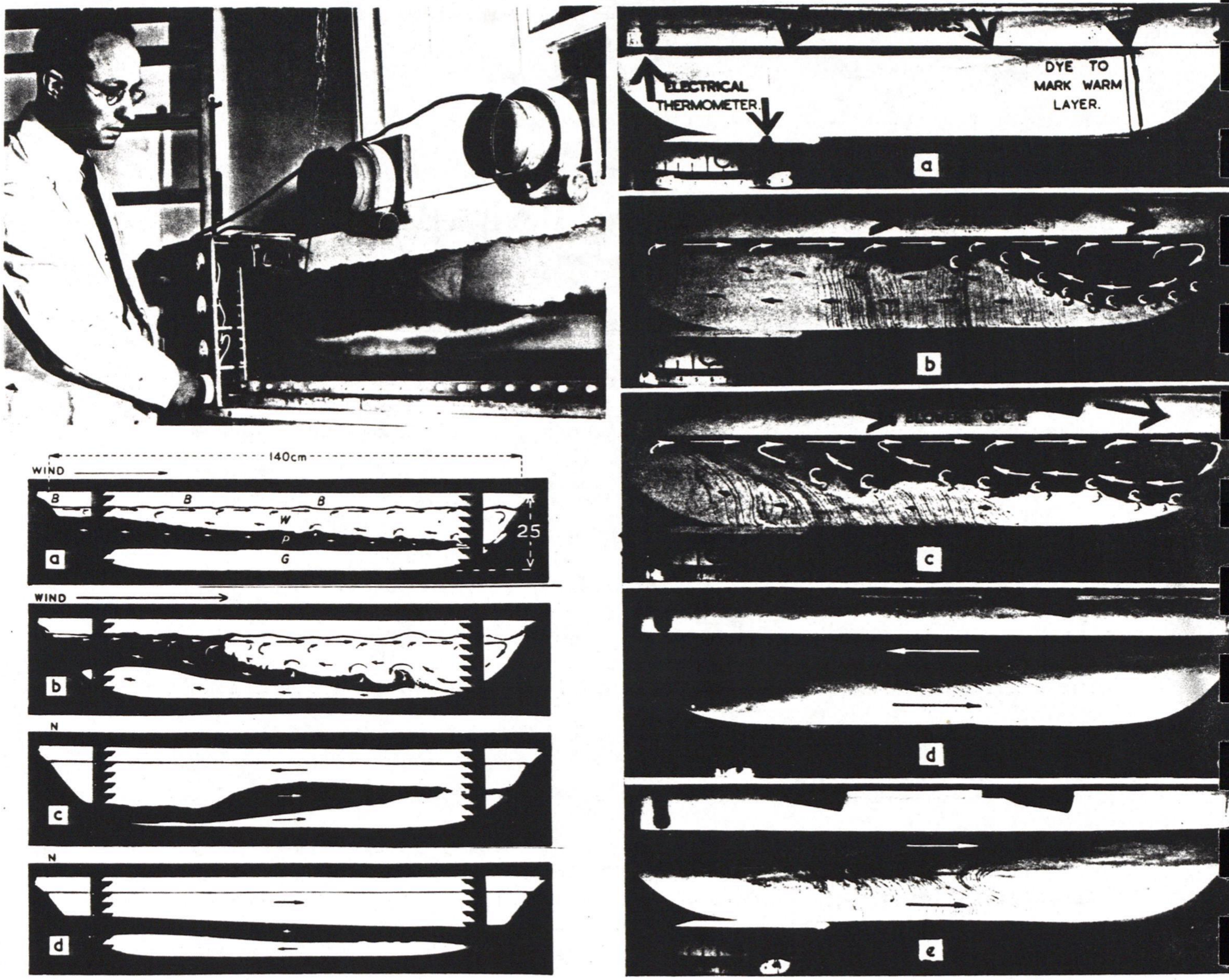


Fig. 8. Wind-forced and subsequent free internal seiche responses to wind of 3-layered and continuously stratified lake models. The upper left panel shows the arrangement of the tank and "wind" blowers (and the author, in a 1950 press photo). The bottom left and the right panels (from Mortimer 1954) illustrate responses during and after wind in, respectively, a 3-layered model (with approximately 3% jumps in density across each of the 2 interfaces) and a model continuously stratified by means of heating wires laid along the water surface. The fluids in the lefthand model were: water, phenol, and salt solution.

The forced and free phases of internal seiche generation, observed in Windermere; were simulated in physical models (3-layered and continuously-stratified, Fig. 8) and in a 3-layered numerical model (M. S. Longuet-Higgins' appendix to Mortimer, 1952). With the initial perturbation (Fig. 7d) as input, that model closely reproduced the principal features of Fig. 6, first, the dominant 19 h seiche on the thermocline and, second, the 50 h oscillation of lower isotherms. The thermocline oscillation, for which the first vertical mode is dominant, is satisfactorily simulated by a 2-layer model (Fig. 4, for example); but the sub-thermocline oscillations involve the second, third, or higher vertical modes and can only be interpreted in terms of multi-layer models or those in which density varies continuously with depth.

Heaps (1961) constructed a three-layered numerical model, in which a linear depth-dependence of density in each of the layers could be adjusted to fit observed profiles. That model -- realistic, but less complex than those in which density varies continuously with depth -- promises to be a useful tool for the analysis of sub-thermocline oscillations involving vertical modes greater than one; but it has yet to be put to work to interpret data sets now available from several lakes. More recently, a three-layer model was constructed (Kanari, 1984) to interpret observations in Lake Biwa, Japan.

a

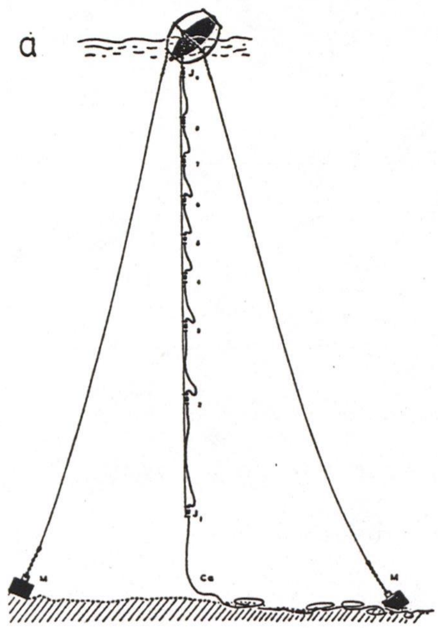
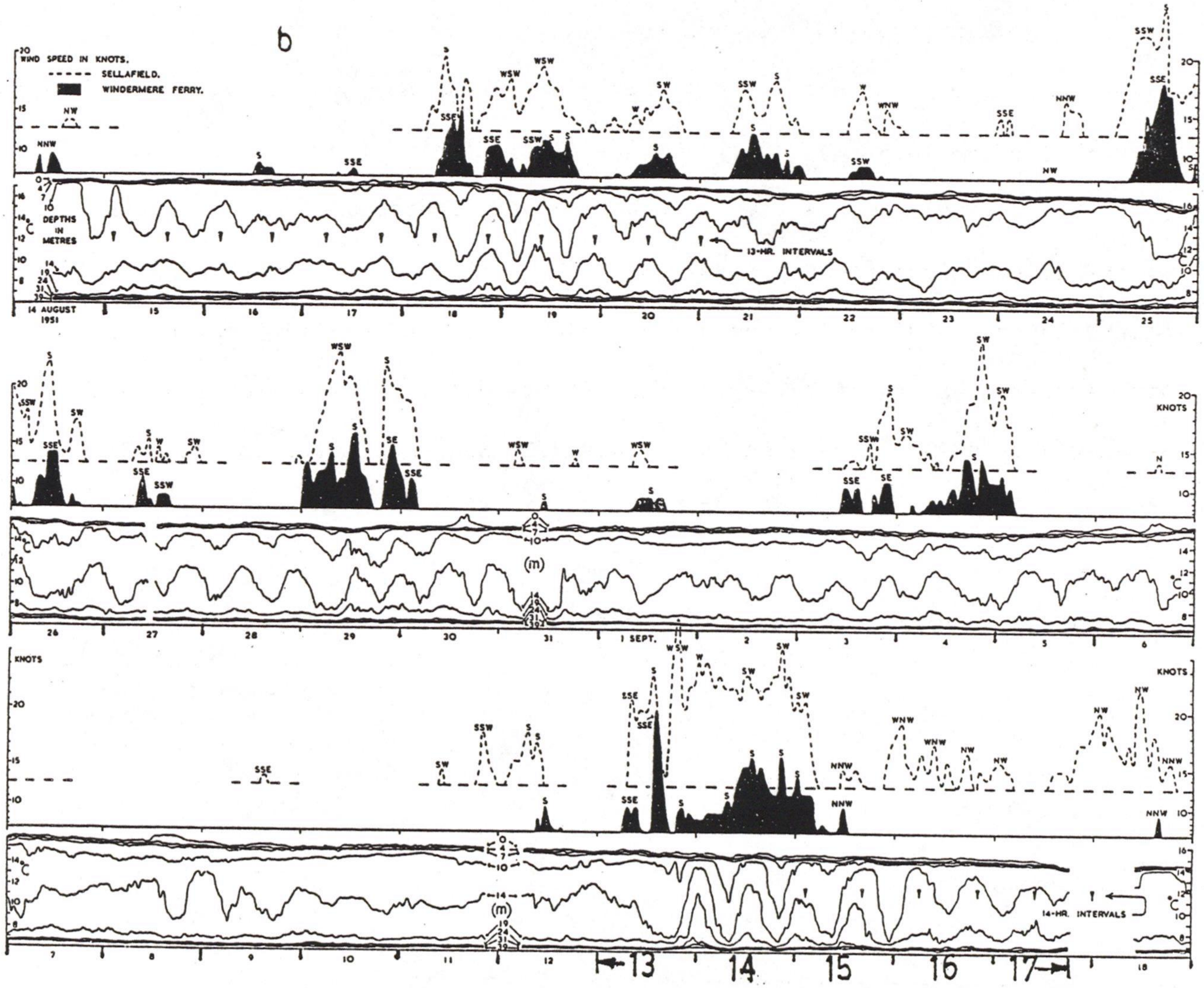


Fig. 9: (a) the prototype thermistor chain*, and (b) the records of temperature (^{o}C) at 9 depths (0-39 m) in Windermere, 14 Aug, 18 Sep, 1951, compared with wind speeds (knots, square-law scale) locally (solid black) and at Sellfield (dashed lines). The periodogram (c) represents the spectrum of "energy" in the 14 m temperature record in (b). Portion (a) is from Mortimer (1951); portions (b) and (c) are from Mortimer (1953).

*(a) was constructed from Navy surplus equipment and cable, to record (onshore) the temperature at 9 depths, repeating every 18 min.

b



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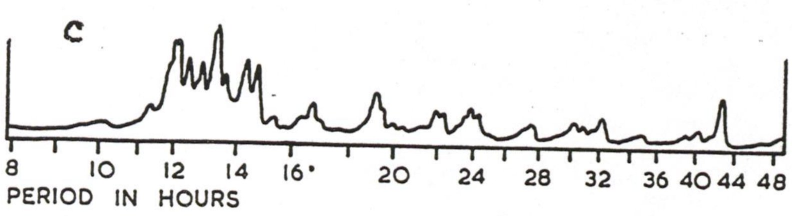


The advent of "continuous" recording with moored thermistor chains

Measurement of long time series of temperature, in which spectral analysis can reveal the frequency, phase, and coherence relationships of oscillations, is virtually impossible with reversing thermometers. The exceptional effort, described in the next section, serves to confirm this statement.

Moored thermistor chains, first deployed in Windermere (Fig. 9) and later widely used in limnological and oceanographic investigations, permitted detailed examination of that lake's response to wind impulses, and confirmed that the uninodal internal seiche was the most common response after a wind disturbance. That was also the conclusion of a survey of the then limited information from other lakes (Mortimer, 1953).

The Windermere data-set enabled N. S. Heaps (1966) to tackle the problem of the lake's forced response to wind fluctuations. His model -- consisting of two layers of differing but uniform density and eddy viscosity and with wind stress applied at the surface and frictional stress applied at the bottom -- showed remarkable agreement with observed responses (Fig. 10).

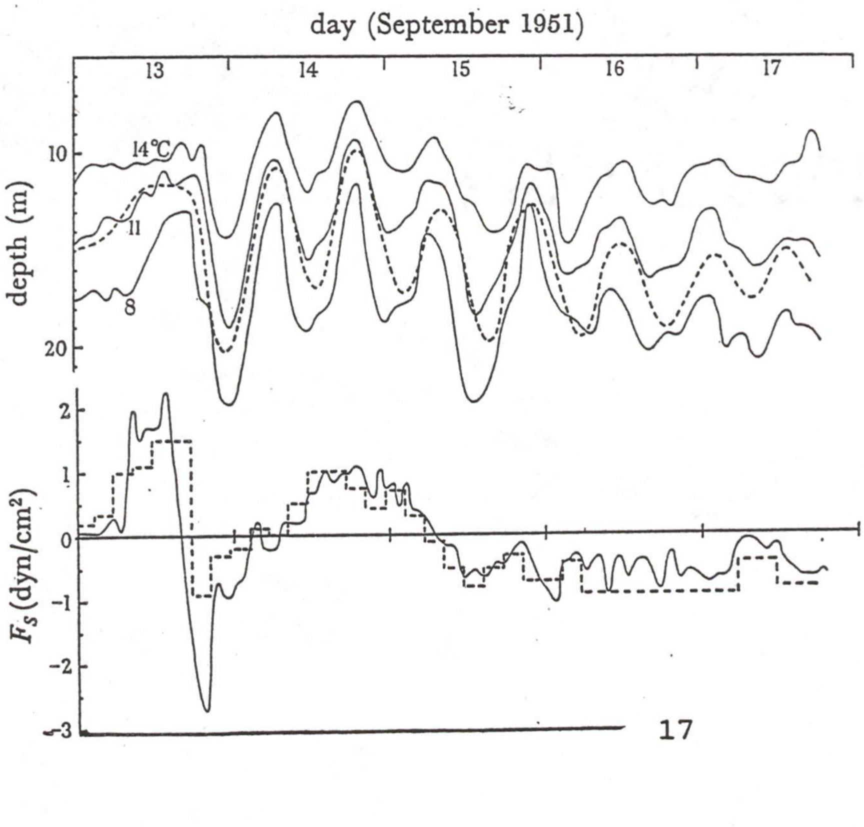


Fig. 10: (a) thermocline depth (dashed line) predicted by Heap's (1966) model compared with isotherm depths (continuous lines) interpolated from temperatures recorded by thermistor chain in Windermere, N. basin, 13-17 Sept., 1951, illustrated in Fig. 9b; (b) the along-lake wind stress component, F_s, deduced from records at Sellafield, 23 miles to the west. Details in Heaps and Ramsbottom (1966).

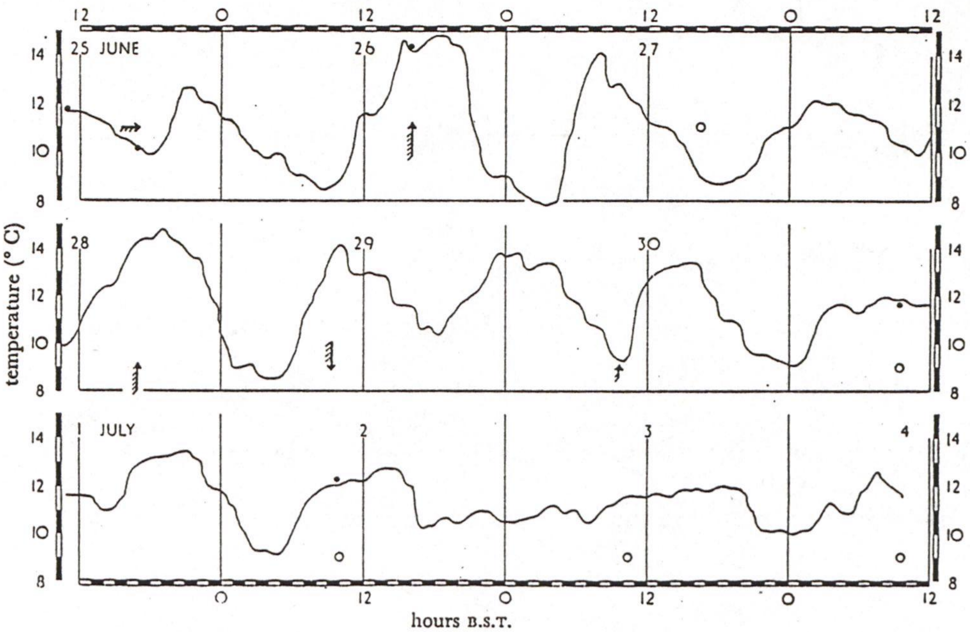
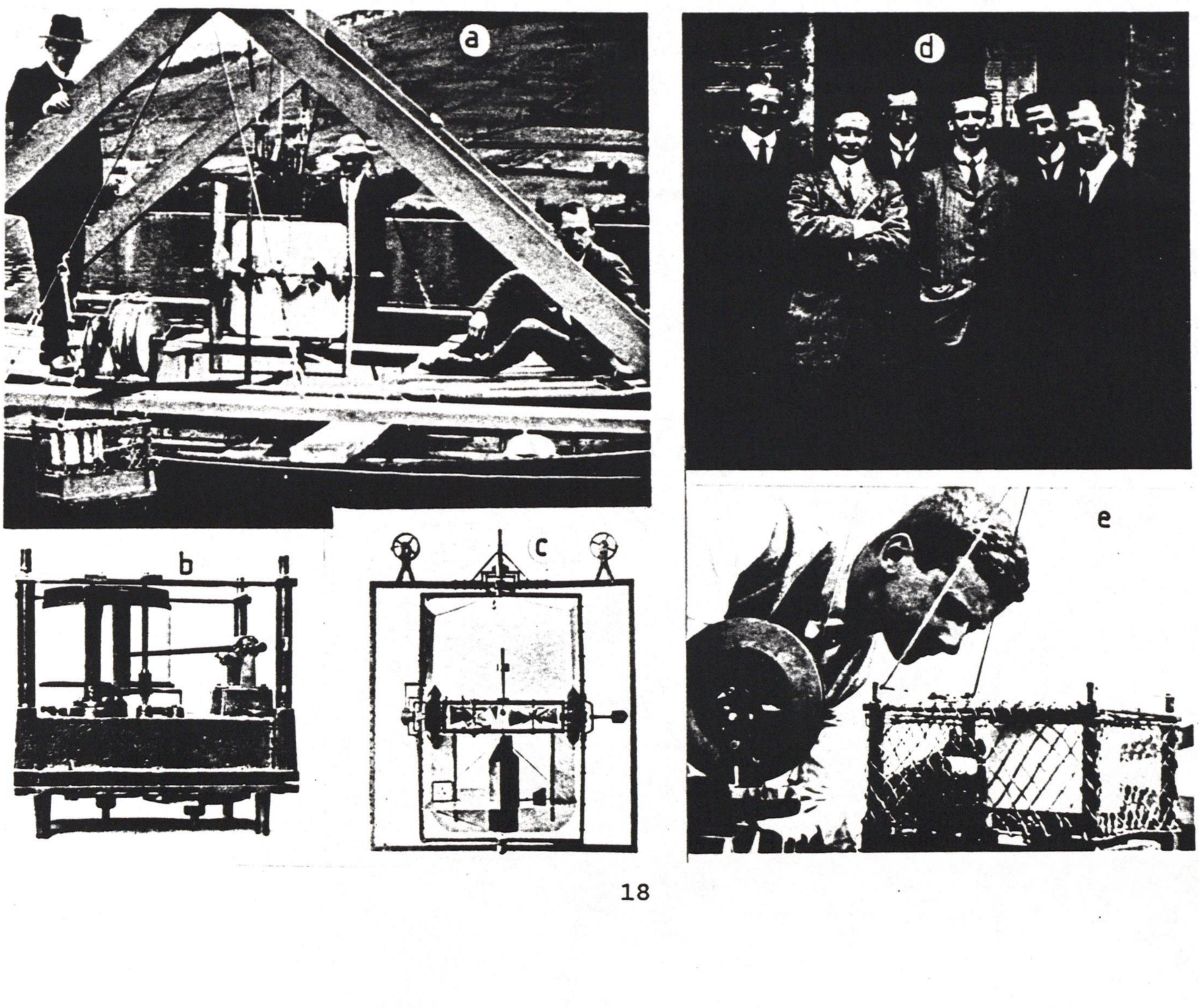


Fig. 11. A continuous record of temperature at 9.5 m depth in Windermere, 1947 (Mortimer 1952).

Fig. 12. (below) The team of investigators and equipment developed for the Loch Earn study (Wedderburn and Young 1915). For details, see footnote on facing page.



Nonlinear features -- internal surges

Previous illustrations include examples in which the internal oscillations deviate from the sinusoidal (linear) form. In most of those cases the isothermal surfaces fall more rapidly than they rise (i.e. fixed-depth temperature rises steeply and falls slowly). Such steep-frontedness arises when the speed of currents induced by the wind or the wave-associated currents are substantial fractions of the speed of wave progress. In some cases that asymmetry is minor -- as in the Fig. 11 example from Windermere -- but in other cases it is more striking, as Wedderburn and Young (1915) discovered in Loch Earn. The crew and equipment, assembled for that study, are photographed in Fig. 12.* Two barges were anchored 650 m apart near the W end of the basin (station A in 45 m, station B in 30 m depth) with six observers working round the clock on each barge. The first series of measurements covered 60 h; the second covered 131 h, during which 9000 measurements were made with reversing thermometers.

The observations, from the first 60 h set and 76 h selected from the second 131 h set, are presented as isotherm-depth fluctuations in Fig. 13. In each case the record from station A (the one furthest from the end of the basin) is placed above the station B record. "A considerable oscillation with a period of about 18 hours appears to have been in progress" or to have just started at the beginning of the 13-15th August record, followed by a regular

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*Fig. 12 illustrates: (a) barge; (b) submersible thermograph; (c) current meter (to measure horizontal and vertical components); (d) group photo with A. W. Young and E. M. Wedderburn at front center; (e) E. M. W. examining the thermograph. (I later borrowed that thermograph from the Royal Scottish Museum to produce the record in Fig. 11.) Photos (a) to (c) are from Wedderburn and Young (1915), photos (d) and (e) by courtesy of Mrs. A. W. Young, located with the help of the Royal Society of Edinburgh.

oscillation of diminishing amplitude. That oscillation was interrupted by 12 h of steady E wind. The average period of the regular oscillation, later in the month (24th to 27th), was longer, about 21 h. "The most noticeable

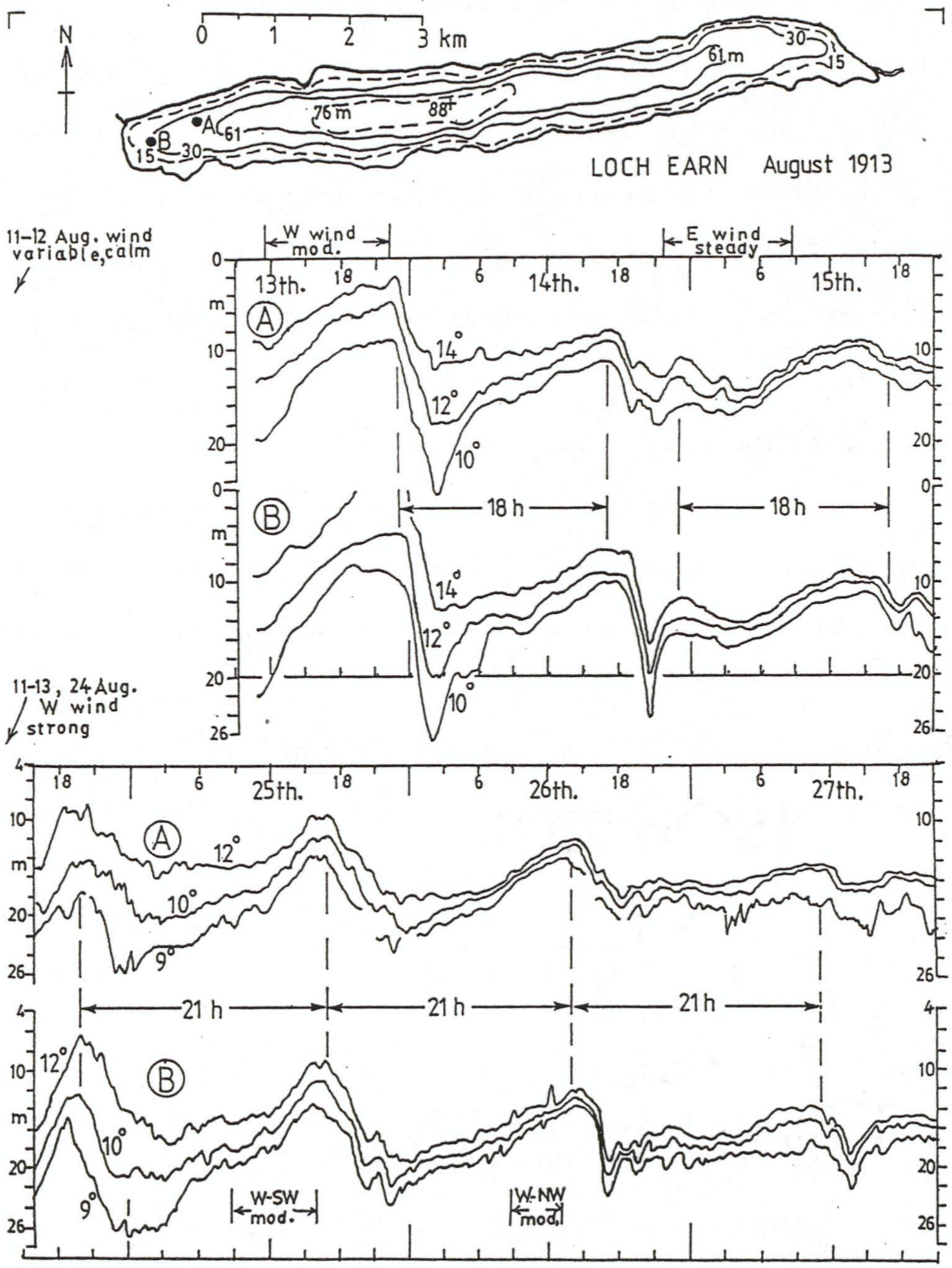
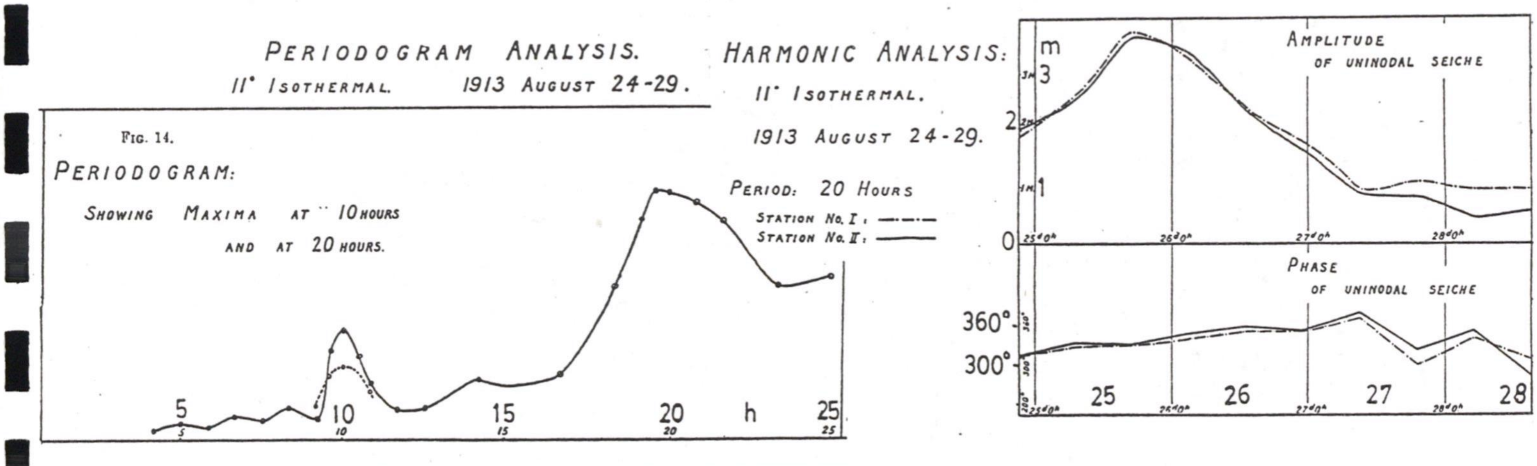


Fig. 13. Depth (m) of near-thermocline isotherms (^{o}C) at stations A and B in Loch Earn, Scotland, 13-15, 24-27 Aug. 1913, interpolated at 5 min intervals from 9000 temperature/depth profiles measured with reversing thermometers (redrawn from Wedderburn and Young, 1915).

point about [both oscillations] is that the isotherms fall very rapidly and rise gradually." Such waves of depression are diagnostic features of an internal surge which, on passing from station A to shallower water at station B, became steeper. This steepening, the authors attributed to "the change in the type of wave in shallow water." Towards the ends of the loch ..., "when the amplitude is large, considerable distortion of the wave surface may be expected." Larger-than-usual currents (6.2 cm/s) were observed during the passage of "this boundary wave or bore."

In order to reveal the principal periodicities present, Wedderburn and Young subjected a series consisting of 140 50-min depth averages of the 11° isotherm at Station A to a periodogram analysis. The result (Fig. 14) discloses a main oscillation of period near 20 h (the uninodal seiche) and a subsidiary peak near 10 h, which the authors interpret as the "binodal" seiche. But the periods predicted for the uninodal and binodal seiches by the Wedderburn (1907) model were 19.6 and 11.0 h, respectively; and it therefore seems more likely that the well-defined 10 h peak in Fig. 14 was not a binodal



When subjected to harmonic analysis the 20-hour oscillation is found to be :—

$$14{\cdot}99 + 1{\cdot}72\cos(\theta - 332^\circ) + 1{\cdot}10\cos(2\theta - 172^\circ) + 0{\cdot}10\cos(3\theta - 185^\circ) + 0{\cdot}14\cos(4\theta - 300^\circ) + 0{\cdot}14\cos(5\theta - 125^\circ).$$

Fig. 14. Facsimile (re-arranged) of figures in Wedderburn and Young (1915).

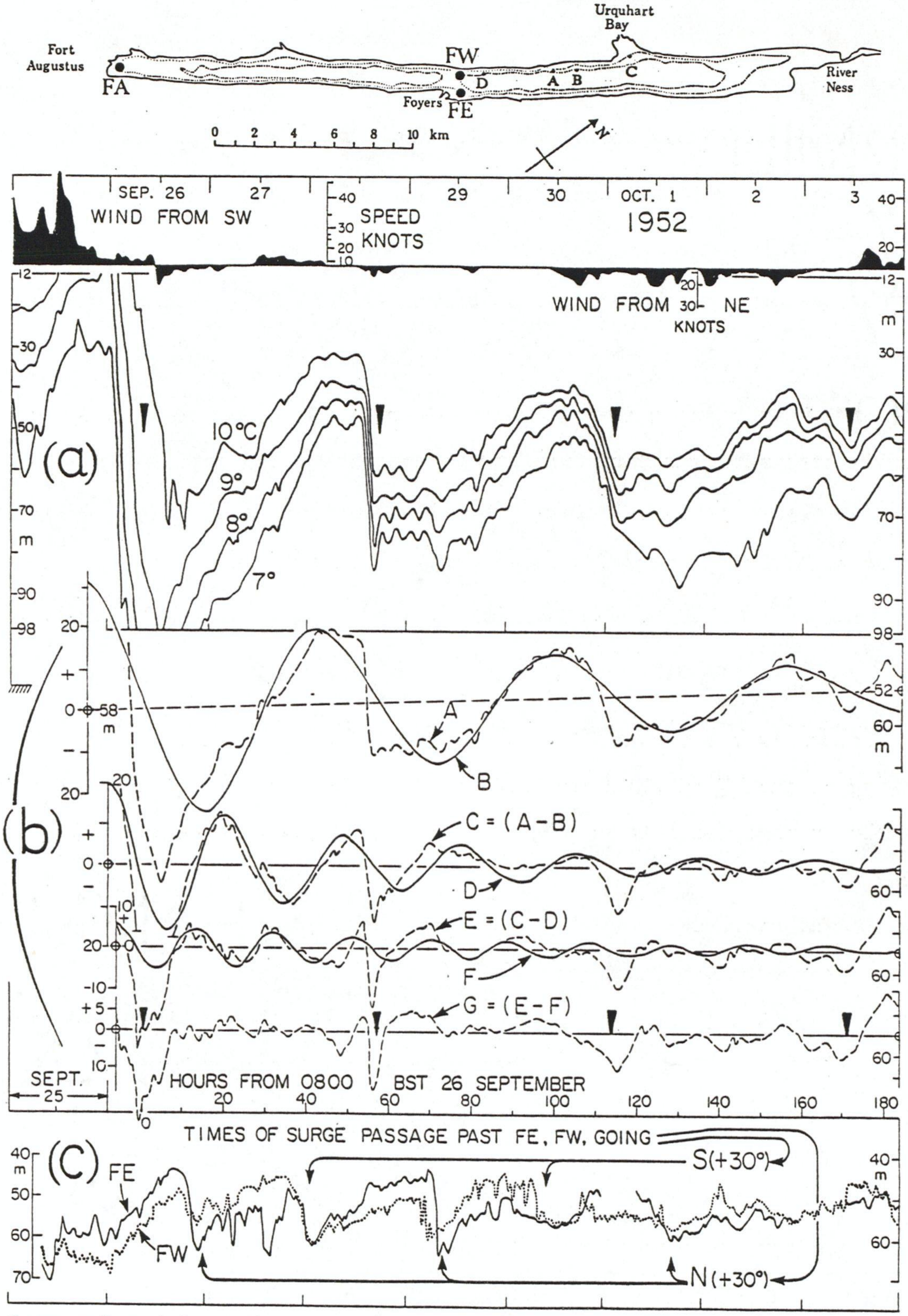


Fig. 15. Internal seiche and surge response after strong SW wind in Loch Ness, Scotland, 25 Sep. to 3 Oct. 1952: (a) hourly mean isotherm depths at station FA, interpolated from thermistor chain records at fixed depths (Mortimer, 1955); (b) graphical harmonic analysis of 9°C isotherm-depth fluctuation (Mortimer, 1979); and (c) 30-min mean depths of the 9° isotherm at mid-lake stations FW and FE, showing repeated passsages of the surge. Solid arrowheads mark the first mode seiche (and surge-passage) period of 57 h.

signal, but was a double-frequency harmonic introduced by the 20 h nonlinear surge wave. Later examples support this conjecture.

Wedderburn and Young also performed harmonic analysis on the same series with a 20 h wave as input. They noted an initial ("uninodal") amplitude of 2.0 m, rising during a windy spell to a maximum of 3.6 m at 1700 h on 25 Aug., then followed by a gradual fall to near 1 m over the next 36 h of relative calm. The phase, it is interesting to note, showed a slow but steady change during the calm interval, thus giving "a very accurate means of determining the true period of the oscillations" (21.1 h in that case). Well ahead of its time, the authors' pioneering procedure would today be called "complex demodulation".

An even more striking surge was discovered when the Windermere thermistor chains were deployed in Loch Ness to study rotation effects mentioned later (Mortimer 1955). In common with the Loch Earn examples and those to be described, the surge in Loch Ness was generated by a (in that case, large) <u>downswing</u> of the thermocline at the SW end of the basin (Fig. 15). Traveling in or nearly in phase with the uninodal internal seiche (period $\simeq$ 57 h), the surge arrived at the NE end near the time at which the thermocline had there swung down to its lowest position. Whether the surge was there reflected, or generated anew and thus locked in step with the downswing, requires further clarification. But it should be noted that, when the main oscillation and its harmonics are filtered out (graphically in Fig. 15b), the to-and-fro passage of the surge signal is seen to be repeated for at least two cycles with decreasing amplitude until the seiche amplitude has, one may speculate, become too small to generate further surges.

Analysis of the strongly-asymmetrical wave in Loch Ness was taken beyond this descriptive stage by Thorpe (1971, 1974, 1977, Thorpe et al. 1972). He employed physical and numerical, two-layered models of internal surges to explain their principal features. Thorpe's profiling current (and temperature) probe, developed for that study, disclosed a new feature: a "packet" of short internal waves which followed directly after the large wave of depression had passed. Similar features have been observed elsewhere (see Fig. 16): in Lake Constance

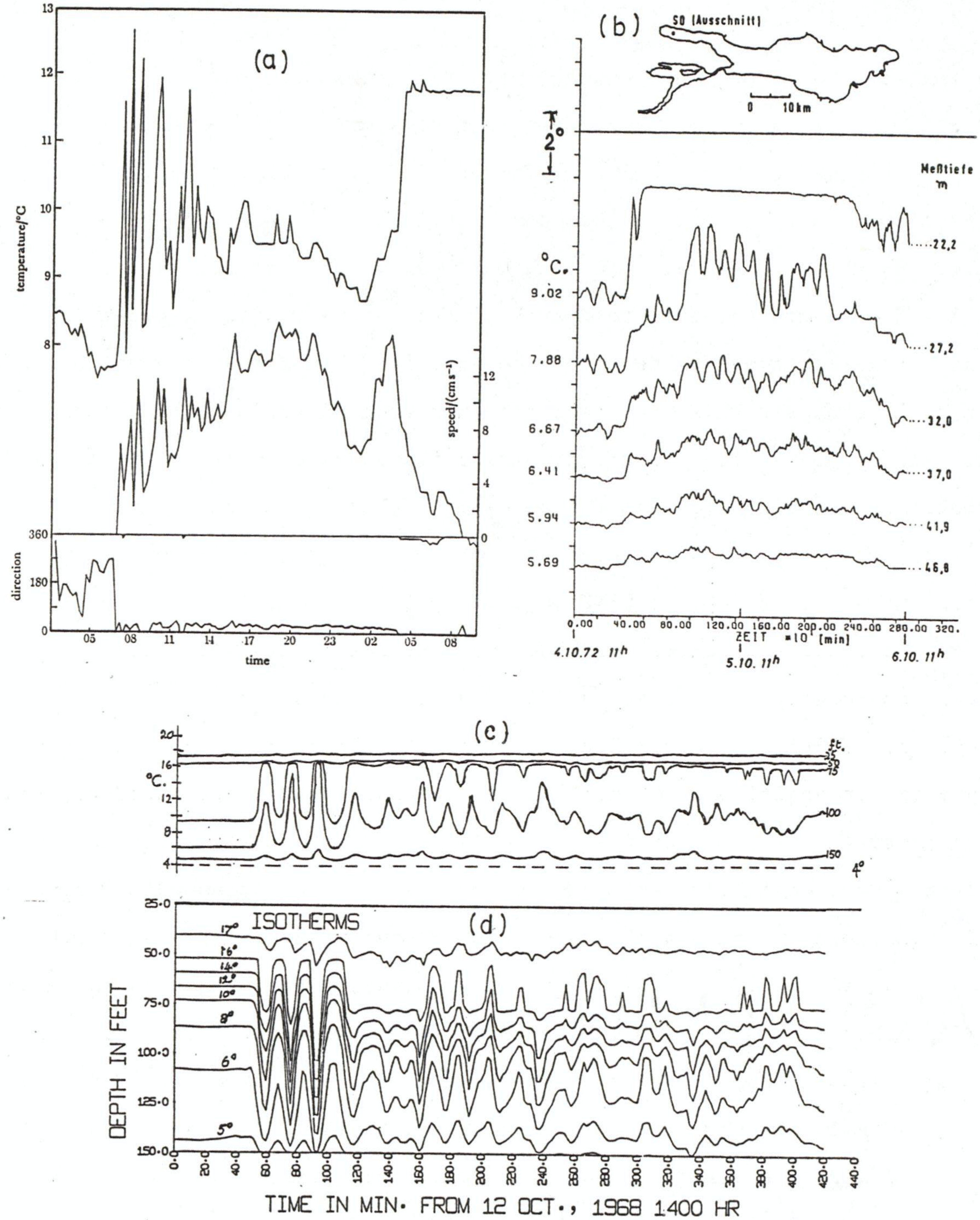


Fig. 16. Internal surges (traveling temperature fronts) and associated undulations in four lakes: (a) Loch Ness, station c, see Fig. 15 (Thorpe, 1977); (b) Lake Constance, station SO (Hollan, 1974); (c) Babine Lake (Farmer, 1978, redrawn); and (d) Seneca Lake (Hunkins and Fliegel, 1973). Illustrated for (a) to (c) are fluctuations in temperature at fixed depths at a mid-lake station. For (d), isotherm depths are interpolated from the fixed-depth temperatures. See also the model surge in progress in c at left in Fig. 8.

(Hollan 1974); Seneca Lake (Hunkins and Fliegel 1973); and Babine Lake (Farmer 1978). In the last two examples, no connection with an internal seiche was evident. The surges were generated by a wind-induced downswing of the thermocline at one end of the basin, and were seen to travel in one direction only. They were not "reflected", as in Loch Ness, from the other end. With a single observing station on Seneca Lake (Fig. 16d), it was not possible to confirm the absence of a seiche or to describe events near the ends. But in Babine Lake nine thermistor chain moorings (not all occupied simultaneously) allowed the progress and form changes of the surge to be followed in sufficient detail to provide a data set, against which Grimshaw (1978) was able to test a two-layered model, and to take cross-channel and along-channel variation of topography into account.

The considerable body of theory relating to nonlinear internal solitary waves and surges is reviewed by Grimshaw (1983) and Mysak (1984) and modeled by Diebels (1991), Schuster (1991) and by Diebels et al. 1993. Apparently, there is a transition from examples of weak nonlinearity (Windermere, Loch Earn) through strongly nonlinear seiches (Loch Ness, Lake Constance, and as later shown, Léman) to cases in which a one-way surge is the dominating or the sole feature. In the latter cases, the inference is that bottom topography at one end of the basin is such that surge reflection or generation is suppressed.

In recent work on the Lake of Zurich (Mortimer and Horn 1982, Horn et al. 1987) it was also found that surges can be generated by large downswings of the thermocline at either end of that basin, but that they emanate more often and more strongly from the N end than from the SE end. An example is illustrated in Fig. 17. Strong wind during the last six hours of 11 September brought about a sharp drop in isotherm level at moorings 9, 10 and 11 coinciding with a rise in isotherm level at 6 and 4. After midnight the thermocline at mooring 11 descended below the bottom thermistor at 18 m; and this strong downwelling stroke generated a surge which traveled away from the SE end at a speed of about 1 km h^{-1} to pass mooring 9 at B, 6 (and nearby 5) at C, and 4 at D. By that time, half a seiche period had elapsed since the start of the oscillation; and the thermocline slope had become reversed,

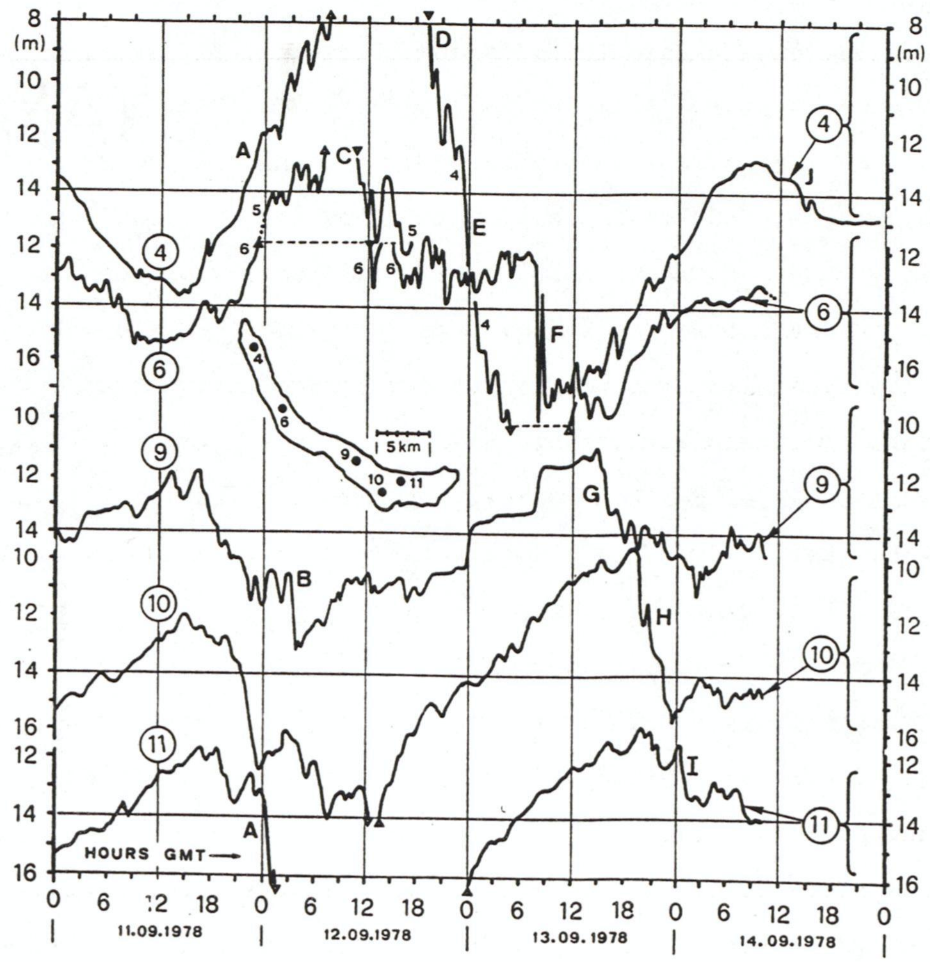


Fig. 17. Depth variation of the 10°C isotherm at mooring stations 4, 6, 9, 10 and 11 from September 11-14, 1978, in the Lake of Zurich. Mooring station numbers, placed near the corresponding traces, are circled. Letters A to J refer to the passages of internal surges past the indicated stations as described in the text.

thereby generating a new surge at the NW end. That second surge then progressed away from the NW end, passing mooring 4 at E, 6 at F, 9 at H, and 11 at I. The fact that the amplitude of the eastgoing surge (H,I) was much less at mooring 11 than at 10, and the further observation that the amplitude difference at those moorings was reversed when the earlier westgoing surge passed at A, are both consistent with Coriolis deflection arising from the earth's rotation.

The results illustrated in Figs. 16 and 17 suggest that, emulating Wedderburn and Young (1915), closer attention should be paid to events and processes near the basin ends, to discover why certain end topographies favor surge generation (or reflection), while others do not.

The influence of the earth's rotation in basins of width order 10 km

The hitherto-described internal seiches and surges were basin-mode oscillations, the structures and periodicities of which were determined by basin morphology, basin dimensions, and by the vertical distribution of density. Generated by wind impulses, those oscillations are also affected by rotation, conspicuously so in basins designated as "large" when basin-width b exceeds the Rossby radius, $a = c_i/f$. For internal waves, the components of that important length scale are: c_i, the internal wave speed* in the absence of rotation, and f, the latitude-dependent Coriolis parameter (inertial frequency). Those terms and the properties of long waves in rotating systems are described and defined in textbooks of fluid dynamics and oceanography (for example, Gill, 1982) and in Platzman's (1970) treatise on tidal waves. As some of those long-wave models (in two-layered constant-depth contexts) will be invoked in later interpretations, their characteristics are, therefore, introduced briefly here.

The influence of the earth's rotation on the predominantly horizontal motions of water masses on the earth is expressed as a deflecting force, the Coriolis force, which is proportional to current speed u and directed (in the northern hemisphere) always 90° to the right of current direction. Therefore, if no other forces act upon a horizontally-moving mass, its track (determined solely by inertia and the horizontal component of the Coriolis force) is a circle (the inertial circle) of radius u/f, traversed clockwise in one inertial period, $12/\sin\phi$ hours. At the latitudes ϕ of Léman (Lake Geneva) and Lake Michigan, to be reviewed later, the inertial periods are near 16½ and 17½ hours, respectively. This is the "inertial motion" referred to on pp. 7 to 9 and later described as "inertial waltzing".

The Coriolis force also acts upon water motions in long waves, conspicuously so in long internal waves with periods of several hours. Gravity is a restoring force and, in the absence of rotation, particle motion in the

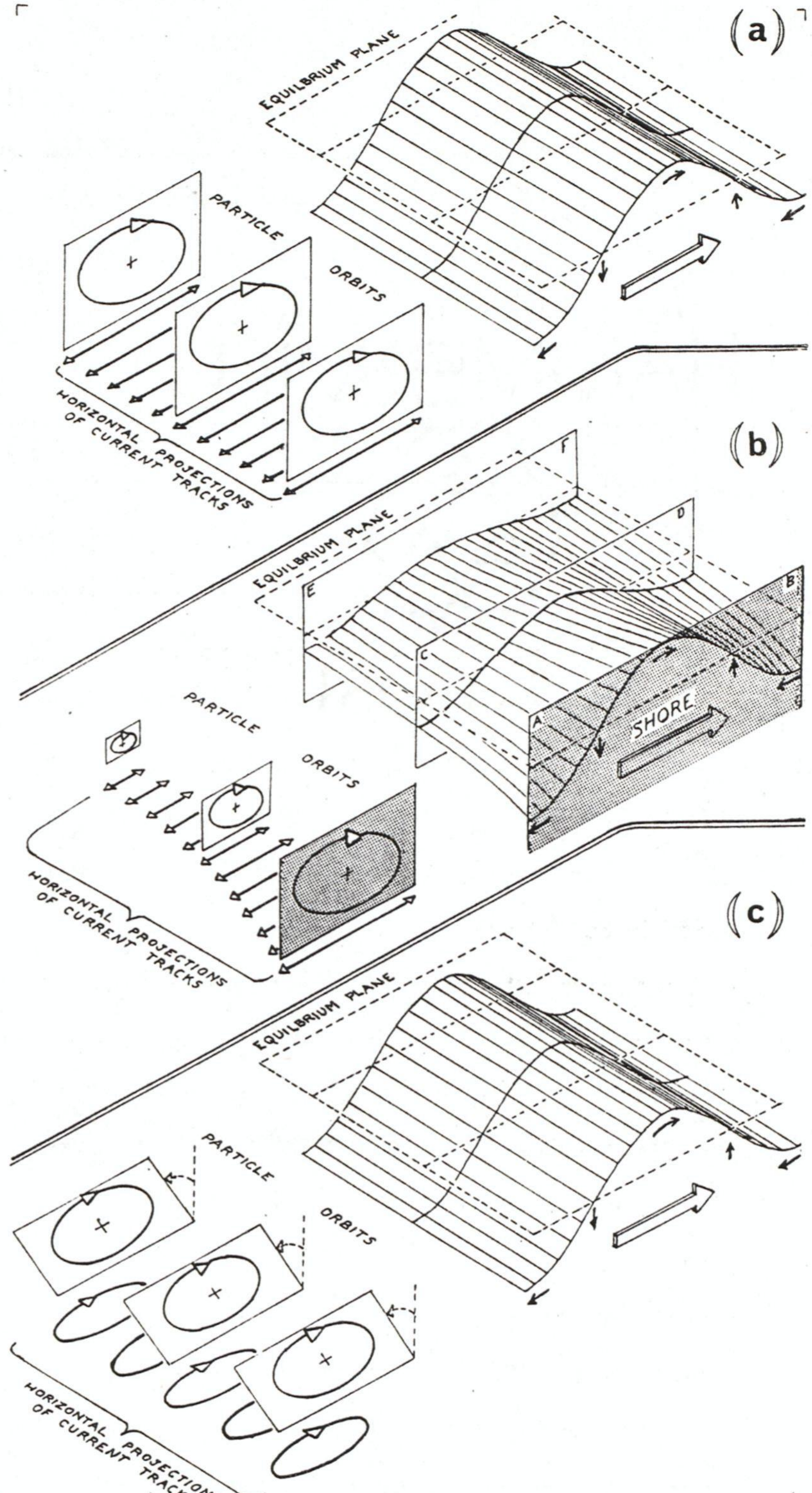
*In the simplest 2-layered model (an upper and lower layer of respective uniform density ρ_1 and ρ_2 with $\rho_2{>}_1$, and respective equilibrium thickness h_1 and h_2) the internal wave speed, c_i, on the interface is the square root of $g(\rho_2-\rho_1)h_1h_2/\rho_2(h_1+h_2)$, in which g denotes the acceleration of gravity.

Fig. 18. Sketches of near-surface current tracks (projected onto a horizontal plane), particle orbits, and waveforms in 3 model waves, each progressing from l. to r. In (a) the influence of rotation is absent or negligible. In (b) and (c) the influence of (N. hemisphere) rotation is fully present.

Model (b) -- a prototype Kelvin wave -- represents long waves generated in the presence of shore barriers (here modeled by a shaded vertical wall).

Model (c) -- a prototype Poincaré wave -- represents offshore waves, which can only exist where shore barriers are absent or sufficiently distant.

In each case, the vertical scale is greatly exaggerated. Applied to internal waves, the waveforms here shown propagate on the interface, and the particle motions are those of the sub-interface layer. Further details are given in the text.



wave is confined to vertical orbital planes oriented along the direction of wave progress (Fig. 18a). Therefore, straight vertical barriers (model shorelines, or channel sides) can be aligned along the direction of progress without impeding or modifying the wave. Rotation, however, imposes an additional restoring force -- the Coriolis force -- perpendicular to the direction of particle motion; and one of two general conditions must be satisfied before rotation-modified waves can exist. Either (i) barriers must be absent, or (ii) if barriers are present (as shores or channel sides) waves are only possible if they conform to particular shore-compatible geometries. Condition (ii) is met by the prototype Kelvin wave modeled with a vertical shore wall, shaded in Fig. 18(b). That wall restricts particle motions to vertical orbital planes. This can only be done by a wave in which amplitude decreases with distance from shore at an exponential rate, such that the pressure forces generated by the sloping wave everywhere cancel the Coriolis forces. Thus, the wave is in geostrophic equilibrium and exists only in association with a shoreline, to which it is "trapped" and along which it can travel, in one direction only, at wave and group velocity c_i, which is (in this case) wavelength-independent. Lake Michigan examples will be described later.

Condition (i) -- absence of shore barriers -- can only be met well offshore, where Coriolis deflection of orbital motion is not counteracted, and the orbits are tilted from the vertical (Fig. 18c). At short wavelengths, the prototype Poincaré wave does not differ greatly from the Fig. 18(a) model; but as its wavelength increases, the following changes become conspicuous: the orbital plane leans more and more (to the left in the N. hemisphere); the celerity rises above c_i; the group velocity (at which wave energy is propagated) falls below c_i; the orbital plane increasingly tilts toward the horizontal; and the wave period approaches the inertial period. At the long wavelength limit (infinity) the orbital plane is entirely horizontal, and particle motion is indistinguishable from inertial motion in a circle, described on pp. 7 to 9. This is the basis of frequent references to inertial or near-inertial oscillations, viewed as limiting cases of Poincaré waves, in which energy is entirely or nearly all kinetic.

It is evident that no barrier could be inserted in Fig. 18c without disturbing the waves. However, particular combinations of standing Poincaré waves can satisfy condition (ii) above and permit shores or channel sides to be inserted (without wave disturbance) along specified lines. For details see Mortimer (1980).

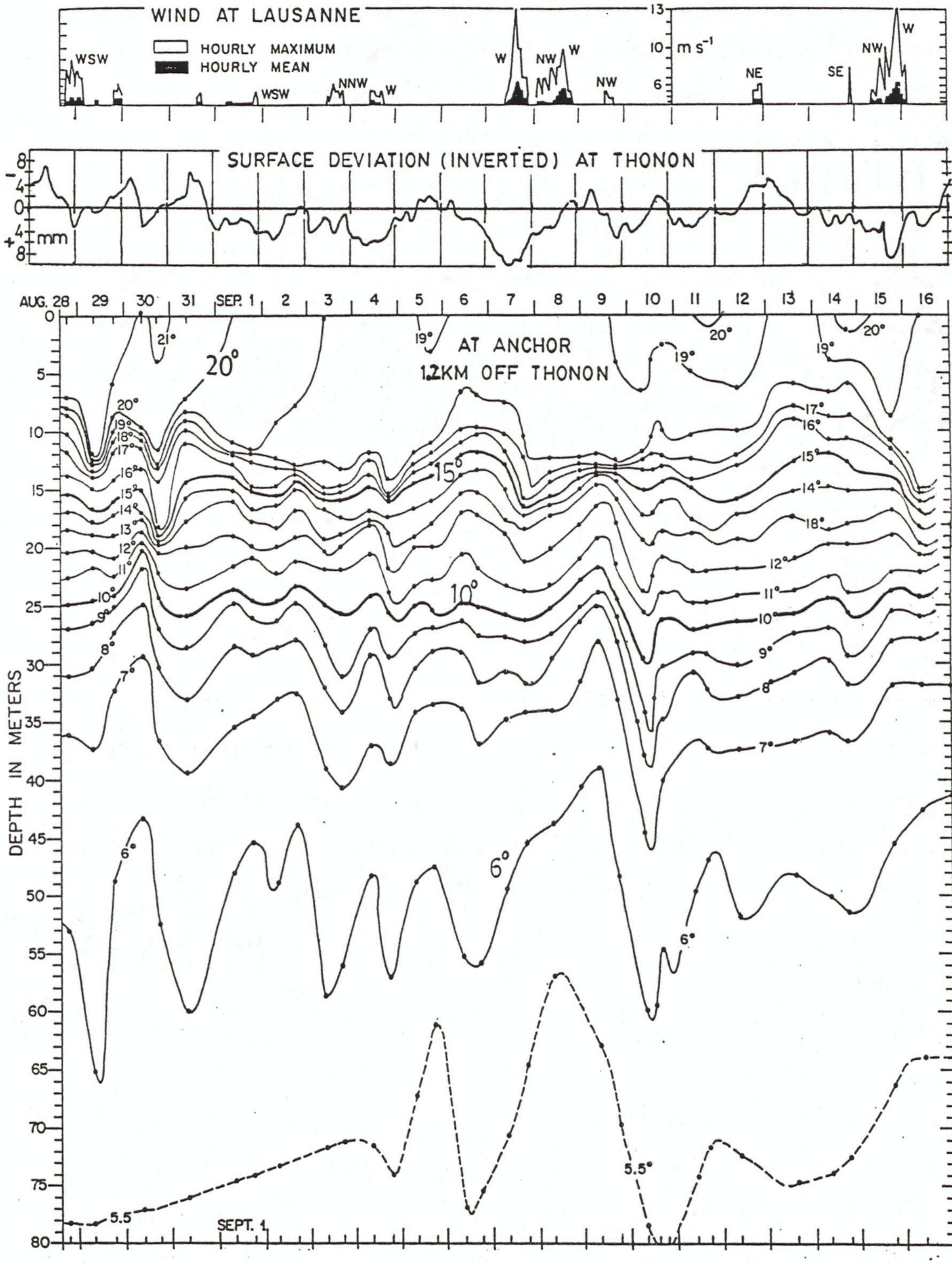


Fig. 19. Léman (Lake Geneva): Fluctuations in isotherm depth based on twice-daily profiles 1.2 km off Thonon (see inset map, Fig. 20), 28 Aug. to 16 Sep. 1950, prepared as described in the text (assembled from figs. in Mortimer 1979).

The exponential rate of which the amplitude of the shore-trapped Kelvin decreases with distance x from shore is $\exp(-fx/c_i)$ in which c_i/f has already been defined as the Rossby radius a for an internal wave. At typical Lake Michigan values of $f = 1 \times 10^{-4}$/s and $c_i = 45$ cm/s, a is 4.5 km, at which distance offshore the amplitude has fallen to 37% of the onshore maximum. At 9 km it has fallen to 14%. Lake Michigan (width b ~130 km) is therefore a "wide" basin or channel for internal Kelvin waves. The mid-Lake amplitude is vanishingly small (see later Fig. 27b, p. 45).

The first evidence of an internal Kelvin wave response in a lake came from Léman (Lake Geneva, Switzerland) where, in 1950 in collaboration with Dr. Bernard Dussart, I measured temperature profiles off Thonon twice-daily for fourteen days during August and September (Fig. 19) and also examined temperature records during that and previous years from the 15m-deep water intake of the City of Geneva at the SW end of the Petit Lac. The combined evidence suggested that the response to strong SW wind over Léman was, first, a surge-like depression of the thermocline along the southern shore, followed by counter-clockwise progression of a shore-trapped internal Kelvin wave at approximate speed c_i (Fig. 20). One or two whole-basin circuits were completed in about four days each, if the post-storm calm persisted for long enough (Mortimer 1953, 1963).

However, after a careful statistical analysis of a later, much longer (35 month) series of Léman current and temperature records, Bohle-Carbonell and van Senden (1990) have recently concluded that basin-mode models are inappropriate interpreters in this case. In such models, the authors pointed out, "the motions are organized on length and time scales determined by the lake basin" with "a flow field which is highly coherent in time and space". But "stringently conducted spectral analysis" of the Léman records failed to find such coherence; and the authors concluded that "currents in large lakes, at least those of Lake Geneva, may be best described as transient in time and only locally organized in space." While this statistic, from those particular

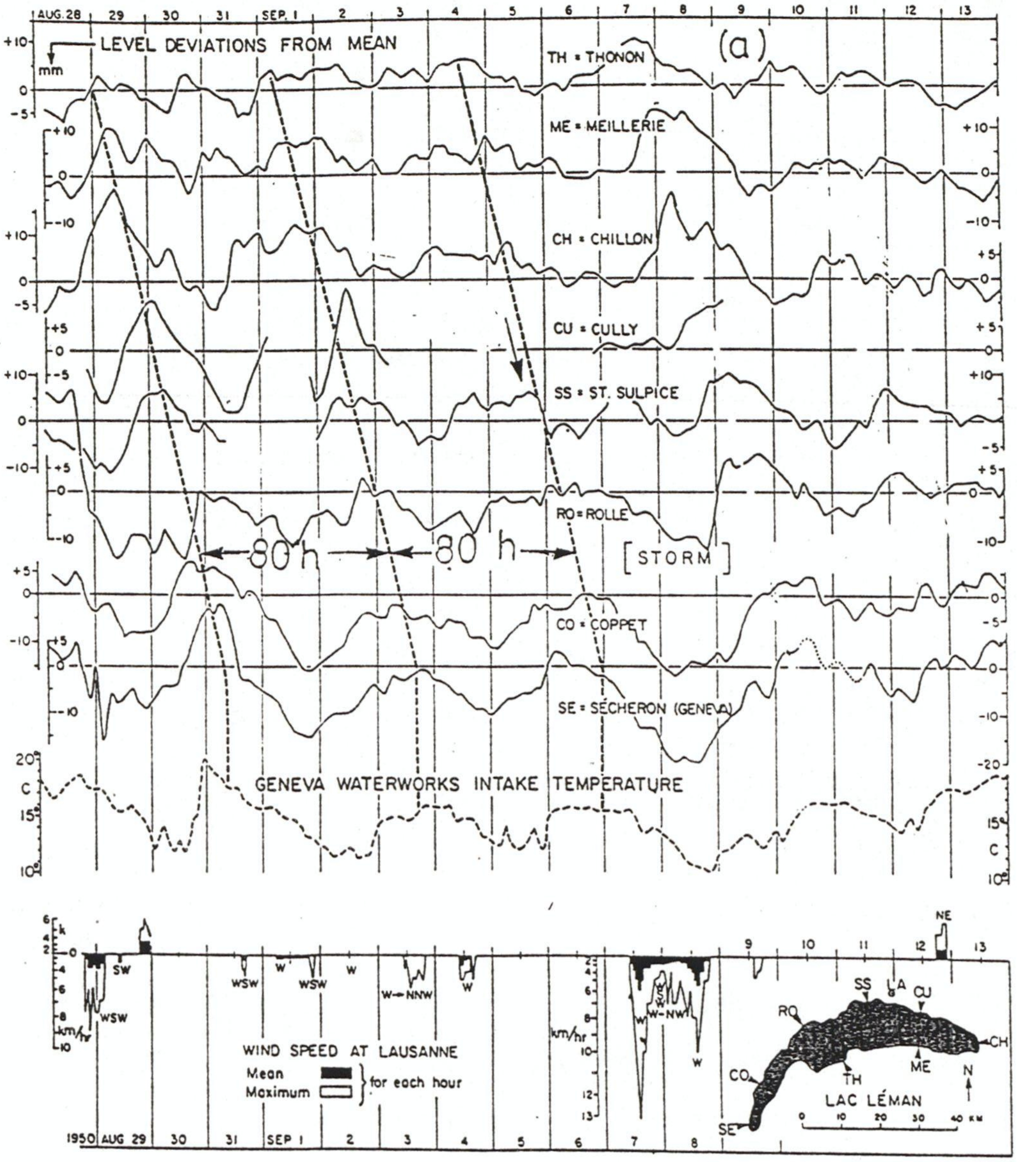


Fig. 20. Léman (Lake Geneva) 1950: Deviations of surface level at eight shore stations, computed as described in the text, compared with wind records at Lausanne and with temperature at 15 m depth at Sécheron (Geneva waterworks intake, shown as a broken line). Source, Mortimer (1963).

groups of recorders, is evidently true for most of the time, it is not (I believe) the whole picture.

First, it should be pointed out that wind-forced upwelling or downwelling -- deflection of the thermocline interface from its equilibrium level -- and the oscillations which follow are accompanied by much smaller but proportional deviations of the surface water level from equilibrium out of phase with the interface deflection. Therefore, in Léman, a thermocline deviation of -1m from equilibrium level may be expected to be accompanied by a surface deviation of about +1mm from equilibrium.

It was, therefore, a fortunate coincidence that the Fig. 19 temperature observations were made in a year (1950)in which the Swiss Service Fédérale des Eaux (1954) had installed sensitive water level recorders around the lake shore, attempting a more precise match between the land levels of France and Switzerland. After low-pass filtration of those surface level records, to remove short-period fluctuations including surface seiches, and after subtraction of the whole-lake mean level, the (inverted) surface deviation reproduced the main features of thermocline depth fluctuation, approximately at Thonon (Fig. 19), more exactly at Sécheron (near Geneva) and, by inference, also at six other shore stations around the basin (Fig. 20). (See Caloi et al., 1961, and Sirkes 1987, for demonstrations of similar interface/surface level correlations in Lake Bracciano and in the Dead Sea, respectively.)

When the deviations at eight stations around the basin were plotted in counter-clockwise order and compared with wind observations at Lausanne (Fig. 20) the positive deviation (thermocline depression) at Thonon, coincident with the 28/29 August wind impulse, was followed by a counter-clockwise progression of positive deviations, attaining their greatest amplitudes at the extremities of the basin (Chillon and Sécheron) and completing two cycles (two basin circuits) each of about 80h duration. further progress was apparently stopped by the 8/9 September storm, which set another cycle in motion, again starting with a positive deviation at Thonon and continuing with the counter-clockwise progress of a steep-fronted surge. The interfacial origin of those surface-deviation waves was indicated, not only by Fig. 19, but also by the remarkably close correlation in Fig. 20 between the surface deviation at Sécheron and the temperature at the nearby Geneva water intake at 15m, near thermocline depth.

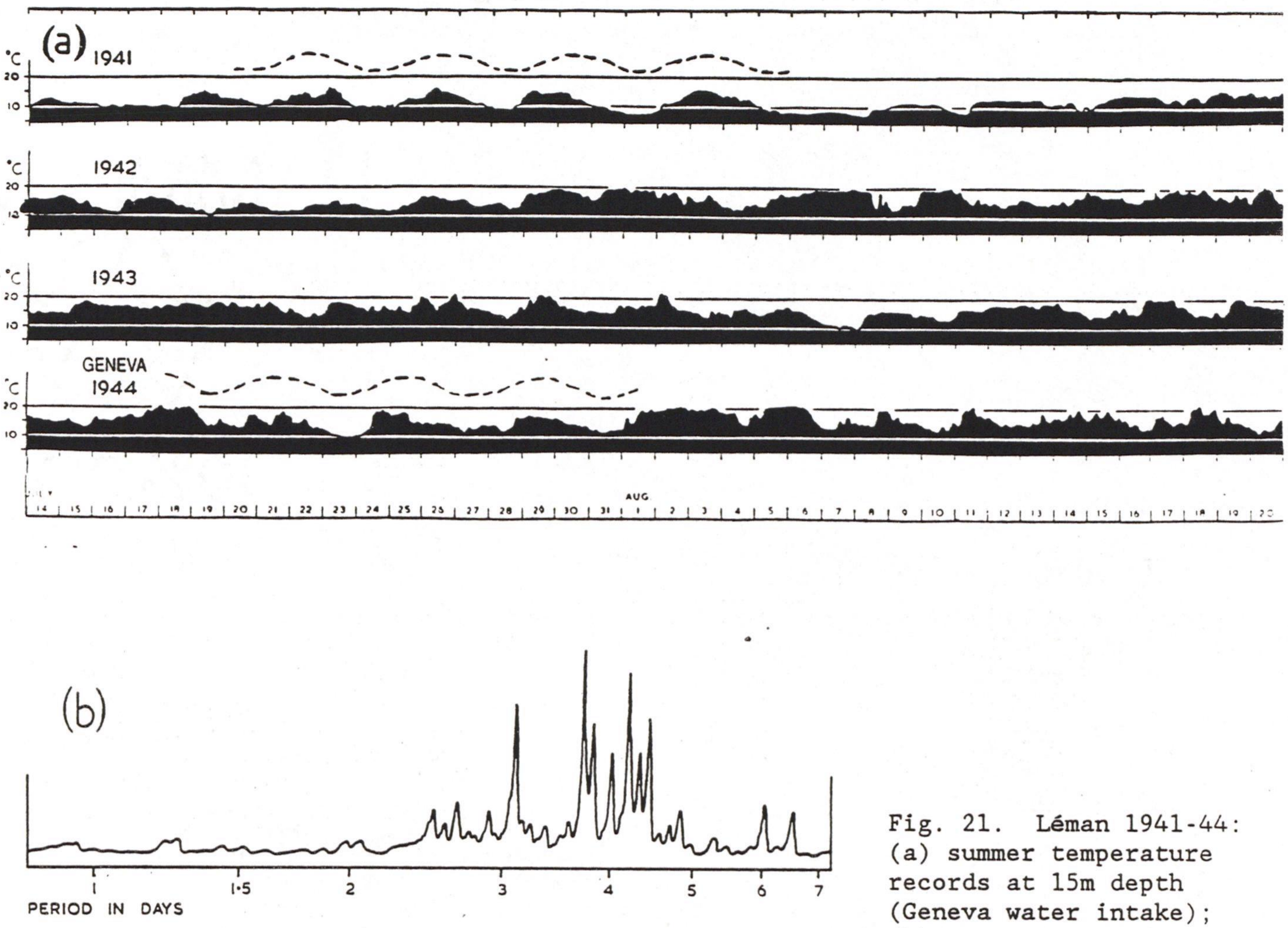


Fig. 21. Léman 1941-44: (a) summer temperature records at 15m depth (Geneva water intake); (b) periodogram of (a). (Mortimer, 1953).

Examination of temperature records from the City of Geneva water intake (15m) for summers prior to 1950 (silhouettes in Fig. 21) reveal a complex mixture of predominantly aperiodic fluctuations, combined with occasional episodes, during which an underlying long-period oscillation persisted for three or more cycles and then faded away or was interrupted by a newly-generated (?) large fluctuation out of phase with its predecessor. Such episodes, of period 3 to 4 days marked by broken sinusoids in Fig. 21, were attributed to thermocline waves. A periodogram, prepared from the silhouettes (with equipment developed for the analysis of ocean waves, Barber et al., 1946), showed that most of the energy was concentrated in the period range 3 to 5 days (Fig. 21b).

The findings displayed in Figs. 19 to 21 I attributed to the excitation and progress of an internal Kelvin wave illustrated by the model in Fig. 22, in which the "thermocline" is a shaded surface. Without rotation the "thermocline" would swing as a see-saw about the hingeline; but, with rotation, the rightward deflection of the upper and lower layer currents imposes a transverse oscillation, as shown in mid-basin, about the pivot (amphidromic point) P. Such a shore-trapped wave, traveling at internal wave speed c_i calculated for the temperature structure in Fig. 19, would complete the circuit of the basin in 84 hours.

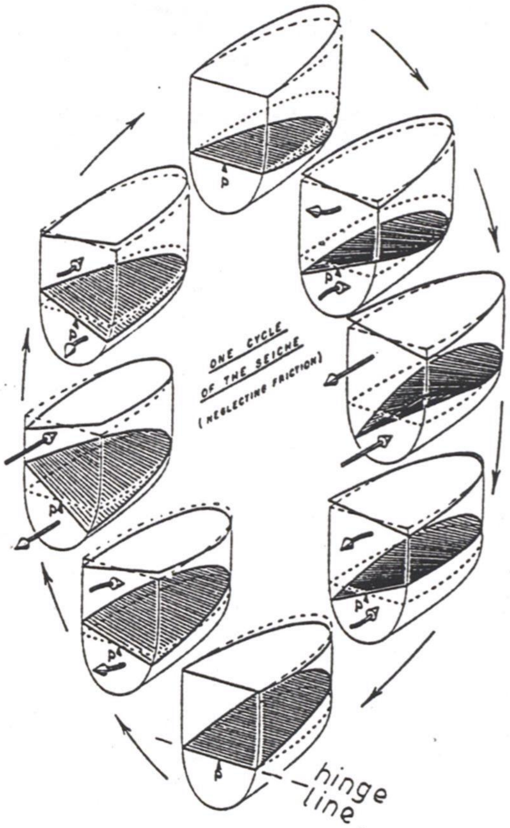


Fig. 22. A rotating two-layered model of (half a) lake to illustrate the thermocline oscillation (shaded) and the coupled oscillation in surface level (unbroken line). Broken lines indicate the equilibrium levels of surface and thermocline. P is the amphidromic point (see text). Long thin arrows indicate the oscillation sequence; conical arrows depict current at mid-basin in the upper and lower layers.

Nearly thirty years elapsed before the above speculations could be tested on summer records obtained at a mooring 293m deep 2.7km off Lausanne by instruments described in Graf et al. (1979). An extract of the temperature record at 9.8 and 19.1m is presented in Fig. 23(a) with a spectrum of fluctuations in the 19.1m record in Fig. 22(b). The latter, produced with the program of Kielmann et al., 1979), shows a distinct peak at 2.2 cy/d and a broad concentration of energy between 0.2 and 1 cy/d (Mortimer et al., 1984).

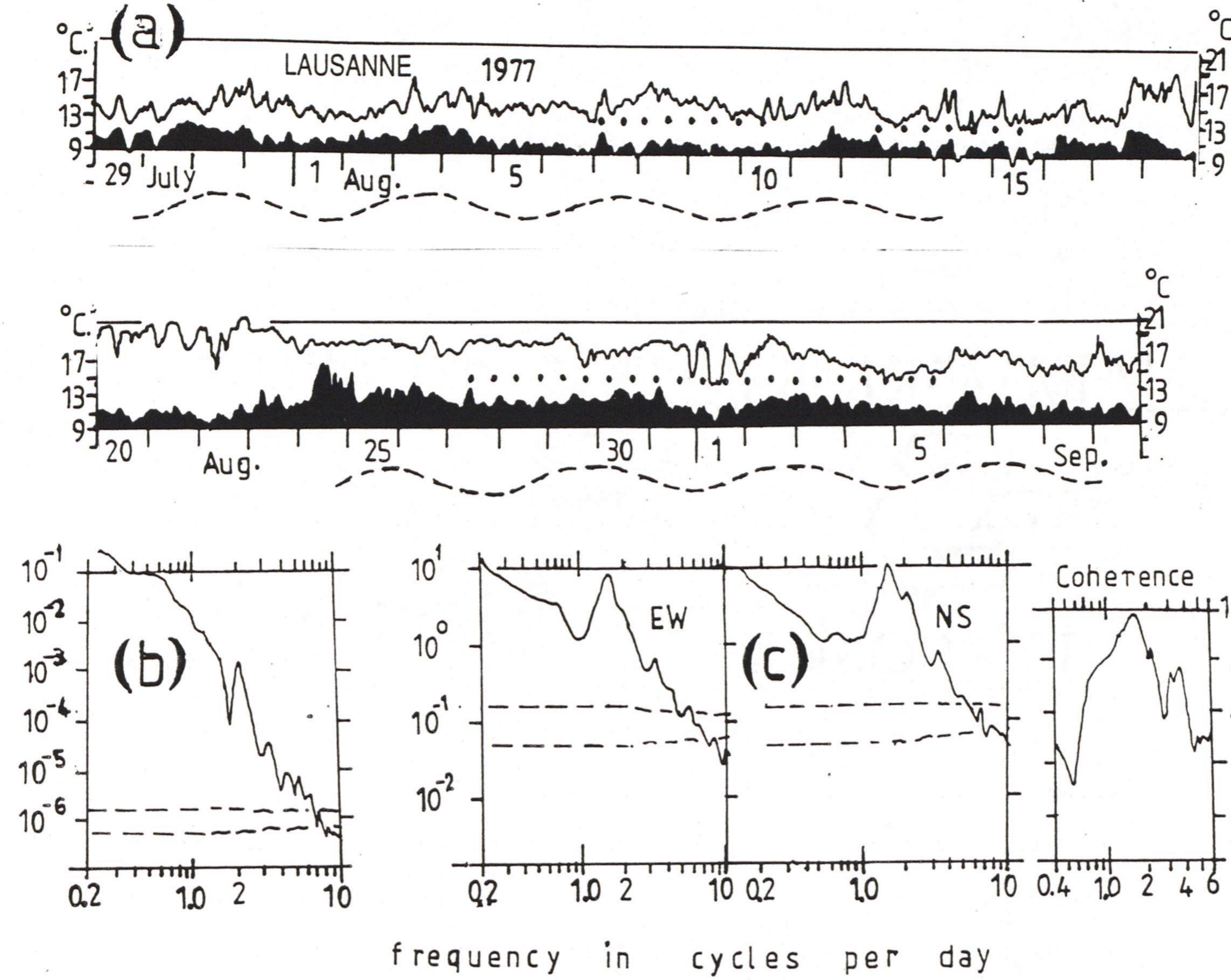


Fig. 23. Léman 1977: (a) temperature at 9.8 and 19.1 m (black silhouette) 2.7 km from Lausanne (water depth 293 m) 29 July - 9 Sept.; (b) spectrum of temperature fluctuations at 19.1 m at the Lausanne station, 17 June - 12 Aug.; (c) spectra of EW and NS current components and their coherence at 19.1 m at the Lausanne station,, 17 June - 3 Aug. 1977. The broken lines in (c) indicate 95% confidence intervals. (From Mortimer, Perrinjaquet, and Bohle-Carbonell, 1984). The broken-line sinusoids in (a) are added here to indicate a 4-day periodicity; and the dots are at 11 h intervals.

The temperature fluctuations off Lausanne in 1977 were less regular and of much smaller amplitude than those at the 15m Geneva water intake in 1950; but there were occasional episodes in 1977 which coincided with abrupt increases in current fluctuations (not illustrated) presumably in response to wind impulses. There were also occasional bursts of relative regular oscillations of approximate period 11h (2.2 cycles/day) indicated by dots in Fig. 23.

The current records off Lausanne are not illustrated here; but their spectra, Fig. 23(c), show an additional feature -- a concentration of energy near 1.7 cy/d. The phase relationship (not illustrated) and the high coherence at 1.7 cy/d are consistent with clockwise rotation of the current vector, a characteristic of internal Poincaré waves. Why the current energy should peak at 1.7 cy/d while the temperature "energy" peaks at higher frequency (2.2 cy/d) awaits a tested explanation. Although those oscillations and the occasional 4-day temperature "waves" seen in Fig. 23 may be signals from basin modes, Mortimer et al. (1984) had to conclude, from an examination of all records then available, that (i) Léman "is particularly sensitive, not only to the wind force, but also to the duration and timing of the impulse", and (ii) that the "responses are episodic, short-lived, and characteristically intermittent". That conclusion was reinforced by Bohle-Carbonell and van Senden's (1990) statistical analysis of a much longer (35 month) series of temperature and current records from later years. But it should be noted that their recorders were deployed not in optimal arrays for the detection of basin modes, but were localized small groupings designed to answer other questions. Also, to accommodate a commercial fishery, recording had to be confined to Oct. to Mar. (1981-83), seasons of weak stratification.

Another major complication was the strong non-uniformity of the wind stress on the Grand Lac, sheltered from W and SW by the mountains of Savoy. Other complicating factors, to which Bohle-Carbonell and van Senden drew attention, were: differential warming which gave rise to local winds and local current fields; the effect of shore topography ; and nonlinear interactions which exchanged energy between motions of different space and time scales (Bohle-Carbonnel & Lemmin, 1988). Nonlinear interactions were

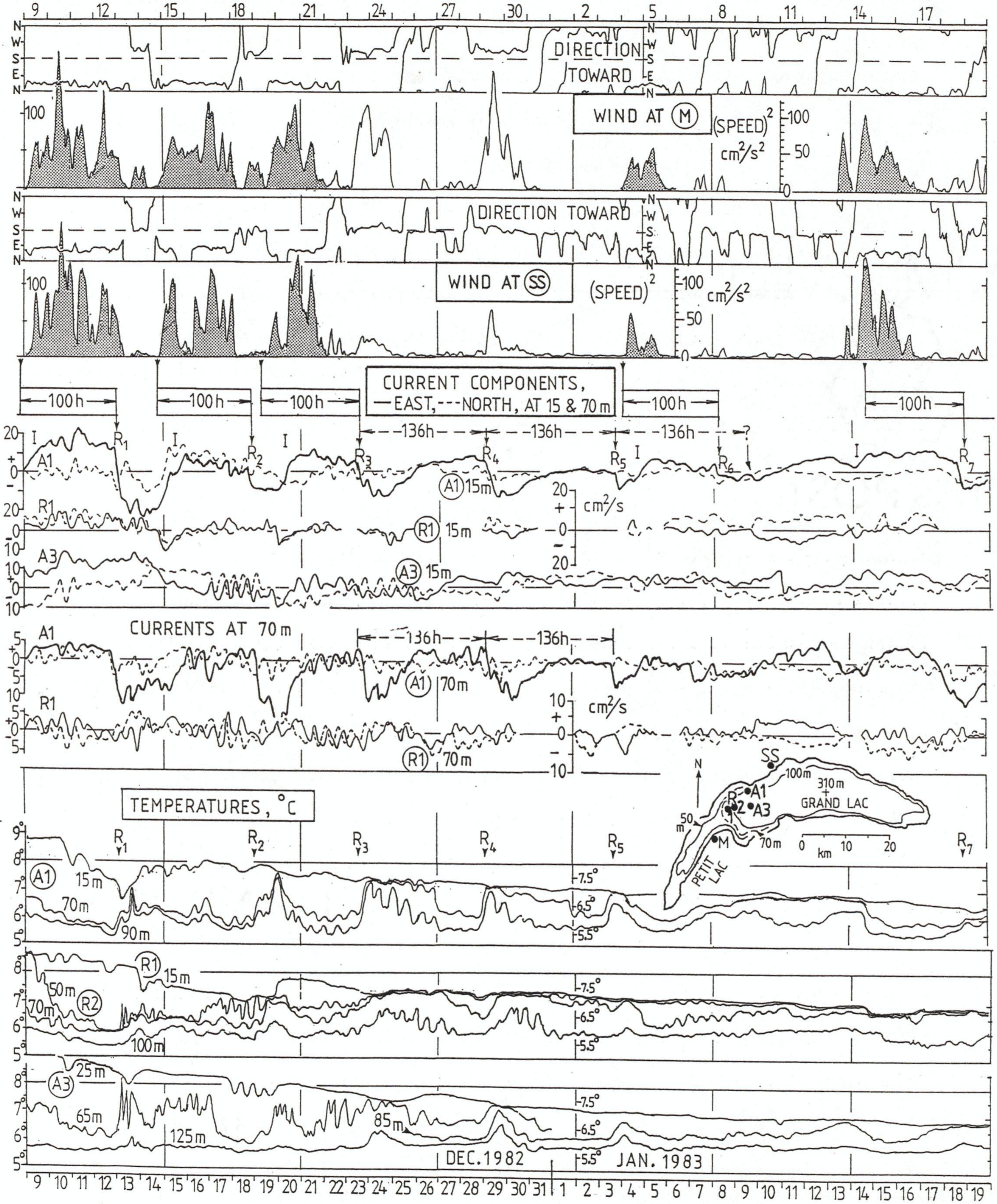


Fig. 24. Caption on opposite page.

strongest during the weak stratification of winter, when the internal wave celerity c_i fell to the level of ambient current speeds.

Nevertheless, in spite of the above-described complexity in wind forcing and response, my search through all the winter records disclosed one clear example of a basin-mode response (Fig. 24). During the interval Dec. 1982 to Jan. 1983, Léman was still stratified with a thermocline at about 100m depth. Current meters and thermistor chains were moored in groups A and R at the west end of the Grand Lac. Anemometers at Messery (M) and St. Sulpice (SS) revealed four distinct pulses of wind from SW to W (and one weak event) all shown shaded in Fig. 24. At roughly 100h after the start of each wind event, the currents at 15 and 70m at mooring A1 (note double speed scale at the latter depth) reversed suddenly from eastgoing to westgoing at the mooring nearest shore. Those reversals, R_1, R_2, R_3 and R_7 (with R_6 questionable), were each accompanied by a sudden depression of the thermocline (i.e. a sudden temperature rise) at mooring A1, marking the passage of a surge. Similar saw-toothed temperature "waves" were also seen at moorings further offshore at Rolle 1 (R1) and A3. The reversals in current direction, however, were confined to the nearshore instrument, A1, except for weak delayed signals at 15m, at mooring R1, discussed below.

With so localized an array of moorings, one can only speculate on what the whole basin response (if any) might have been. The evidence from 1950 in Figs. 19 and 20, however, suggested that W-wind-induced transport of surface water from the western half of the lake, including that transported from the Petit Lac, was deflected to the right by the earth's rotation (Ekman transport) forcing a depression of the thermocline along the W-wind-sheltered shore near Thonon and

Fig. 24. (See opposite page.) Current components (cm/s) and temperatures (^{o}C) recorded by the Laboratoire Hydraulique, EPFL, Lausanne, at 3 moorings at the W end of the Grand Lac (Léman) for 42 days starting 9 Dec. 1982, compared with wind speed and direction at Messery (M) and St. Sulpice (SS). Further details in the text.

perhaps further east. That depression then traveled, it was conjectured, as a steep-fronted shore-trapped internal Kelvin wave, counterclockwise around the basin's periphery.

If that conjecture is applied to Fig. 24, and if it is assumed that the internal surge travels at speed c_i (Kelvin wave speed in a two-layered model) around the particular isobath which corresponds to the equilibrium depth of the thermocline, then the perimeter track lengths and the circuit times can be estimated. This is done in Table 1. Because the equilibrium thermocline depth was at 20m in 1950 and at 90 or 100m (two cases considered) in 1982, the latter perimeter tracks were much shorter and were confined to the Grand Lac.

Table 1. Parameters of two-layered models (see footnote on p. 27) fitted to the Léman basin and to average temperature profiles estimated for the interval R3 to R5 in Fig. 24, (Dec. 1982, cases A, B) and from Fig. 19 (Sep. 1950, cases C,D). The bottom layer thickness h_2 is taken as 200 m in cases A and C and as 300 m in cases B and D. T_1 and T_2 are uniform temperatures assigned the upper and lower layers, respectively; a is the Rossby radius.

Case	h_1 m	T_1 °C.	h_2 m *	T_2 °C.	$(\rho_2-\rho_1)/\rho_2$	c_i cm/s	$a=c_i/f$ km	Interface perimeter km	Basin circuit time, hours	Time TH to A1, hours
										(76 km)
A	90	7.3	110	5.75	0.0000605	17.1	1.8	100**	162	123
	100	7.2	100	5.7	0.0000569	16.7		95**	158	126
B	90	7.3	210	5.7	0.0000619	19.6	1.9	100**	142	108
	100	7.3	200	5.65	0.0000582	19.5		95**	135	108
										(78 km)
C	20	17.5	180	6.0	0.001254	47	4.4	146	86	46
D	20	17.5	280	5.8	0.001260	48	4.4	146	84	45

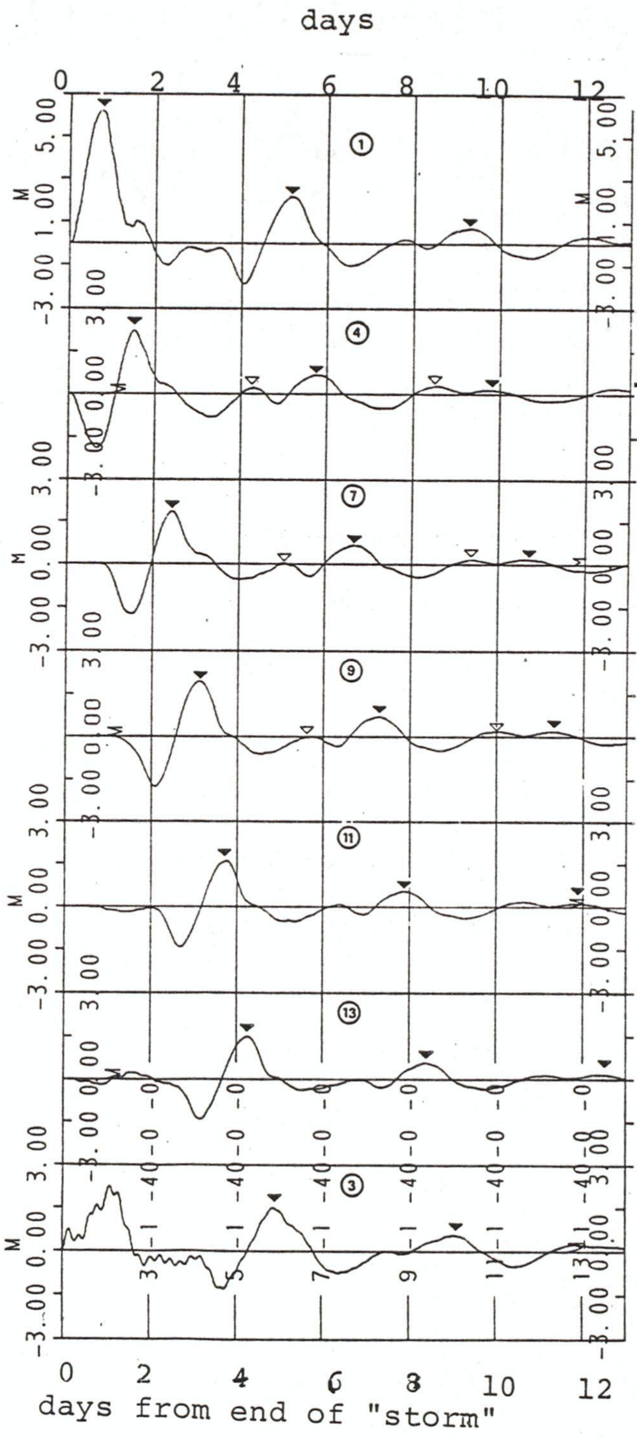
*The total depth of the model basin is taken as 200 m in cases A and C, and as 300 m in cases B and D.

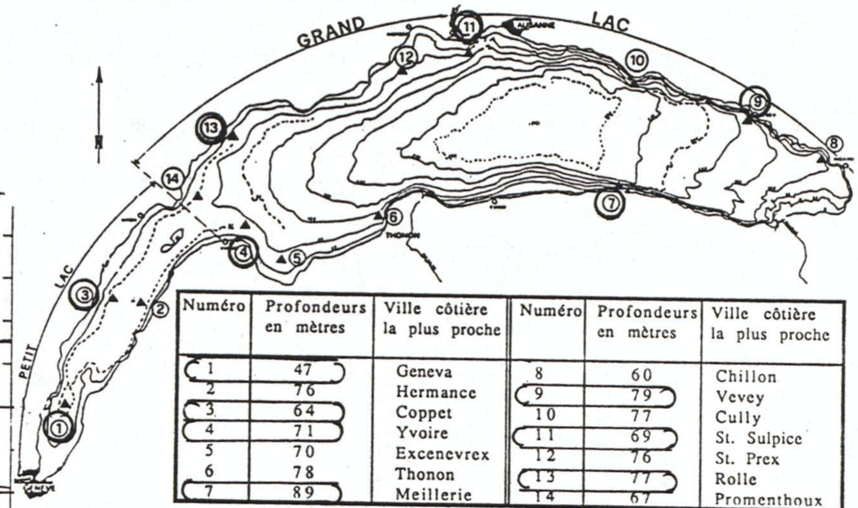
**Does not enter the Petit Lac, the entire bed of which lies above the 70 m depth contour.

Table 1 also lists the corresponding Rossby radii, a, as 4.4km in 1950 and 1.9km in 1982. Therefore the postulated 1982 wave was more closely trapped to the shore than was the 1950 wave.

The above interpretation suggests that the three current reversals R_1, R_2, and R_3 mark the arrival at mooring A1 of three separately generated, nonlinear internal surges which had traveled around the eastern and northern shores of the Grand Lac after their respective initiations on the southern shore by the three SW-W storms shown shaded at top-right in Fig. 24. The surges corresponding to R_1 and R_2, it is supposed, were unable to complete a full circuit of the basin before a new SW wind pulse intervened. But R_3 was able to do so at least twice, re-passing A1 at 136h intervals, as R_4 and R_5, during the long absence of strong SW-W wind (23 Dec. to 10 Jan.). (The NE wind pulses seen at M -- weak at SS -- during that interval were coincident with but not, I believe, causally connected to reversals R_3 and R_4.) As one would expect in a free oscillation persisting for three cycles, the amplitudes of the current reversals R_3 to R_5 and the amplitude of the associated saw-toothed temperature waves at moorings A1 and A3 decreased steadily with time.

The choice of maximum depth (200 or 300m) for the model basin in Table 1 makes little difference in the 1950 case. But for 1982, the calculated value of c_i is more sensitive to the choices of h_1, h_2, T_1, and T_2. The postulated shore-trapped nature of the wave/surge amplitude is demonstrated by the difference in temperature-wave amplitude at A1 and A3 (respectively 1 and 4km from shore) and, notably, by the confinement of the large current reversals to the nearshore mooring A1. The reversals at mooring R1 occurred about 1 d after reversals R_1, R_2, and R_3 at A1. This is evidence of the wave's cyclonic progress, but the speed of about 6km/d along the 100m contour there is less than half of c_i calculated in Table 1. While one should expect faster progress in the deeper eastern half of Grand Lac and slower progress in the western half, so large a discrepancy has yet to be explained. Decreases in speed of nonlinear surges, when compared with the celerity of the corresponding linear waves, were observed in Lake Zurich (Mortimer & Horn, 1982) and Kootenay Lake (Wiegand & Carmack, 1986).





Numéro	Profondeurs en mètres	Ville côtière la plus proche	Numéro	Profondeurs en mètres	Ville côtière la plus proche
1	47	Geneva	8	60	Chillon
2	76	Hermance	9	79	Vevey
3	64	Coppet	10	77	Cully
4	71	Yvoire	11	69	St. Sulpice
5	70	Excenevrex	12	76	St. Prex
6	78	Thonon	13	77	Rolle
7	89	Meillerie	14	67	Promenthoux

Fig. 25. Output from a two-layered mathematical model (Beckers, 1989, facsimiles) showing oscillations in "thermocline" level (m) at seven stations around the Léman basin (see circled numbers, with depths, in the map) during 12 days following an 18 h model wind pulse (from SW at 5 m/s) applied only to the Petit Lac. (The Grand Lac is sheltered by mountains.) Filled triangles follow the decaying internal surge peak on two counterclockwise, 4-day circuits around the basin, thus simulating the progression observed in Fig. 20.

In a recent thesis, incorporating linear and nonlinear, 2-dimensional, 2-layered models of Léman, Beckers (1989) simulated numerically the observations in Figs. 20 and 24. The former simulation, shown here in Fig. 25, follows a counterclockwise-propagating internal surge after application of a model SW wind pulse imposed for 18 hours on the Petit Lac only, thus taking account of the strong sheltering of the Grand Lac by mountains. The main features of Fig. 20 are reproduced.

The answer to the question of whether, in Léman and other large lakes, basin-mode responses develop depends on how those responses are defined. If, with Bohle-Carbonell and van Senden (1990), one regards a basin mode oscillation as one which persists for "more than five crossing periods" (2.5 cycles) then few internal basin modes were caught in their statistical net. However, the evidence in Figs. 19, 20 and 23 to 25 suggests to me that start-ups of basin modes are not uncommon in Léman, but that they are usually soon aborted or modified by a new wind stress. It also appears that the wave fronts are steepened into surges by nonlinear interactions and that the Fig. 22 model must be modified accordingly (see Bennett, 1973). Only rarely is a strong, extensive wind impulse followed by a calm interval long enough to permit internal oscillations to circumnavigate the basin for several cycles.

Whether or not the three separate wind-forced internal surge responses, followed by three cycles of free circum-basin progression of the third response (Fig. 24), are to be regarded as basin-mode responses is a matter of definition. They are clearly basin-steered, but completion of one or more cycles in unaborted form in Léman is rare. More common are the 2.2 cy/d oscillations in temperature and the 1.7 cy/d oscillations in current seen in Fig. 23(a), (b), and (c). These <u>may</u> be Poincaré waves and/or transverse internal seiches, i.e. basin-mode responses; but that question can only be settled by an appropriate basin-wide array of recording instruments and by repeated whole-basin transect surveys.

Whole-lake responses have been recorded in basins of dimensions similar to those of Léman. Examples are Lake Biwa, Japan (Fig. 26, Kanari 1976, 1984) and Vättern (Funkquist, 1979).

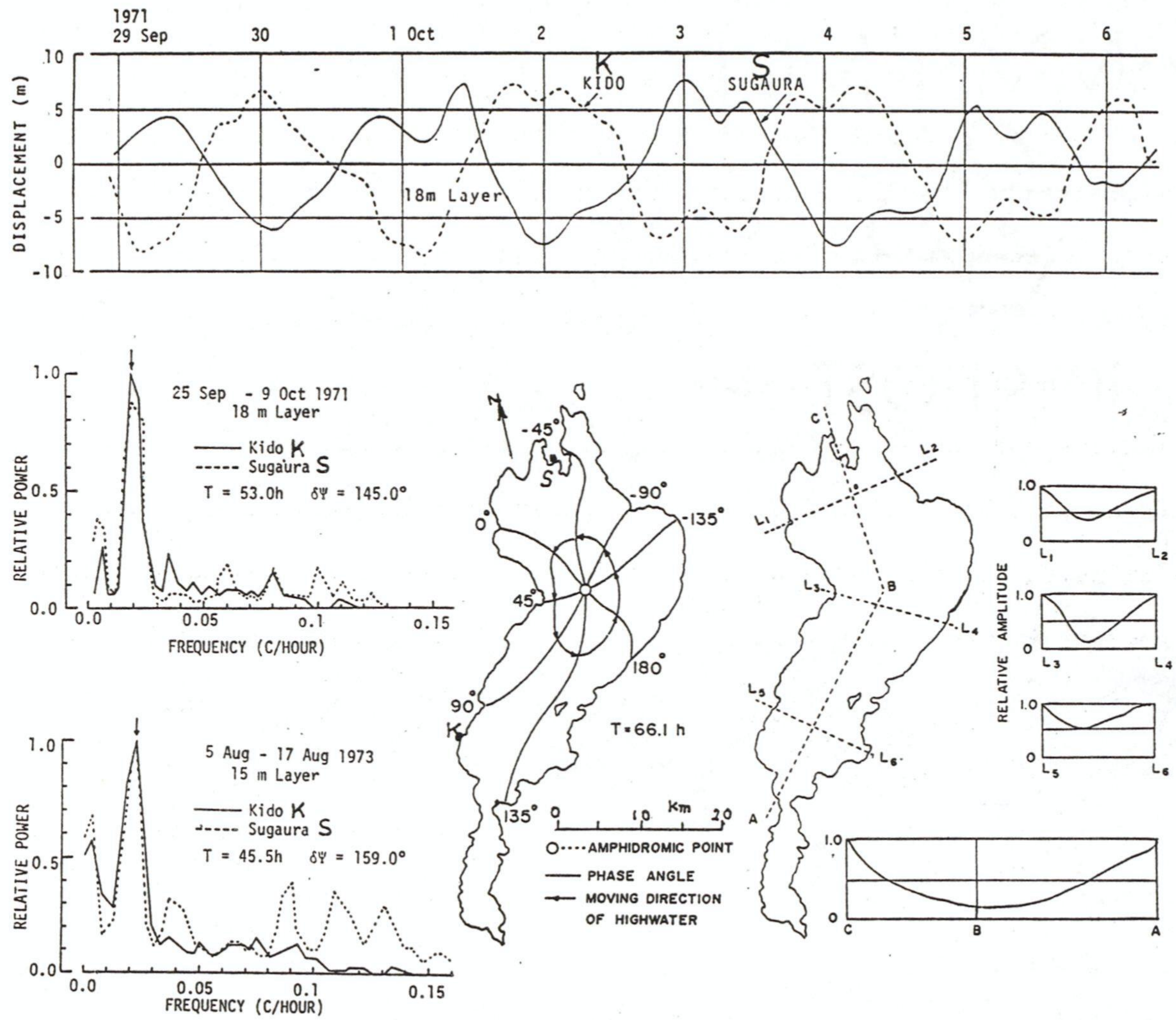


Fig. 26. Internal Kelvin wave in Lake Biwa, Japan, observed (above) with spectra (at left) and modeled (Kanari, 1984).

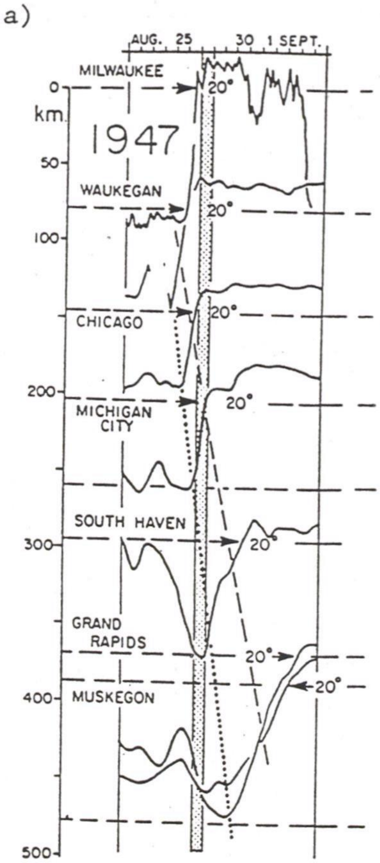
In narrower basins, Coriolis effects have also been demonstrated and interpreted as internal, nonlinear (i.e. steep-fronted) Kelvin waves (Loch Ness, Mortimer 1955; Kamloops Lake, Hamblin 1978; Mjøsa, Rye 1979; Mjørk et al. 1979, 1980). In the case of Mjøsa, basin topography determines that travel of the wind-induced wave/surge is predominantly unidirectional.

The influence of the earth's rotation in basins of width order 100 km (The Great Lakes)

(i) Nearshore responses of the thermocline to wind impulses

Léman had alerted me to the clues from internal motions, which temperature records from water intakes can provide. Therefore, during my first visit to North America in 1953, I began to collect such records from around Lakes Michigan and Ontario. And it turned out that, in those basins also, occasional evidence of a response with Kelvin wave features (cf. Fig. 18b) appeared (example at left in Fig. 27) when extensive wind-induced tilting (downwelling) of the thermocline was followed by some weeks of relative calm. Because Kelvin wave progress is slow (at speed c_i) it is not surprising that whole-basin circuits were never seen. In Lake Michigan, as in Léman, the Kelvin-type waves were steep-fronted, a feature explained by nonlinear theory (Bennett 1973). The situation is further complicated in Lake Ontario, where Simons (1980) demonstrated an interaction between a shore-trapped internal Kelvin wave and a topographic Rossby wave, not further treated in this review.

(a)



(b)

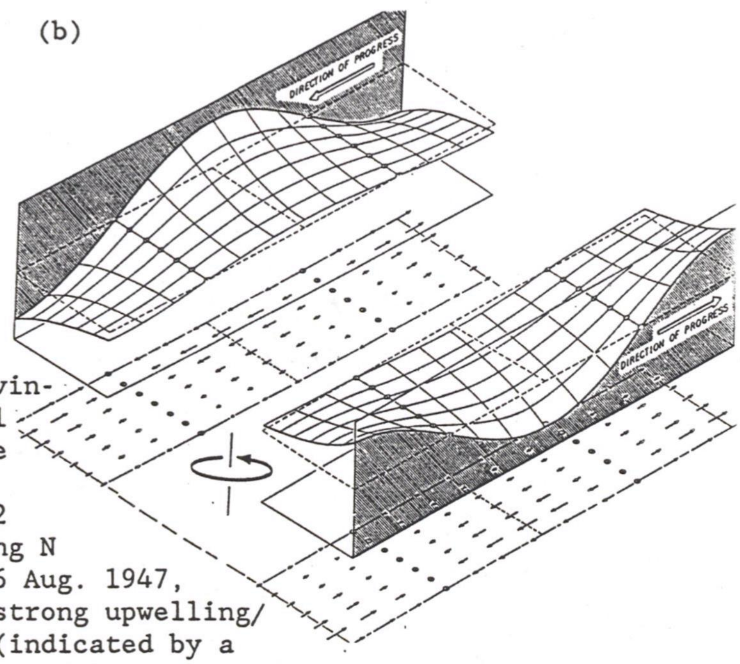


Fig. 27. Kelvin-type internal waves in Lake Michigan in response to 2 days of strong N wind on 25/26 Aug. 1947, followed by strong upwelling/downwelling (indicated by a shaded strip): (a) evidence from water intake temperature records; (b) a model Kelvin wave in a wide channel of uniform depth.

Entirely different periodic signals were occasionally seen in the temperature records, when Lake Michigan's thermocline lay for several days near intake depth. These were interpreted as Poincaré waves (see Fig. 18c) of frequency close to or slightly above f. The corresponding dominant periods in the episodes shown in Fig. 28, for example, were either indistinguishable from the local inertial period (17.5 h) or distinctly less (e.g. 15 h). The probable explanation of this, initially puzzling, behaviour will be presented later as part of geostrophic readjustment to wind-induced downwelling nearshore.

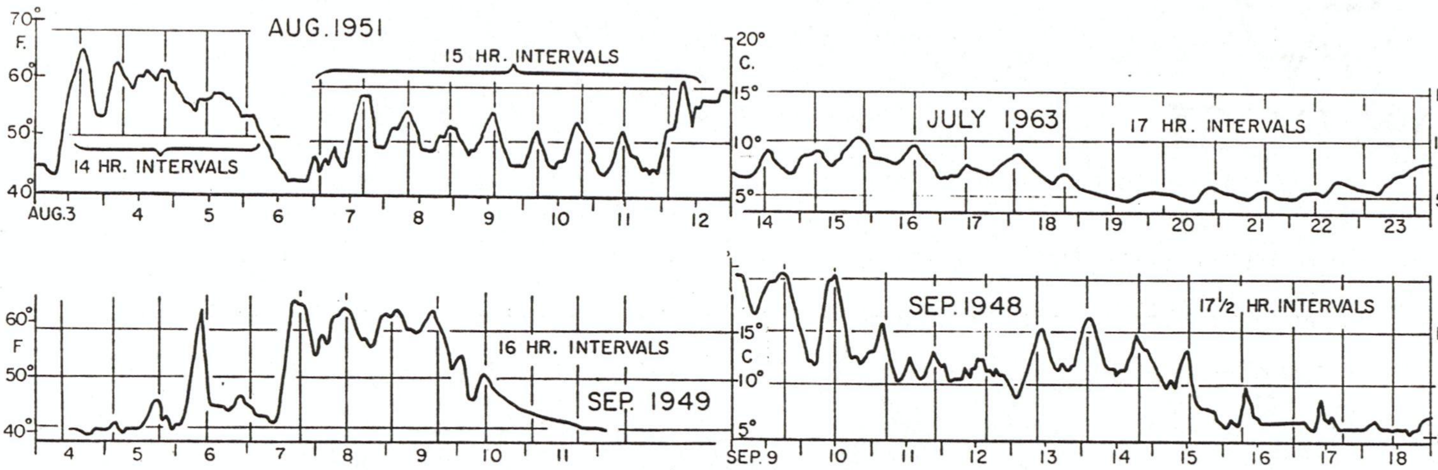


Fig. 28. Episodes of internal Poincaré-type waves in Lake Michigan as disclosed in temperature records from the Milwaukee water intake at 18 m depth, when the thermocline was near that level.

For help in interpreting the Lake Michigan findings, I turned (in a 1963 paper) to well-known theoretical models of tidal waves in two-layered, rotating wide channels, in which the "thermocline" displacement responses to wind impulses included: (a) shore-trapped low-frequency Kelvin waves of the type already described; (b) offshore inertial motion proper (called Sverdrup waves in 1963, here called prototype Poincaré waves, Fig. 18c); and (c) particular combinations of (b) which, if given time, can produce cross-basin standing waves (seiches). Stages in this process will be discussed later.

(ii) Offshore inertia-dominated responses of thermocline and currents to wind impulses

Publication of the 1963 hypothesis coincided with my tenure of a visiting professorship in Wisconsin and was specifically prompted by the plans of the (then) U.S. Federal Water Pollution Control Administration (FWPCA) to deploy an array of oceanographic current meters and thermographs in each of the Great Lakes. Lake Michigan's turn came in 1962/64; and I organized a parallel study of the thermal structure and internal oscillations, with the help of graduate students wielding bathythermographs on over 70 regular railroad ferry crossings from Milwaukee to Muskegon (see later Fig. 34). At the same time current and temperature profiles were obtained at mid-lake anchor stations occupied by a small research vessel (later Fig. 30). Some of the results of these joint studies have appeared in data reports (U.S. Dept. Interior, 1967, Mortimer 1968, 1971) and some appear in a later section, but much awaits analysis and publication. The same must also be said of the subsequent, large-scale campaign in 1972 on Lake Ontario, the International Field Year for the Great Lakes (IFYGL, Aubert and Richards 1981), some results from which are presented later. It is regrettably often easier to obtain funds to collect the ingredients for a meal than to properly serve and digest it. But discoveries during those two campaigns contributed notably to the development of physical limnology and have also influenced oceanographic theory, demonstrated, for example, by many subsequent papers in the Journal of Physical Oceanography.

Presentation of the 1963 findings, concerned with long internal waves in Lake Michigan, is here confined to a few samples and to later discussion of unanswered questions. Some of the figures may appear over-congested; but this is a deliberate attempt to display the principal features and connections, not the details. Placed center-stage in Fig. 29 is the wind record (2-hourly speed ranges) from 17 July to 8 August at FWPCA mooring 18 (see chart at top

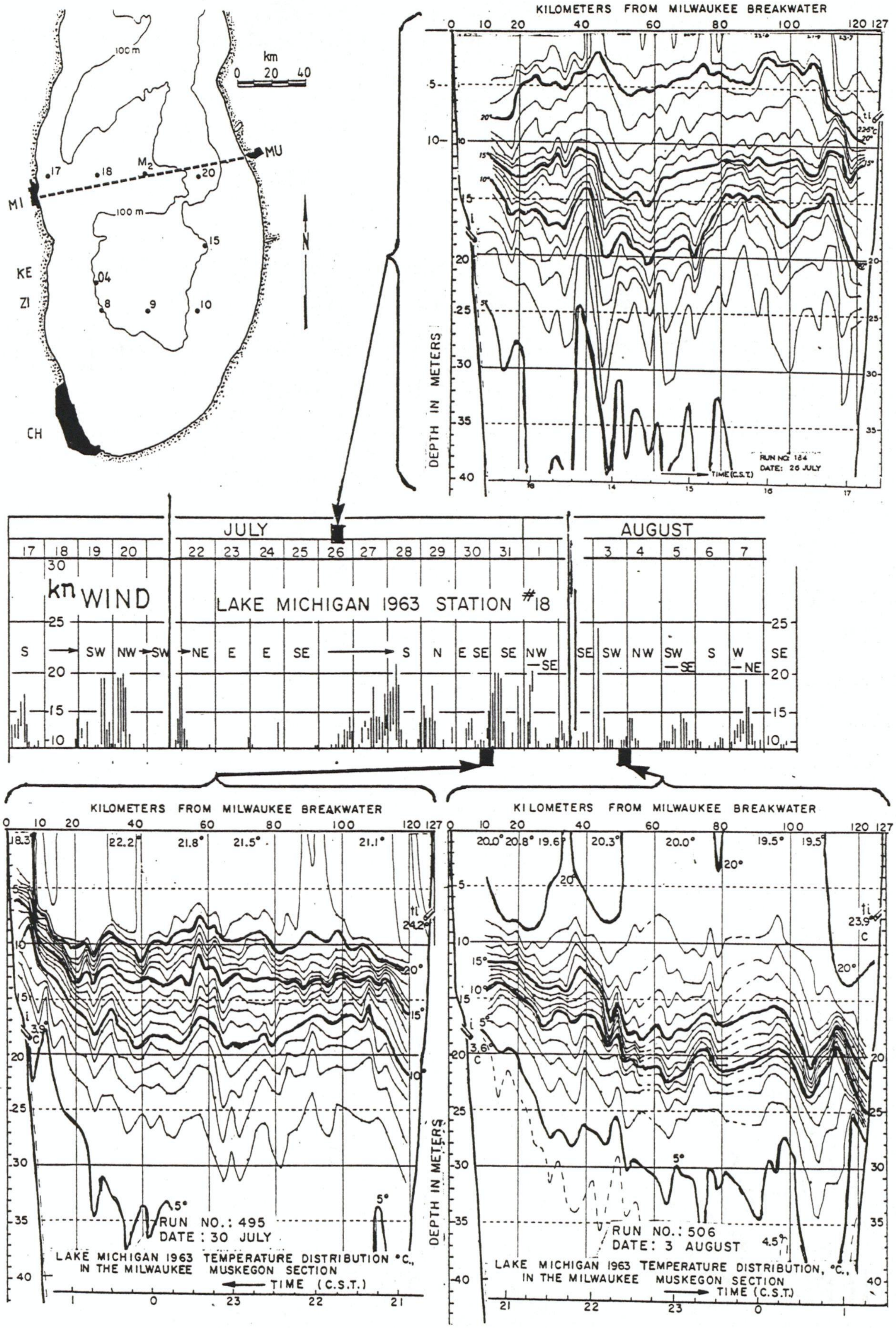


Fig. 29. Caption on opposite page.

left for selected FWPCA 1963 mooring positions, the Milwaukee to Muskegon ferry track, and the position of anchor station M2). Three representative plots of cross-lake temperature distribution, obtained from the railroad ferry, are displayed. The first, 26 July after 5 days of calm, reveals the influence of surface heating with little mixing in the upper layer. The second, 30 July after 4 days of weak variable wind, shows increased mixing in the upper layer, but little internal wave activity on the thermocline. The third, 3 August after a short but strong SE wind impulse the day before, shows a strongly tilted and wavy thermocline. These temperature cross-sections and those presented in later figures are not synoptic, because the ferry takes, on average, 5.5 h for the crossing. Thus, near-inertial motions are aliased. But some features of the cross-lake temperature distribution can be revealed in time/distance diagrams of the type illustrated in later Fig. 35. Already visible in Fig. 29 and more conspicuous in later Fig. 32 are the many abrupt temperature fronts encountered on the crossings. Some of these could be followed in their propagation from shore, a phenomenon to be introduced and discussed later.

The 30 July and 3 August sections in Fig. 29 may be compared with the corresponding 30/31 July and 3-5 August plot of isotherm depth oscillations at anchor station M2 (Fig. 30). The influence of the 2 Aug. wind impulse is visible at M2 as an increase of internal wave activity, dominated by a repeating 17 h oscillation with nonlinear features, including the presence of a substantial double-frequency component.

Corresponding current profiles at the anchor station, measured at approximately 2 h intervals and illustrated in Fig. 30, show a slab-like,

Fig. 29. (on opposite page). Three transects of Lake Michigan (by railroad ferry, Milwaukee-Muskegon, see map, top left) on 26 and 30 July and 3 Aug. 1963 on which the temperature/depth distribution is plotted from repeated bathythermograph casts (Mortimer 1968, 1971). Wind speed (2-hourly speed ranges at mooring 18, see map) is plotted on a $(\text{knots})^2$ scale to indicate relative wind stress. The equipment is illustrated in later Fig. 34.

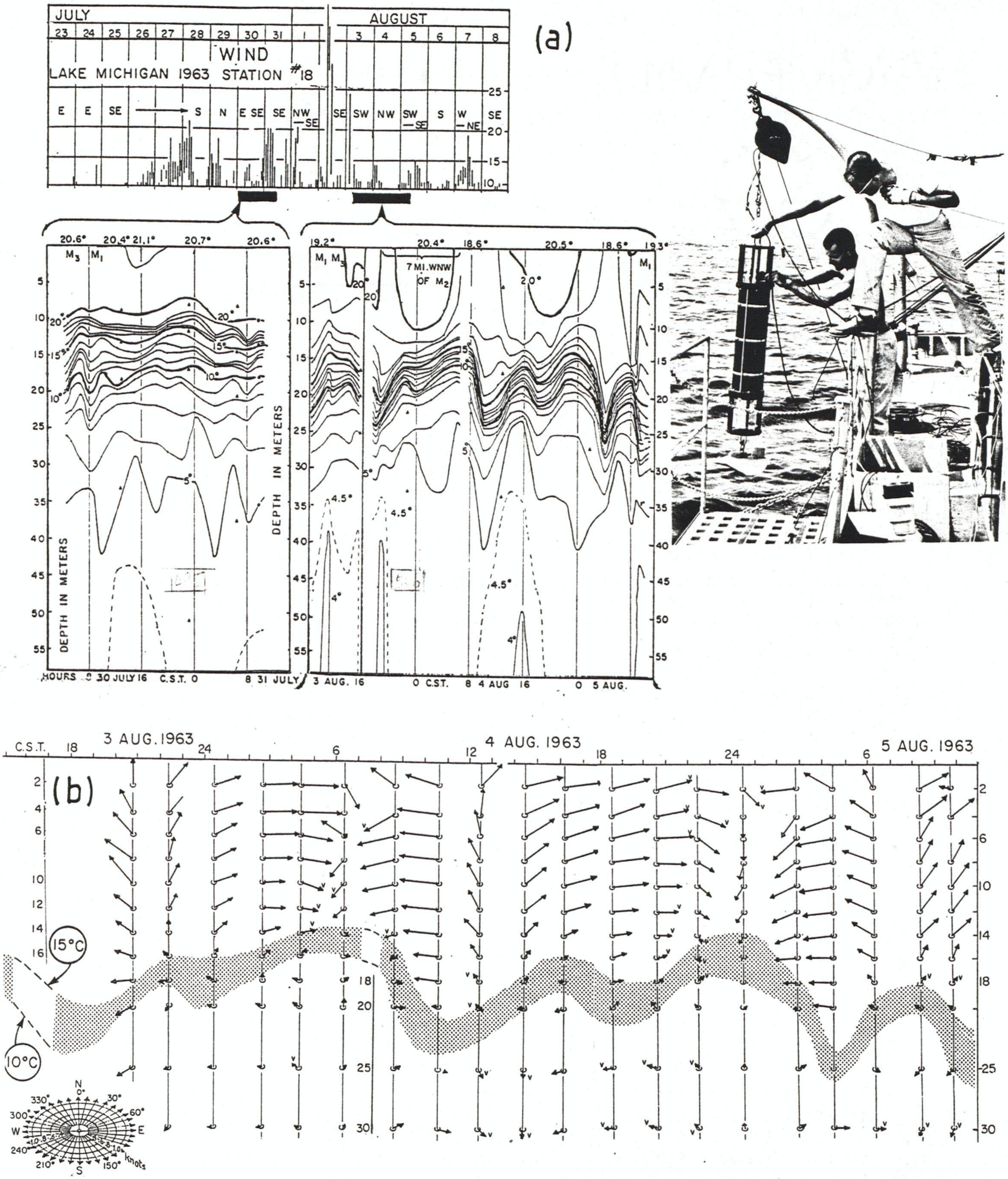


Fig. 30. Caption on opposite page.

waltzing motion of the upper layer (compare similar behavior in the Baltic Sea, Krauss, 1981). Between the thermocline and the surface, the current is fairly uniform and its direction rotates clockwise at a frequency close to f (i.e. with a periodicity close to 17 h). Below the thermocline a much smaller current also rotates clockwise, but nearly 180° out of phase with the current above. Looking down and proceeding downward through the thermocline, the current vector shifts clockwise until the 180° direction reversal is achieved. This is consistent with downward transmission of energy. The thermocline is, therefore, a layer in which shears are generated. If these are large enough, instability and mixing ensue (examples from the Baltic Sea are given in Krauss 1981).

The responses, illustrated in Figs. 29 and 30, to the 2 Aug. wind impulse are the first record of inertial motion in a lake basin. During that summer similar responses were recorded at most FWPCA moorings, but not at those within a few km of the shore (examples in Verber 1964 and Mortimer 1971). To demonstrate the almost continuous but intermittent excitation of inertial wave responses to wind changes in Lake Michigan, it is instructive here to take, not a 1963 record, but a later set (Mortimer 1980). In three panels, Fig. 31 displays continuous records of current and temperature at 17 m, and temperatures at three lower depths, at a mooring 30 km from shore, from 3 Aug. to 4 Oct. 1976, accompanied by the wind record from a nearby onshore anemometer. At the beginning of the series the thermocline lay near 25 m; at the end it was about 10 m deeper. Currents are displayed in two

Fig. 30. (see opposite page). Lake Michigan, measurements at anchor station, mid-lake between Milwaukee and Muskegon: (a) isotherm depths, 30/31 July and 3 to 5 Aug. 1963; and (b) 2-hourly (approx.) profiles of current vectors (displayed on isometric projection) 3 to 5 Aug. (Mortimer 1968, 1971). Also shown (top) is wind speed $(knots)^2$ and K. M. Stewart and D. C. McNaught preparing to lower the current meter.

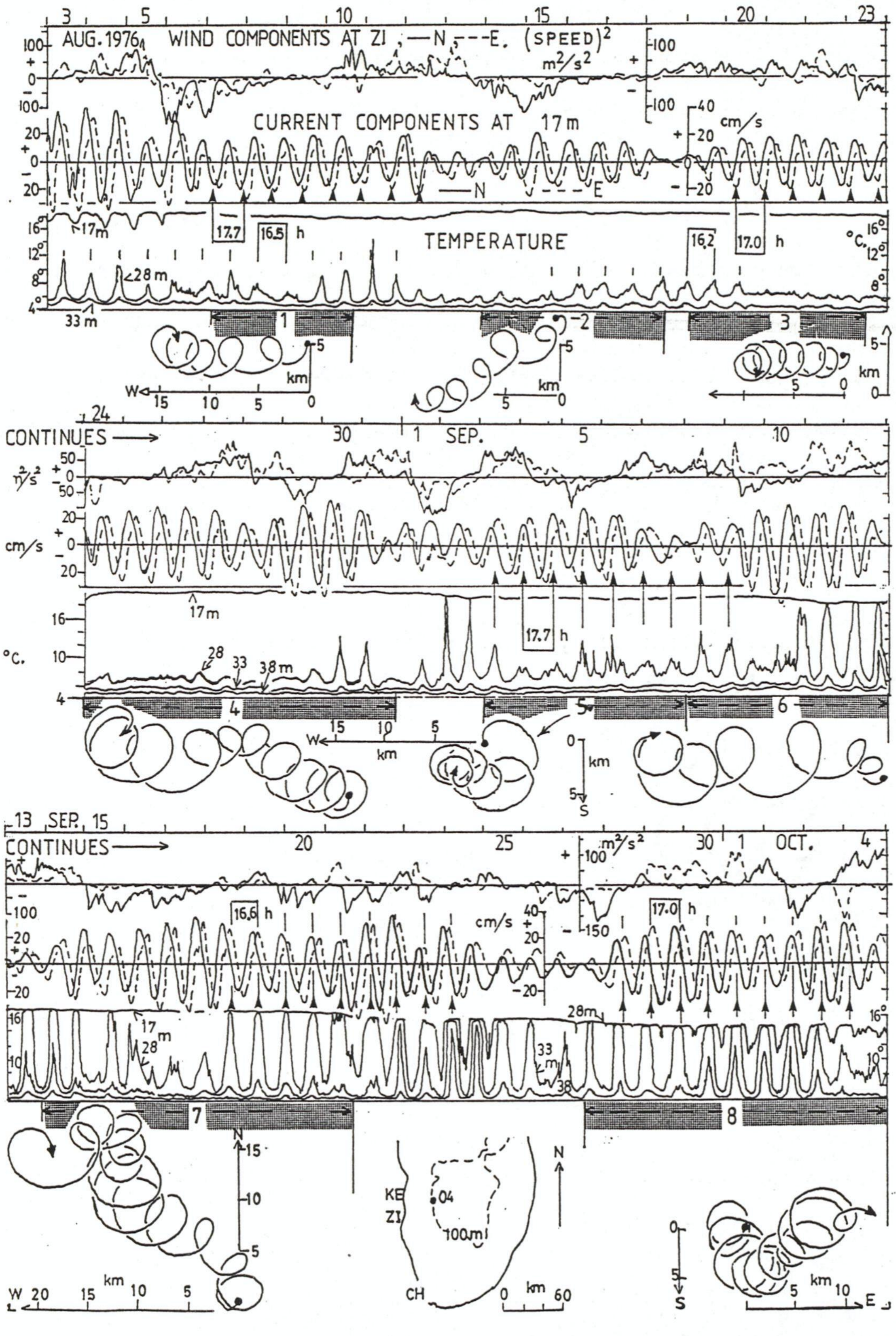


Fig. 31. Caption on opposite page.

ways: as N and E components; and as progressive (head-to-tail) vector diagrams for selected episodes (numbered and shaded).

The most obvious feature of the current pattern in Fig. 31, also generally seen during the 1963 campaign, was a regular clockwise rotation at or slightly above frequency *f*, sometimes combined with steadier drifts. The current oscillations tended to arrive in groups, rarely lasting for more than eight cycles (142 h); and there was usually a change of phase between the groups, often coinciding with a change in wind.

In Fig. 31, and in the 1963 records generally, the frequencies of the current oscillations were close to *f*, whereas the frequencies of the accompanying temperature oscillations were a few percent higher (e.g. episodes 1 and 3 in Fig. 31). Similar frequency differences between spectra of current and temperature oscillations were later also seen in the Baltic and Mediterranean Seas (Krauss 1981, Millot and Crépon 1981). But occasionally, temperature "wave" frequency is indistinguishable from *f*. An example was illustrated in Fig. 28. The explanation may lie in the Poincaré wave frequency changes which occur during the geostrophic readjustment process, to be described later.

The responses to wind change in Fig. 31 and in the 1963 records were rapid and their strength apparently depended, not only on the <u>force</u>, but also on the <u>duration</u> of the wind impulse. A similar dependence was also demonstrated by Pollard (1970) and modeled by Pollard and Millard (1970) for

Fig. 31. (see opposite page). Current components and temperatures at mooring 04 on the 100 m depth contour in Lake Michigan, 30 km E of Kenosha, Wis.; 13 to 23 Aug. 1976 (Mortimer, 1980). Displayed in three panels from top down are: the NS and EW (dashed line) components of wind speed squared (m^2/s^2) onshore at Zion, Ill.; current (cm/s) at 17 m depth; temperature at 17, 28, 33 and 38 m; progressive current vectors (looping tracks) during the numbered, shaded episodes.

oceanic inertial oscillations. The channel model of Krauss (1979), constructed to predict Ekman flow and inertial responses in the Baltic, also helps to disentangle the responses to varying wind stress in Lake Michigan.

The dominance of inertial wave responses in Lake Michigan during stratification (but not in the fully mixed water columns of winter) was confirmed by spectra of currents (Malone 1968) which displayed a large peak centered on f (Fig. 32, top left). The width of that peak (as suggested by Munk and Phillips 1968) may be a consequence of the above-noted low persistence of the inertial oscillations. But there was also a small peak near $2f$, which Malone (I believe mistakenly) attributed to the uninodal surface seiche. As peaks near $2f$ were also demonstrated later in spectra of inertia-dominated currents in Lake Ontario (Marmorino and Mortimer 1977) and in the Baltic Sea (Krauss 1981), a more probable attribution is to nonlinearity in the inertial wave, consistent with the waveforms observed in Fig. 30 and later Fig. 40. With appropriate choices of amplitude and phase, the waveform in those figures can be reproduced by a combination of two sinusoids of frequencies f and $2f$.

Nonlinearity in the inertial response may, therefore, be common. This is suggested by current spectra from mid-Lake Ontario where intermittent inertial responses to wind changes are common (later Fig. 45). In mid-Lake Erie also, Boyce and Chiocchio (1987) encountered "pure inertial motions forced by the wind in the surface layer and by the opposing surface pressure gradient or setup in the subsurface layers. Stratification controls the vertical distribution of turbulent stress in the water column." The authors found that the largest inertial response occurred in the mid-water column above the seasonal thermocline. That occurrence in Lake Erie was related to

> "the details of the thermal stratification. The hypothesis that the principal response of the water column to a changing wind field is the direct result of surface stress and pressure gradient is tested with simple diagnostic models that explain the subsurface maximum of the inertial-period motion in terms of a water column alternating between three and two moving layers."

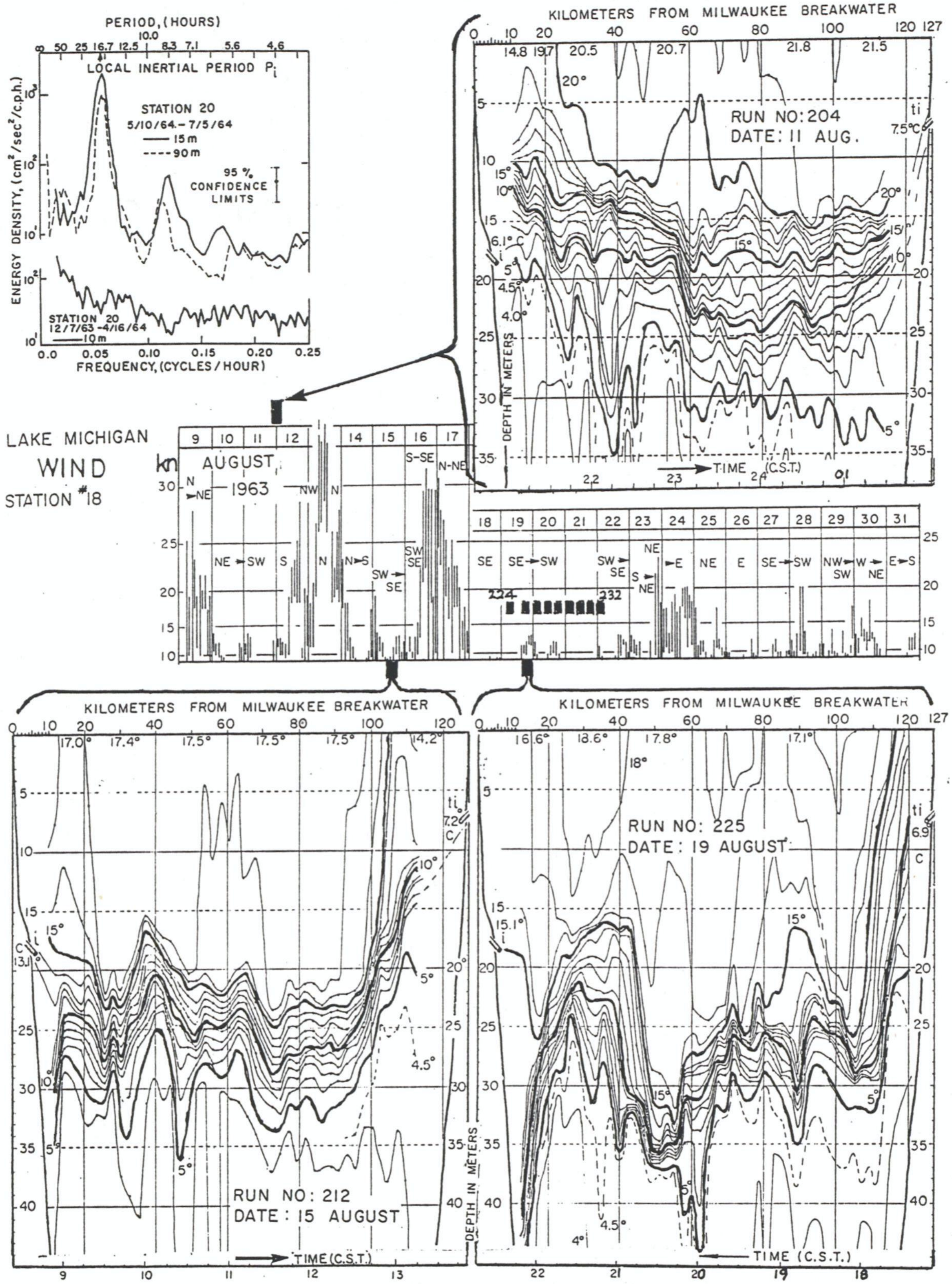


Fig. 32. Top left, frequency spectra of currents at two depths in summer and one depth in winter (Malone, 1968). The remainder of the figure presents temperature transects before (top right) and after (bottom left) the storm of 12/13 Aug. 1963 and (bottom right) after the storm of 16/17 Aug. Filled rectangles in the central diagram of wind speed (knots, square law scale) at Milwaukee indicate the time of ferry runs 224 to 232 (Mortimer 1968, 1971), re-numbered in Figs. 52 to 55.

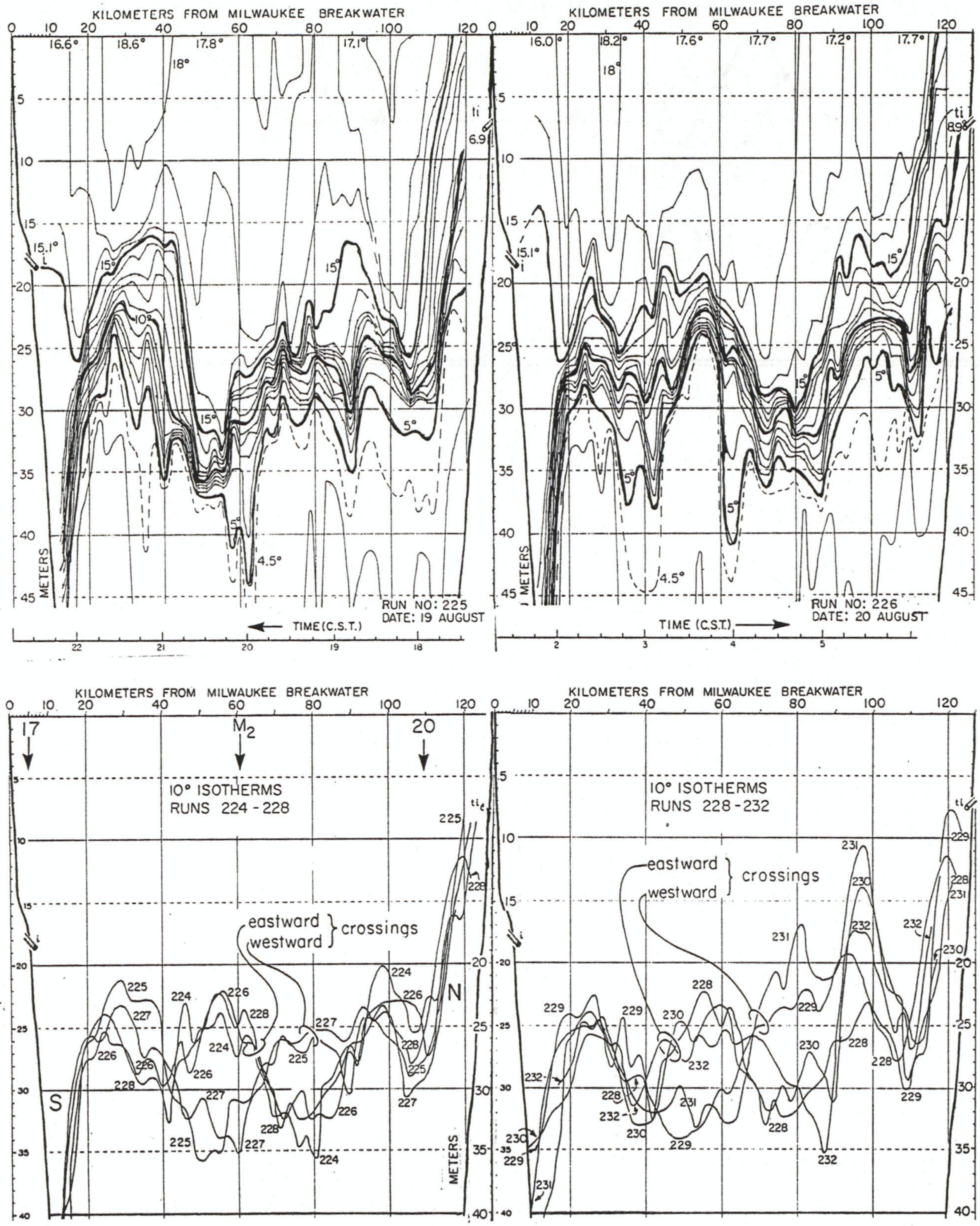


Fig. 33. Temperature (^{o}C) distributions on consecutive cross-lake Michigan runs, 19/22 Aug. 1963, after a major storm on 16/17 Aug. (see Fig. 32): upper portion, isotherm distributions from 0 to 45 m depth, mid-lake crossings westward (top left, 1945 h, 19th) and eastward (top right, 0400 h, 10th); lower portion, $10^{o}C$ (mid-thermocline) isotherms only on nine consecutive crossings (224 to 232, 1000 h 19 Aug. to 0500 22 Aug., re-numbered in Table 2 and Figs. 52 to 55).

(iii) <u>Cross-basin thermal structure after strong wind stress; an initial interpretation</u>

While inertial motions were shown to be the common, everyday responses of the open waters of a stratified Lake Michigan to variable wind, there were occasions in 1963 when strong and sustained storms strongly distorted the cross-basin structure of the thermocline. That structure, revealed by repeated transects of temperature distribution, was explored with bathythermographs operated at roughly 6 min intervals from a railroad ferry on its to-and-fro passages, Milwaukee-Muskegon. Church pioneered this technique in a 1942 study of the heat cycle of the lake. Twenty-one years later, in 1963, we recovered and employed his winch on the ferry which he used (Fig. 34).

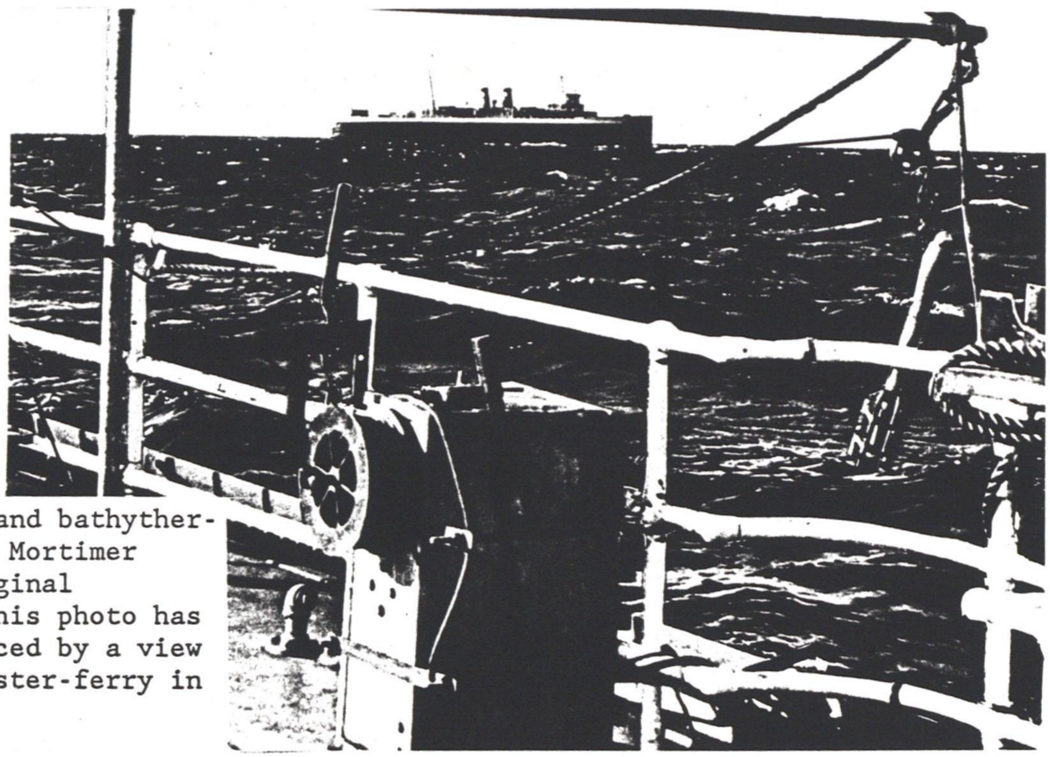


Fig. 34. Winch and bathythermograph used by Mortimer (1968). The original background of this photo has here been replaced by a view of a passing sister-ferry in mid-Lake.

Three storms on 9, 12/13 and 16/17 August progressively perturbed the thermocline isotherms (Figs. 32, 33 and more detail in later Figs. 54 and 55). The principal distortion was strong upwelling along one shore, accompanied by downwelling on the other, confined to a nearshore region of width less than three Rossby radii. Figs. 32 and 33 (which are continuations from Fig. 29)

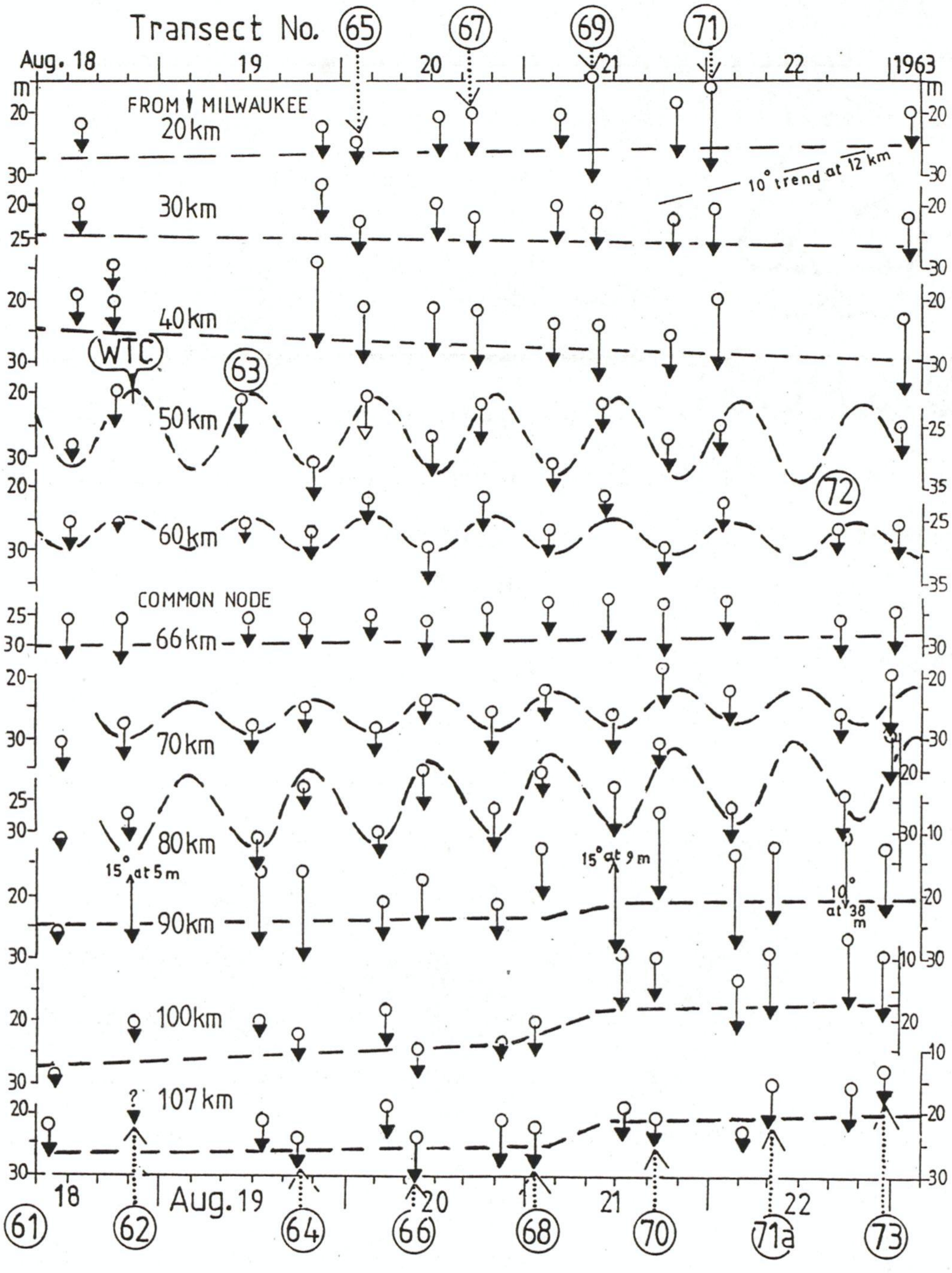


Fig. 35. Depths (m) of the thermocline-embracing isotherms 15° (circles) and 10°C (triangles) at selected distances (km) along the Milwaukee-Muskegon ferry track, interpolated in 14 temperature transects (re-numbered as in Fig. 55) traversed between 1800 h on 18 Aug. and 2400 h, 22 Aug. 1963, during calm after the 16/17 Aug. storm. This figure retrieves the actual thermocline motion (at the selected distances from MI) from the ferry-scanned thermocline topography illustrated in Fig. 33 and later 55. Revealed is a mid-Lake standing wave pattern of near 16 h period with a phase change at a node near 66 km from MI. To the west of the node, the thermocline reaches its crest (at WTC) at 2100 hours on 18 Aug. and at 16 h intervals thereafter.

illustrate the buildup of that distortion, with an overall deepening of the of the upper layer and the generation of large interfacial waves and fronts. This was the combined result of the two strong 2-day storms on 12/13 and 16/17 August, each of which ended with wind blowing strongly from the north. I postulated that the thus-created cross-lake asymmetry of the thermocline, with deflection concentrated near shore, favored generation of odd-numbered cross-lake internal seiche modes with higher modes strongly represented. This is in contrast to the usual domination in small lakes. Csanady (1973) formalized this concept in a model (see also Lee and Mysak 1979).

During the six days of calm following the 16/17 August storm, eleven full and two partial cross-lake temperature transects were obtained, the first during the evening of 18 August. Two of those consecutive transects and the superimposed traces of 10^{o} isotherms from each of ten runs are illustrated in Fig. 33. Consecutive crossings of the ferry often happened to pass the mid-lake point at intervals of about 8 h. Because odd-numbered Poincaré seiche modes would be expected to possess a common node near mid-basin and to occupy the period range 14 to 17 h, the mid-basin isotherm slopes would therefore be expected to change sign from one crossing to the next. That expectation appears confirmed in the lower half of Fig. 33. The timings of the ferry crossings were fortunate in that they brought the apparent seiche structure into focus near the center of the basin, but were less suited for that purpose near the sides. Nevertheless, an attempt (Fig. 35) to counteract the aliasing effect of the ferry timetable and to fit 16.5 h sinusoids to the observed fluctuation in 10^{o}C isotherm level at fixed distances from Milwaukee, disclosed a well-defined node near 65 km, less well-defined nodal regions near 40 and 90 km, but no clear picture in the regions of persistent steep isotherm slopes nearshore. Without adequate current data*, I viewed the observations as a set of transverse seiches (Mortimer 1971). A less likely interpretation of this particular episode was put forward by Lee and Mysak (1979), who suggested possible resonance of the 7th transverse mode with semi-diurnal wind forcing. Examination of the wind record does not support that.

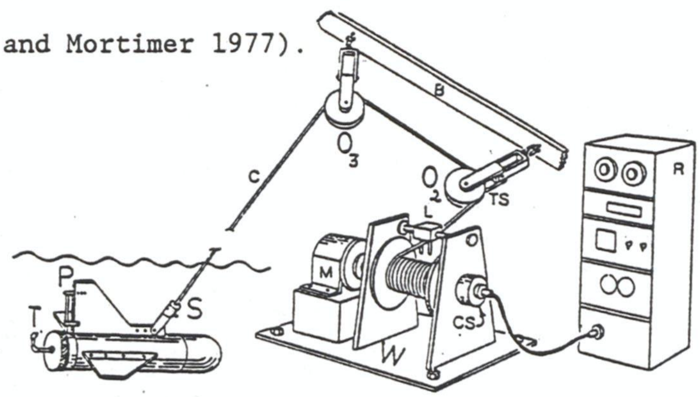
The apparent rarity of the transverse seiche response in Lake Michigan, in contrast to the every-day inertial responses, stems from the rarity of a sustained extensive thermocline deflection followed by a calm interval long enough to permit cross-basin transfer and reflection of Poincaré wave energy and the establishment of the standing wave pattern. Preliminary analysis of the IFYGL records from Lake Ontario (Mortimer 1977, 1980) demonstrated that inertial motion was also the most common response in that Lake and that, while transverse seiching does occasionally take place there, it is not common. This will be demonstrated by selected examples in a later section. Transverse internal seiches have not yet been seen in the only other large-lake basin for which comparably comprehensive records are available (Lake Erie, Boyce and Chiocchio 1987).

Recently, the transverse seiche interpretation of the 1963 observations in Lake Michigan has been called into question (Fennel, 1989). I therefore return to examination of this matter in the final discussion section -- unexplained phenomena and unanswered questions -- starting on p. 75.

*It was unfortunate that only two moored current meters operated within 15 km of the ferry transect.

(iv) <u>Transversely propagating surges, observed and modeled</u>

In narrow-lake examples cited earlier (Earn, Ness, and Zürich) internal surges propagate away from the basin ends, when wind-induced downwelling or the thermocline downswings of the ensuing longitudinal seiche are large enough there. In wider rotation-influenced basins, transverse motions arise; and one may therefore ask whether transversely-propagating surges are generated, when large downswings of the thermocline occur along the shore lying to the right of the wind direction. Repeated observations along cross-basin transects of Lake Ontario during the International Field Year for the Great Lakes (1972), with the towed depth-undulating sensors illustrated in Fig. 36, answered that question affirmatively (Mortimer and Cutchin, 1974; Mortimer 1977, 1980; Boyce and Mortimer 1977).



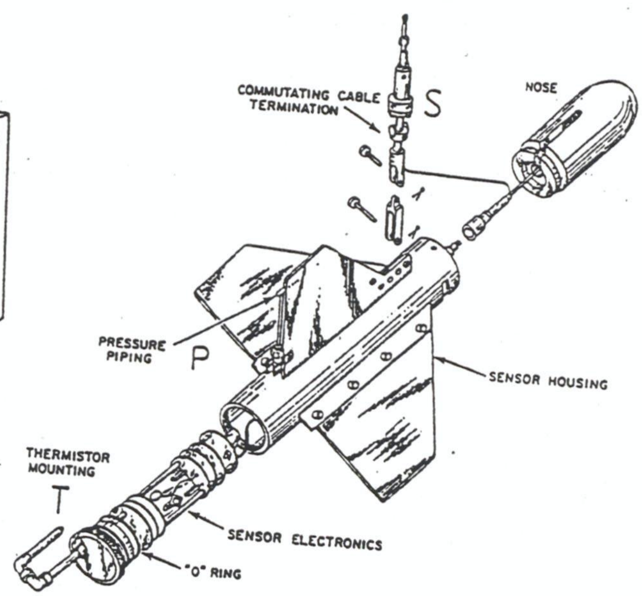
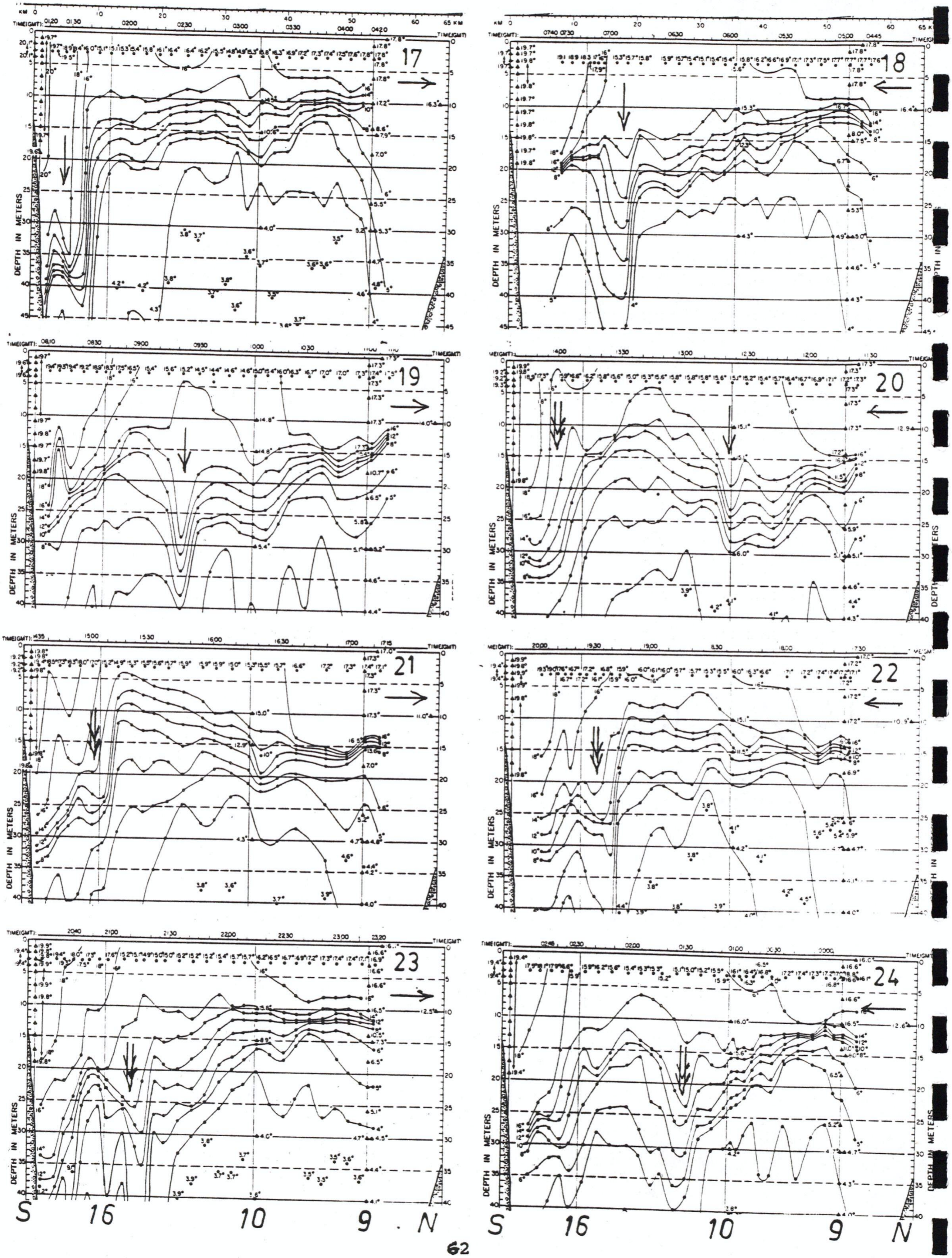


Fig. 36. Towed depth-undulating temperature/pressure (depth) sensors for repeated monitoring of cross-Lake temperature distribution during IFYGL (details in Boyce and Mortimer, 1977).

The orientation of the long axis of Lake Ontario is W to E; and that is also the prevailing wind direction. Therefore, currents generated by strong eastgoing wind are deflected southward by the Coriolis force and bring about frequent strong downwellings along the southern shore.

One such episode, illustrated in Fig. 37, was associated with a storm moving eastward along the basin during late afternoon on 8 August (see later Fig. 40 and Simons 1975 for the effect on the Lake as a whole). The storm


17
18
19
20
21
22
23
24
KM
65 KM
TIME(GMT)
DEPTH IN METERS
S
16
10
9
N

produced strong downwelling along the southern shore, first seen in transect 17, at top left in Fig. 37. An internal depression surge then started to move (northward, to the right in the figure) away from the shore (vertical arrows in transect sequence 17 to 19) at a speed ($\approx$ 2km/h) slightly greater than c_i. Concurrently the nearshore thermocline executed an up-and-down oscillation in a near-inertial period (15 h). The next downswing was large enough to generate a second northgoing surge, marked by double-headed arrows in transects 20 to 24. No surge was observed to radiate from or to be reflected from the northern shore, perhaps because downwelling there was too weak.

In a two-layered model fitted to Lake Ontario conditions, Simons (1978) obtained analytical and numerical solutions for the response to a wind impulse of finite duration and showed that the fronts with nonlinear features could be generated "without invoking excessive wind speeds". His conclusion was that the appearance of the second surge in Fig. 37 was to be viewed

> "as part of the oscillatory rather than the quasi-geostrophic response of the lake to wind. Although disturbances from the opposite upwelling shore may eventually combine with those from the downwelling shore to create standing Poincaré waves (Mortimer 1977), the scale of the frontal zone is sufficiently small that it can be treated independently of this effect. Thus, similar phenomena can be expected to occur in any near-shore region for suitable stratification conditions."

Simon's model, illustrated in Fig. 38 (b and c), extended Cahn's (1945) analysis of the geostrophic adjustment process to a two-layered system. The model was started just before the onset of the storm. Records from the anemometer at mooring 9 provided the wind stress input. The model satisfactorily explained the observed phase relationships between thermocline

Fig. 37. (see opposite page). Temperature ($^{\circ}$C) in the upper 40 m in six transects of Lake Ontario (Braddock Pt. to Presqu'ile) 9 Aug. 1972 (from Boyce and Mortimer, 1977; further details in the text). Three vertical lines indicate, l. to r. respectively, the positions of moorings 16, 10, and 9, from which temperature, current, and wind records are shown in Fig. 40.

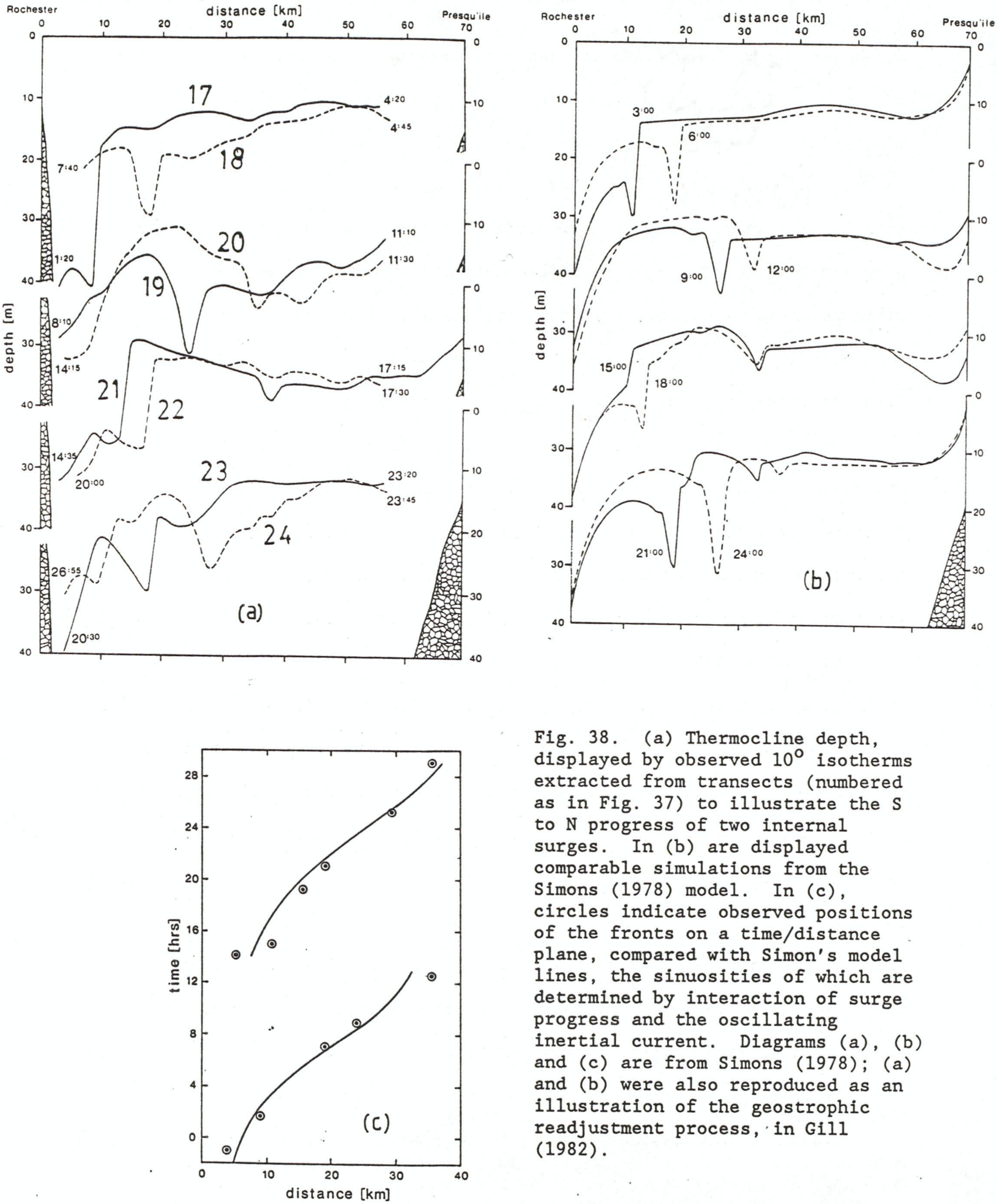


Fig. 38. (a) Thermocline depth, displayed by observed 10^{o} isotherms extracted from transects (numbered as in Fig. 37) to illustrate the S to N progress of two internal surges. In (b) are displayed comparable simulations from the Simons (1978) model. In (c), circles indicate observed positions of the fronts on a time/distance plane, compared with Simon's model lines, the sinuosities of which are determined by interaction of surge progress and the oscillating inertial current. Diagrams (a), (b) and (c) are from Simons (1978); (a) and (b) were also reproduced as an illustration of the geostrophic readjustment process, in Gill (1982).

displacements and the associated rotating currents. It also demonstrated that the degree of nonlinearity (steepness of the surges) depends on the strength and duration of the wind impulse. This explains why, as we shall see, some storms generate surges, while others do not. An impulse lasting for half an inertial period is the most efficient generator. As the surge moves offshore, it advances into regions previously subjected only to inertial waltzing. The interaction between that motion and the surge explains oscillation in the speed of surge advance, illustrated in Fig. 38(c).

(v) Surge generation as a process of geostrophic readjustment

Whenever a rotating fluid system is forced out of equilibrium, a geostrophic readjustment process ensues. This involves the dispersal of Poincaré waves from the perturbed region. Examples of such perturbations are the strong nearshore downwellings illustrated in Figures 32 and 37. Poincaré waves are vehicles for the radiation of energy away from the perturbed region at a rate which depends, not on wave speed, but on wave group velocity. For Poincaré waves that velocity is wavelength-dependent. At wavelengths short compared to the Rossby radius a, the group velocity is maximal at c_s for surface and c_i for internal waves, in the absence of rotation. As wavelength lengthens, the group velocity decreases, and approaches the lower limit (zero). As that limit is approached, the wave frequency shifts closer to f; and the orbital plane of particle motion (see Fig. 18c) tilts closer to horizontal. At the limit, the wave is indistinguishable from inertial waltzing motion, described earlier. I postulate that, at a point offshore, the dominant frequency (initially inertial) rises as the surge passes through, and then falls again to f when the longest Poincaré waves arrive.

As Gill (1982) explained with the help of Fig. 39, any initial transient deviation (of the water surface or thermocline) from equilibrium can be

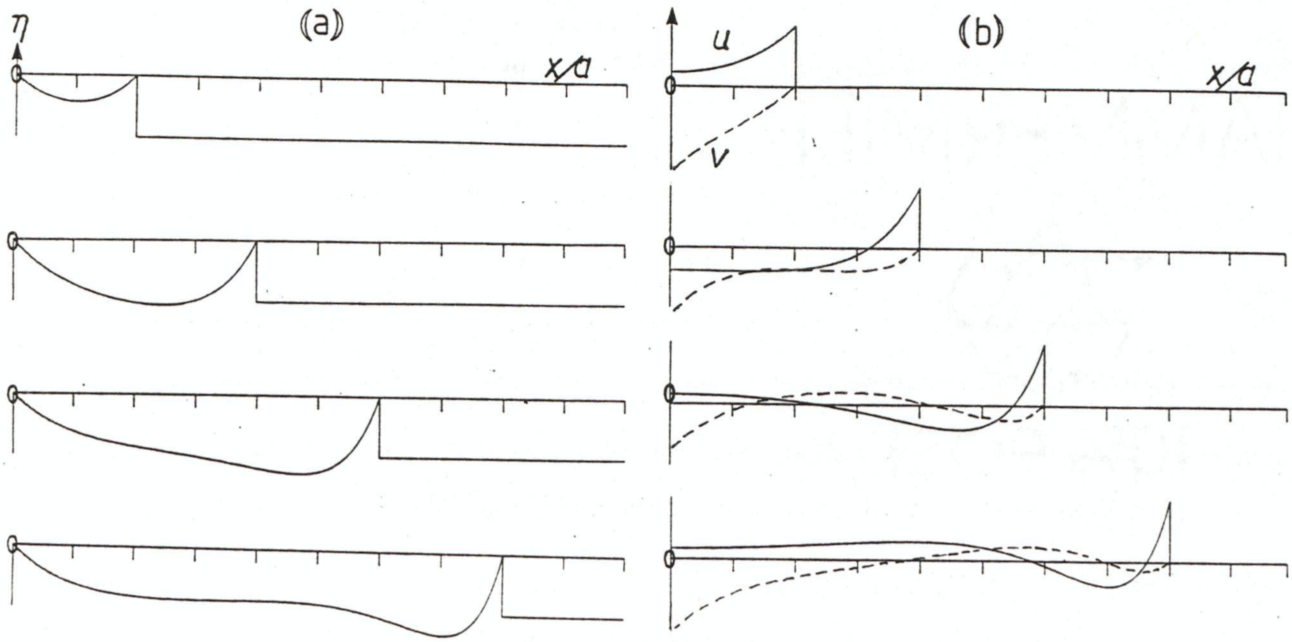


Fig. 39. Geostrophic readjustment in a fluid of infinite extent, uniform density and depth (h), and rotating about a vertical axis at rate $f/2$ (rearranged from Gill, 1982). At time, $t = 0$, fluid level in regions $x>0$ and $x<0$ are, respectively, lowered and raised by an infinitesimal increment η, imposing an initial discontinuity of 2η at $x=0$. Here illustrated, for the $x>0$ region only, are subsequent transient profiles of (a) elevation and (b) associated current components u and v after 1, 2, 3, and 4 rotation periods. The x axis is graduated in Rossby radii, $a=c/f=\sqrt{gh}/f$, in which g denotes the acceleration due to gravity. Full and dashed lines in (b) respectively display the distribution of the current components, u and v, positive u directed to the right and positive v directed into the page.

represented by superposition of an appropriate ensemble of Poincaré waves of different wavelengths. The subsequent dispersal of the ensemble is illustrated by a model fluid system of uniform depth and density in Fig. 39 (see caption for details). When the perturbing force is switched off, the Poincaré waves start to radiate from the perturbed region (from zero in Fig. 39). Because the shortest waves travel the fastest (at speed c_s for surface and at speed c_i for internal), a step-like advancing front forms. The longer waves follow behind as a "wake," with the longest, slowest waves of near-inertial frequency bringing up the rear. "At a fixed point, this is made evident by the fact that the frequency appears to decrease with time after the wave front has passed (i.e. the time between wave crests increases) and soon approaches the inertial frequency f . . . " (Gill, 1982). Herein lies the probable explanation of the different periodicities seen in the occasional near-inertial internal wave episodes seen in Lake Michigan water intake temperature records (Fig. 28). At the beginning of the above-described geostrophic readjustment process, before the shorter Poincaré waves have moved away from the vicinity of the intake pipe, the dominant frequency is higher. Later in the process, the longer waves, which are slower to depart, possess frequencies closer to f.

"The wave fronts moving away from the initial discontinuity carry energy with them, so for any finite region energy is lost . . . by radiation of Poincaré waves until the only energy left is that associated with the steady geostrophic equilibrium" (Gill, 1982), i.e. in the form of shore-parallel currents and associated Kelvin-type waves (Fig. 27).

The model system illustrated in Fig. 39 can also be applied, with c_s replaced by c_i, to two-layered and other stratified systems, in which there is an initial perturbation in thermocline level at the shoreline, $x = 0$. In fact, in his 1982 book, Gill chose the observed and modeled internal surge progression in Fig. 38 as an illustration of the geostrophic readjustment process in progress in nature, even though Simon's model embodies some additional nonlinear features and the observed rate at which the fronts migrate offshore is about 50% greater than c_i. Why this rate is higher is one of the unanswered questions posed in the following discussion.

No examples of reflection of internal surges, when they arrive at the opposite shore, or the consequent formation of standing waves, have been unambiguously demonstrated in Lakes Michigan and Ontario. Therefore, the apparent transverse internal seiches, occasionally generated or reinforced after storms and now to be described, must have other origins.

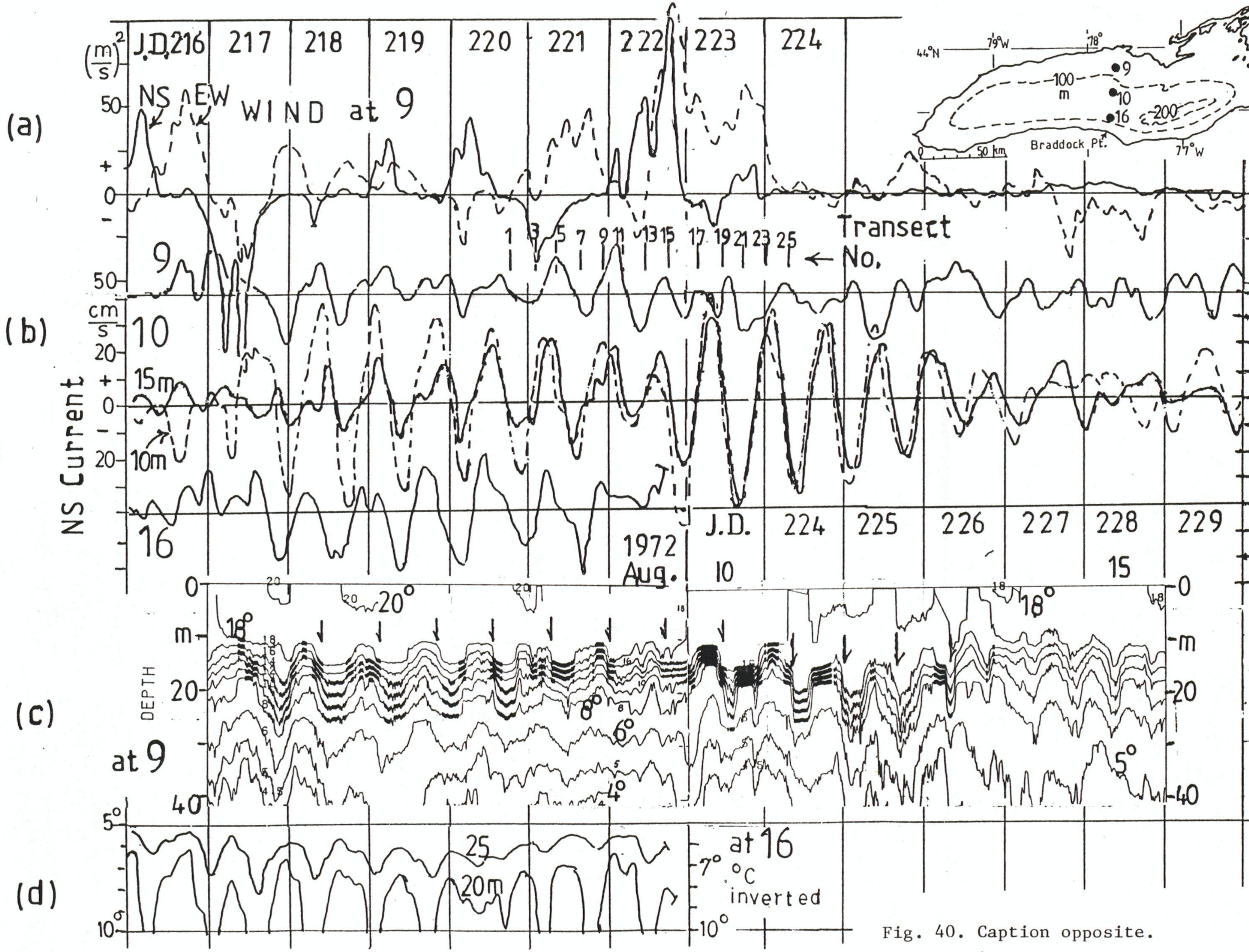


Fig. 40. Caption opposite.

(vi) Evidence of transverse internal seiches in Lake Ontario

The northward-propagating surges, described and modeled in the previous section, were not the only responses of that Lake to wind-induced thermocline displacement. Evidence of occasional whole-basin responses emerges from the examination of the fragmentary records of current and temperature distributions during repeated cross-basin transects from Braddock Pt. to Presqu'ile. For example, the 216-229 Julian Day interval, covered by Fig. 40, included two strong wind impulses. The first, from NNE on day 217, was followed by four days of weak to moderate winds and by strong rotating inertial currents at moorings 16, 10, and 9, located respectively at 12, 36 and 54 km along the transect. The second storm peaked from WSW late on day 222 with wind continuing from W on 223. The following six days were relatively calm. That storm and its aftermath coincided with the series of transects illustrated in Fig. 37.

It is evident that both storms were followed by oscillations in temperature and current, for which only the NS component of the latter at 15 m is shown. (Comparison with the EW component confirms the presence of clockwise-rotating inertial waltzing.) At 15 m, the current recorders were in the upper part of the thermocline. Records from 10 m, which would have better represented the upper layer, were available only at mooring 10 (shown here as a dashed line, for the most part cl-early in phase with the record at 15 m).

Fig. 40. (on opposite page). Lake Ontario, 1972. Wind, current, and temperature oscillations at moorings in the Braddock Pt. to Presqu'ile section, 3 to 16 August 1972.
(a) NS and EW (dashed line) components of wind speed squared at 3 m height at mooring 9;
(b) NS component of current at 15 m depth at moorings 9, 10, and 16 and at 10 m depth (dashed line) at mooring 10;
(c) isotherm depths interpolated from thermistor chain records at mooring 9; and
(d) temperature (inverted) at 20 and 25 m at mooring 16. Mooring positions and bathymetry are sketched at top right; and the intersects (at 9) with towed undulator transects (some illustrated in Fig. 37) are shown as numbered vertical lines.

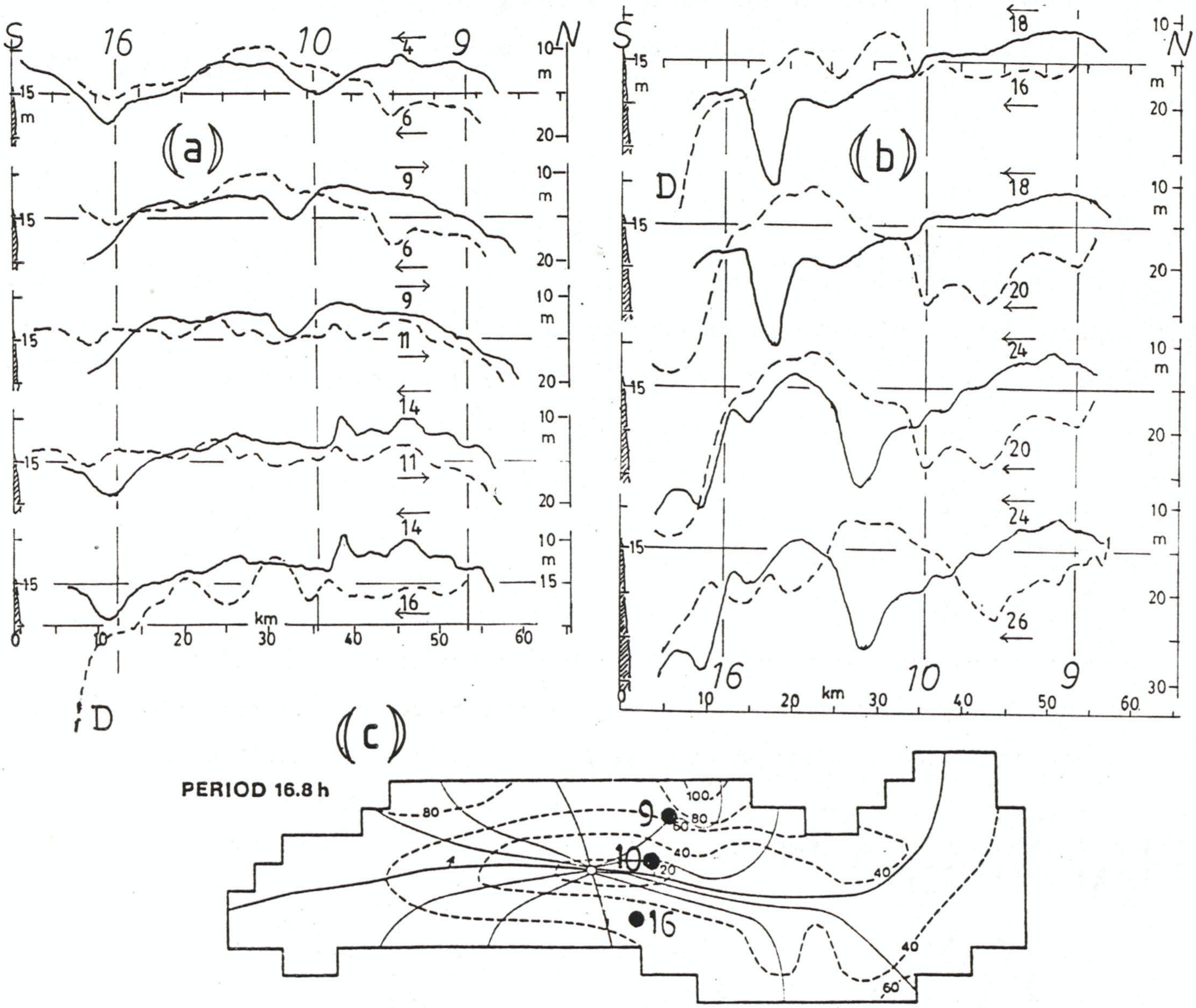
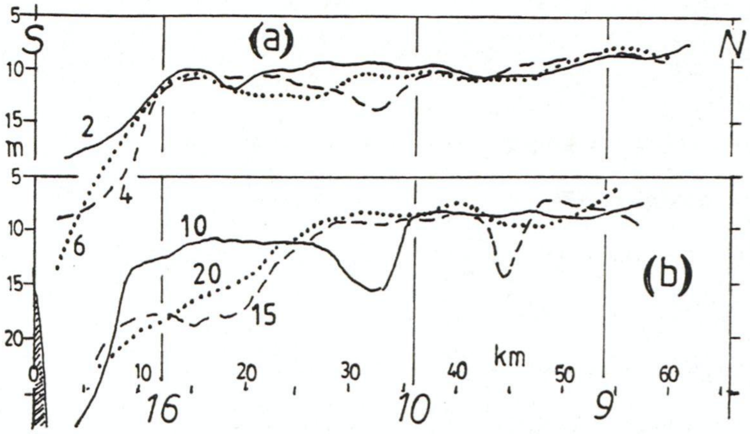


Fig. 41. Lake Ontario 1972: depth of the 10°C (mid-thermocline) isotherms in selected transect pairs, Braddock Pt. to Presqu'ile; (a) 7 to 9 August (before a storm late on the 9th) and (b) 9 to 11 August. The storm forced the downwelling at D in transect No. 16. The pairs were selected, as described in the text, to display the structure of transverse internal seiches, if present. Panel (c) illustrates a model of such a seiche calculated by Schwab (1977) for a model two-layered Lake Ontario basin of irregular plan and uniform depth. Full lines in (c) are co-phase lines at 30° intervals. Dashed lines are contours of relative amplitude.

Significant observations during the four relatively calm days after the first storm were: (1) the NS current component was nearly in phase at all three moorings; and (2) thermocline-depth oscillations at 9 and 16 were nearly out of phase. Those observations were consistent with a predominantly uninodal internal seiche pattern. Roughly estimated by eye, the average periods of the current and temperature oscillations were 17.2 h; and the peaks of N-going current coincided approximately with the times at which the isotherms at 9 were falling through their equilibrium levels. This, also, is consistent with a predominantly uninodal transverse internal seiche. During the four days in question (JD 218-221) the amplitude of the thermocline-depth oscillations decreased, consistent with expected damping by friction. Observations of cross-basin temperature distribution began late on JD 220; and the distributions of the 10^{o} (mid-thermocline) isotherms in the nine transects completed before the onset of the second storm provide evidence of a weak uninodal transverse seiche on a slightly domed thermocline. That evidence appears (in Fig. 41a) when transects are selected to coincide as nearly as possible with crests and troughs in the thermocline-depth oscillation at mooring 9 (see Fig. 40c). The trough-corresponding isotherms in Fig. 41 are shown as dashed lines. Weak transverse seiching during the pre-storm intervals was also illustrated in Fig. 52 of Mortimer (1980).

During the second storm on day 222, the picture changed dramatically (Fig. 41b). The mean level of the first post-storm transect No. 16 was depressed after storm-deepening of the upper layer; and strong downwelling at the southern shore initiated the surge sequence described in the previous section. Although perturbed by large surges and perhaps by progressive internal waves (Mortimer 1980, Fig. 54); the isotherm pairs in Fig. 41b illustrate the development of a predominantly uninodal transverse seiche with a node near mooring 10. That pattern is consistent with Schwab's (1977) model of the first Poincaré-type basin mode, calculated for a two-layered Lake Ontario of irregular plan and uniform depth, illustrated in Fig. 41(c). In



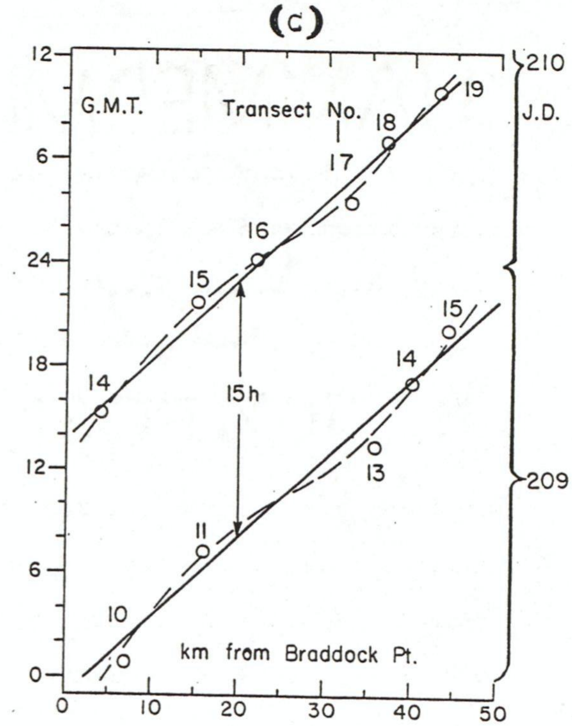


Fig. 42. Lake Ontario, 1972: Depth of 10°C isotherms on Braddock Pt. to Presqu'ile transects, (a) 24 to 26 July and (b) 27-28 July, 1972, respectively during and after a storm; (c) shows 27 to 28 July progress of two internal surges (from Mortimer 1980).

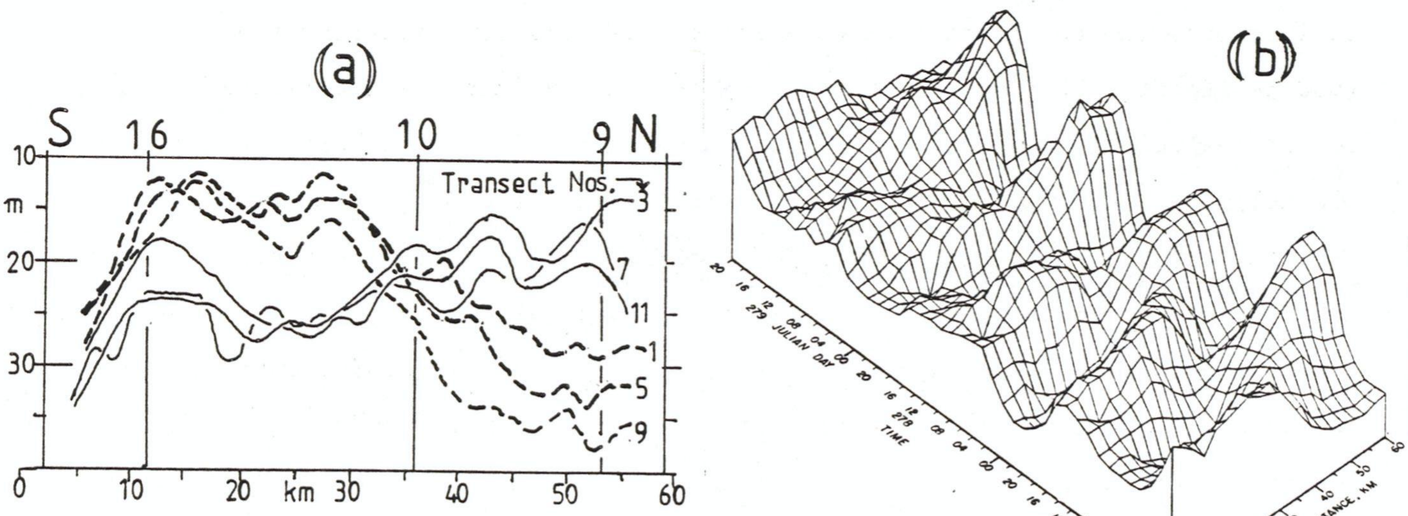


Fig. 43. Lake Ontario 1972. Braddock Pt. to Presqu'ile section, 2-4 October: (a) Mid-thermocline (10°C) isotherms in transects chosen to coincide as nearly as possible with crests and troughs in the thermocline-depth oscillations at mooring 9;
(b) 3-dimensional (isometric) representation of the thermocline surface, i.e. hourly-averaged deviations of the 10° isotherm from its equilibrium level, 2-5 Oct., using all available mooring and transect data (from Mortimer, 1977). Prior high-pass filtration through a 30 h triangular filter removed the effects of low-frequency motions.

that model, the free basin modes separate into two distinct classes: Kelvin-like modes with frequencies much less than f; and Poincaré-like modes with frequencies slightly above f. The corresponding period for the illustrated first mode Poincaré model is 16.8 h.

The observations illustrated in Fig. 40, however, show that other complicating factors are at work. While the current and thermocline-depth oscillations after the first storm both exhibited 5 cycles of average period 17.2 h, after the second storm the current oscillation at mooring 10 (6 cycles near 17 h) was accompanied initially by a thermocline-depth oscillation of higher frequency (3 cycles near 16 h) at mooring 9. The explanation of this often-observed frequency difference between current and thermocline-depth oscillations (also in Lake Michigan) may be the combination of two responses to wind impulses; geostrophically readjusting Poincaré waves emanating from the basin's rim; and inertial waltzing further offshore. The beat effects, derived from the combination of oscillations of slightly differing frequency (modeled in Fig. 20 of Mortimer 1980) may partly explain the intermittent amplitude-decaying features of offshore near-inertial current bursts after the storms in Fig. 40 and in other Lake Ontario and Michigan figures.

Examination of the aftermaths of other storms also reveals considerable variability. An earlier set of transects (JD 206 to 210), divided into pre-storm and post-storm groups in Fig. 42(a) and (b), demonstrated the following storm effects: intensification of previous south-shore downwelling (transect 10); subsequent northward progress of internal surges (Fig. 42c); but no conspicuous transverse seiching, although weak evidence of a binodal transverse seiche (not shown) emerged during later windless days (211-213).

As a contrast, the set of transects (Figs. 43 and 44) which followed a storm on day 274, displayed a conspicuous transverse internal seiche during post-storm days (275-278) of relative calm. But, in spite of southern downwelling after the storm, no conspicuous surges were seen during this episode. The uninodal character of the seiche is demonstrated by the assemblage of information on the depth of the 10° isotherm, derived from all moorings and transects, in

Fig. 43, 44. As in Fig. 35 for Lake Michigan, the aliasing effect of vessel progress is removed by confining the information to selected distance intervals across the transect. Current records are, unfortunately, only available from mooring 9. The seiching was particularly persistent in this example. Seven cycles of current and thermocline-depth oscillation showed an average period of 16.8 h.

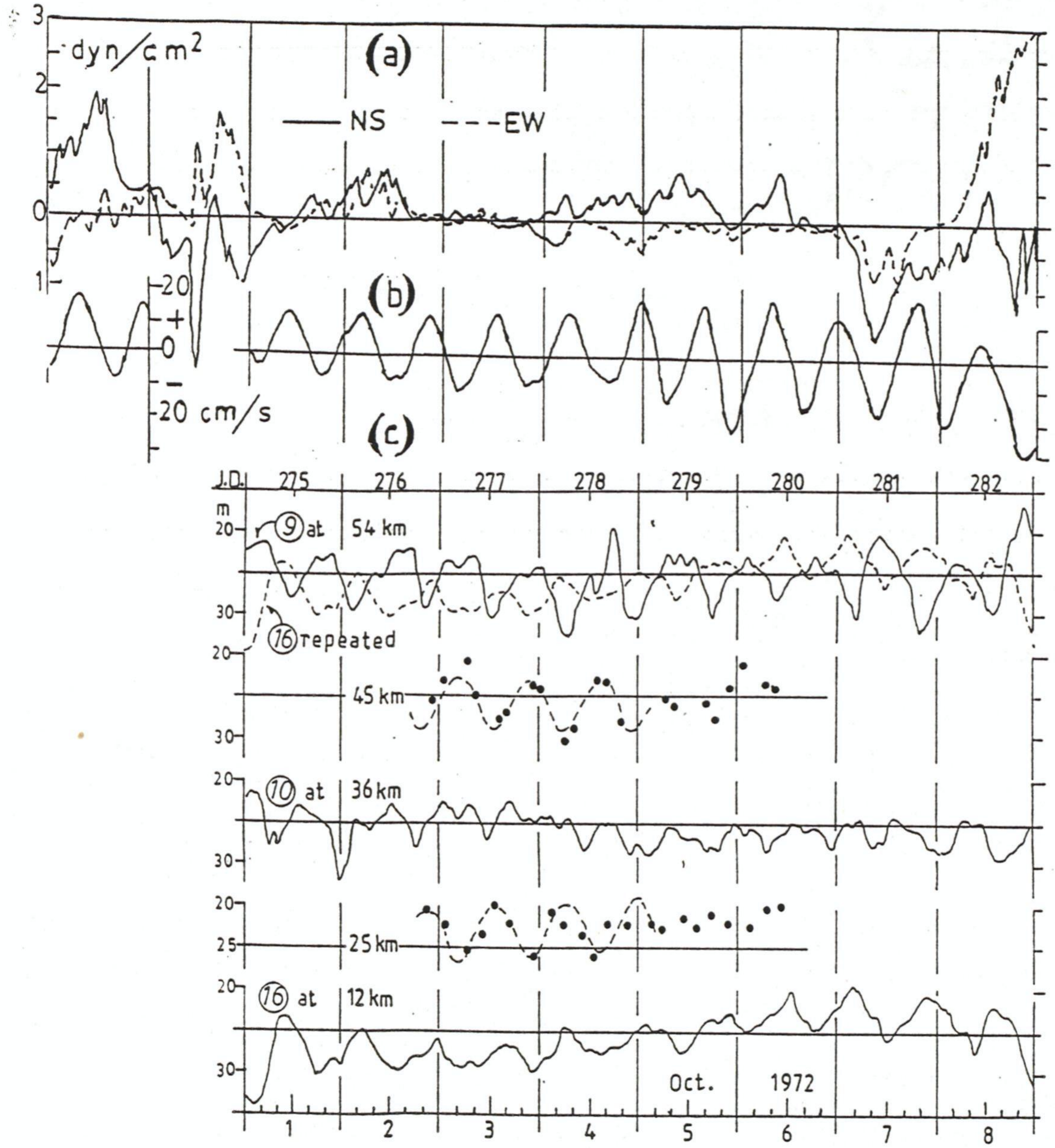


Fig. 44. Lake Ontario, Sept. 30 to Oct. 8, 1972, Braddock Pt. to Presqu'ile section: (a) wind stress (dyn/cm^2) and (b) NS component of current at 15 m depth at mooring 9; (c) depth of the 10^o (mid-thermocline) isotherm at distances 54 (mooring 9), 45, 36 (mooring 10), 25, and 12 km (mooring 16) from Braddock Pt. The 45 and 25 km data (dots with 3m added as prompted by discussion of sensor time constant in Boyce and Mortimer, 1977) are from transect passages and are fitted to a sinusoid of range $\pm$3m and period 17 h.

Questions unresolved for large lakes: a concluding discussion.

The previously-described observations in Lakes Michigan and Ontario enable us to list and provisionally interpret the free internal responses of large stratified basins to impulsive wind stress. Large basins, by previous definition, are those which are much wider than the Rossby radius, $a = c_i/f$. Wind action is seen to change the stratification through turbulent mixing and deepening of the upper layer. It also displaces isotherms from their equilibrium positions, thereby generating rotation-influenced internal waves and currents. Understanding of those field observations has been advanced, principally during the last two decades, by model building and testing. That activity was notably stimulated by the IFYGL (Aubert and Richards, 1981) as demonstrated by the Fig. 38 example. An illuminating and comprehensive account of those models and of attempts to verify them is to be found in Simon's (1980) monograph. Simon's outstanding contributions, in particular, are reviewed in the introductory essay (Schertzer, Boyce, and Murthy, 1988) to his Collected Reprints.*

Here we preface the parade of some unresolved questions with a summary of what can now be recognized as four categories of basin response to wind stress: 1, offshore inertial waltzing and Ekman drift; 2, nearshore dynamics of up/downwelling, forced by the Ekman drift; 3, geostrophic readjustment to response 2, marked by offshore migration of Poincaré wave ensembles and fronts and nearshore generation of shore-trapped Kelvin waves; and 4, basin-steered dynamics of internal seiches, when post-storm calms permit. For brevity, it is convenient to refer to those responses by number, bearing in mind that timing and location of the wind stress determines whether they act in concert or in single dominance, together or sequentially. Their expression and

*J. T. Simons' early death robbed us of his inspiration, wisdom, and future promise. Tragically, the same must be said of two other masters of this field: A. E. Gill and N. S. Heaps.

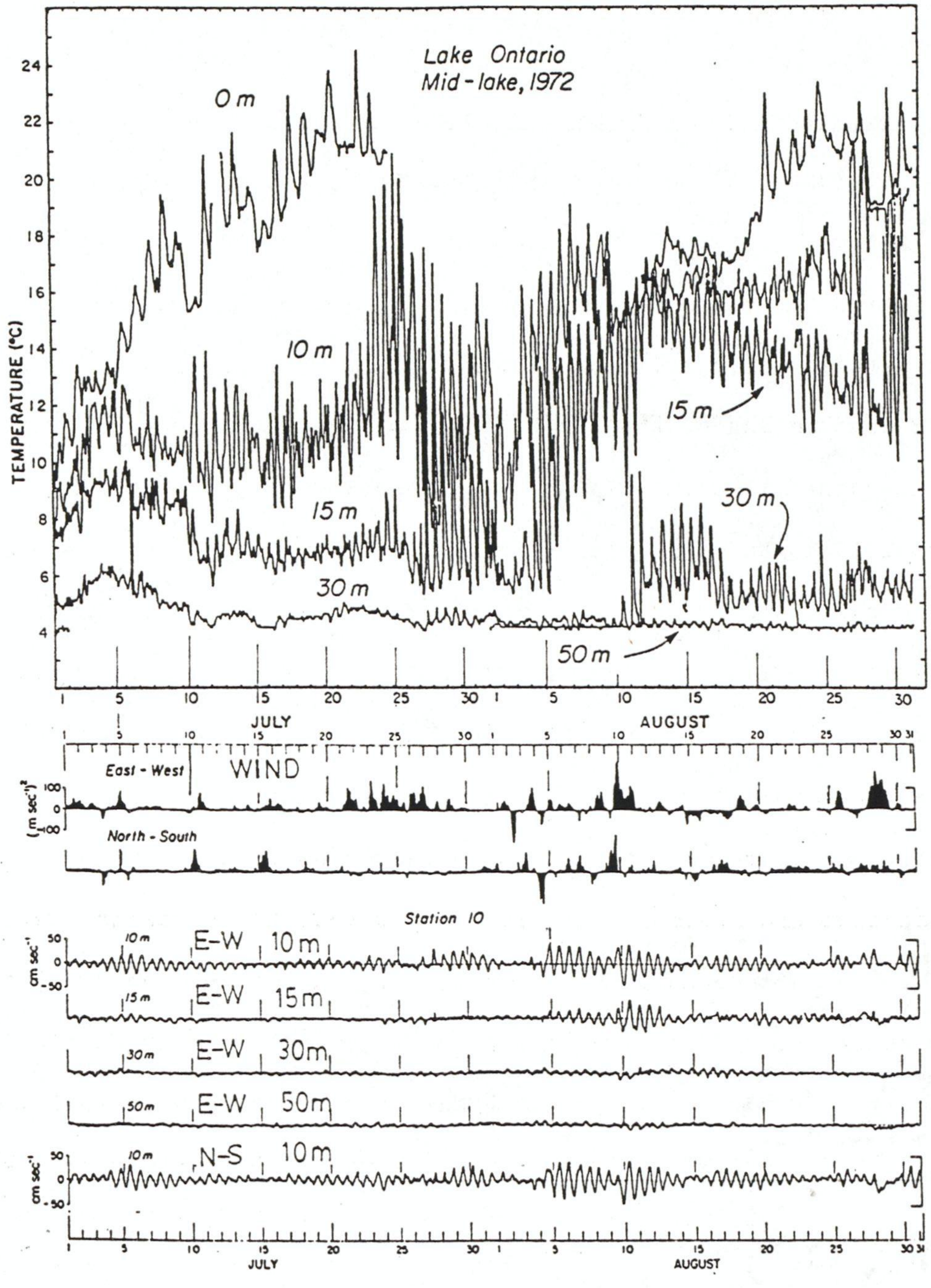


Fig. 45. Intermittent inertial motion (temperature and current oscillations) at mooring 10, Lake Ontario, July-August, 1972 (IFYGL data, Marmorino and Mortimer, 1978).

persistence are strongly dependent on the duration of the wind impulse and on its phasing with pre-existing motions. Expressions of responses 3 and 4 also require an interval of relative calm in the aftermath of the initiating storm. It is more usual for those responses to be aborted by a subsequent wind disturbance.

(i) Interaction with topographic Rossby waves

Before proceeding further, the possibility of interaction between the above-listed responses and a fifth response -- the topographic Rossby wave -- must be considered briefly. Responses 2 to 4 arise as a consequence of wind stress and stratification, i.e. they are baroclinic, and gravity is the restoring force. Topographic Rossby waves, by contrast, are governed, not by gravity and stratification, but by vorticity effects when water masses move over sloping bottoms. Sometimes named "vorticity waves" or "vortex modes", their motion is integrated over the whole water column, whether stratified or not; i.e. it is barotropic. For that reason, vorticity waves as such are not further considered in this review. In Lake Michigan, Saylor et al., 1980, Huang and Saylor, 1982, Schwab, 1983, showed that the dominant period of the vorticity wave is about four days, and that the associated current speeds can be as high as 10 cm/s. For oscillations of frequency near f, i.e. Poincaré waves, the vorticity wave, when present, adds a slowly-varying current. But, for the low-frequency Kelvin and for vorticity waves, nearshore speeds of propagation are similar; and both waves progress cyclonically around the basin. Their strong interaction in Lake Ontario was demonstrated and modeled by Simons (1980). This will probably also be demonstrated in Lake Michigan. Clearly, barotropic/baroclinic interactions pose many, as yet, unanswered questions.

(ii) Questions arising in connection with offshore Response 1 (inertial waves and Ekman drift)

Inertial "waltzing" is an immediate and universal response offshore to change in wind stress, or to other changes in the status quo, where the clockwise-rotating inertial currents are not impeded by boundaries and where the "pure" inertial wave models of Sverdrup (1926) and Pollard (1970) apply.

However, in Lakes Michigan (Figs. 30,31) and Ontario (Figs. 44,45), the observed offshore responses are rarely pure in that sense, because not all the energy is concentrated in currents at frequency f. Some energy also appears in thermocline-depth oscillations, often "blue-shifted" to frequencies slightly above f. That shift raises further unanswered questions. Are we, in fact, observing pure inertial motion "contaminated" by Poincaré waves of higher frequency migrating from an inshore Response 3 event? If that is so, why is the observed current-rotation frequency often distinctly closer to f than is the temperature oscillation frequency (example in Fig. 31)? Research in coastal seas may hold the clue; and the explanation may lie in coastal influences other than Response 3 (Kundu, 1976, 1986, Kundu et al. 1983).

As illustrated in Figs. 31 and 45, Response 1 in Lakes Michigan and Ontario are characteristically intermittent. Which factors, apart from wind variability, contribute to that intermittency? Possibilities are: beat phenomena between inertial motion proper and Poincaré waves of higher frequency passing through from Response 3 events, or rhythmic to-and-fro interchange of energy between the upper mixed layer and deeper water (Gill, 1984).

The thermocline oscillations which often accompany Response 1 (as contaminants?) often exhibit nonlinear features: steep-frontedness and a conspicuous harmonic content of frequency $2f$. That content is revealed in spectra (Figs. 32 and 46) and is sometimes large enough to be seen in untreated records, as in the Lake Michigan and Ontario examples (Figs. 30 and 44, respectively). Variability in the $2f$ content is illustrated by rotary spectra of successive 5-day samples of the July-August current records (Fig. 46). During those months, current rotation was predominantly clockwise; and most of the energy was sharply tuned at frequency f. Much smaller peaks were also present at frequency $2f$, but showed considerable variation in height from one 5-day interval to the next. What is the origin of those $2f$ peaks, seen in both Lakes Michigan and Ontario; and what accounts for the variability in their height?

In general, inertial motion or an approximation to it is the universal response to changes in the offshore status quo, for example, increase or decrease in wind stress, impulsive change in current regime, or a sudden change in frontal dynamics. Probable inertial effects of the latter kind are displayed in Fig. 47, which (in satellite-derived surface temperature images from 16 May and 24 June) illustrates the offshore advance of the thermal front, which characterizes the spring warm-up phase each year in Lake Michigan. On the inshore side of the fronts (dark areas), the water column is stratified and warm at the surface. On the offshore side (light areas), the column is well mixed and cold. Unless disturbed by wind, the front moves steadily offshore. On 16 May, both western and eastern fronts were extensive and relatively stable. But,

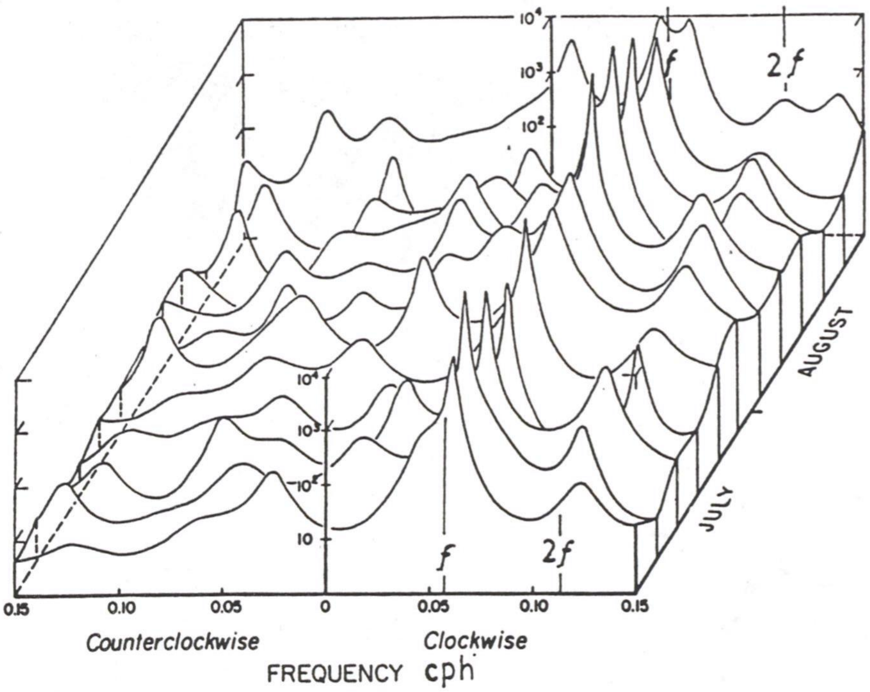


Fig. 46. Rotary spectra of currents at 10 m depth (successive 5-day samples of the record illustrated in Fig. 45, from Marmorino and Mortimer, 1978).

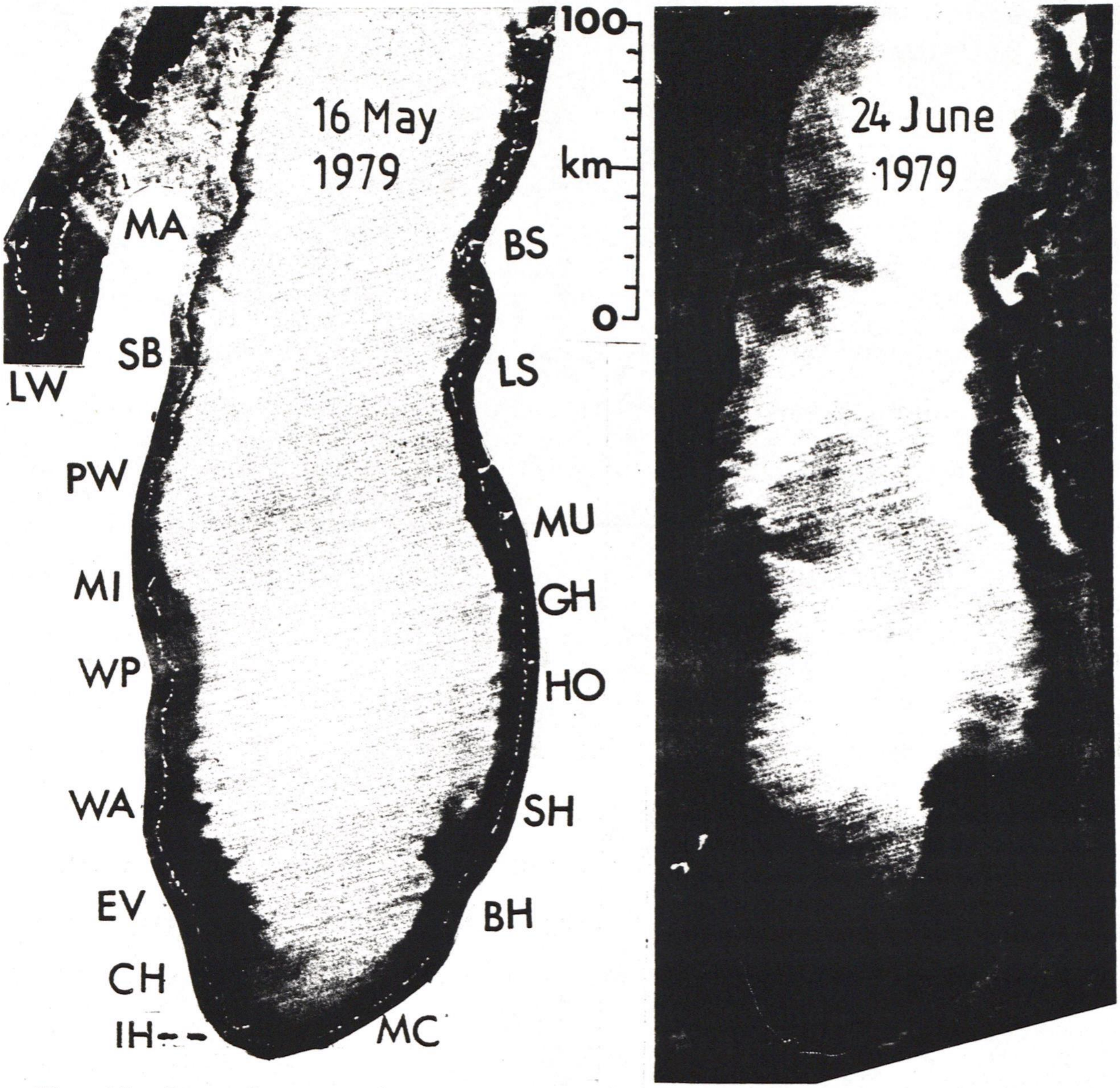


Fig. 47. Coastal Zone Color Scanner images (channel 6, the thermal band) of southern Lake Michigan on 16 May and 24 June 1979 (Mortimer 1988). Darker and lighter areas indicate warmer and colder surface temperature, respectively. On 16 May the eastern temperature front was stable; on 24 June upwelling had occurred onshore; and the front shows evidence of probable inertial response to nearshore upwelling.

by 24 June and after several days of S-directed wind, there was extensive upwelling of cold (near 6°) water, shown as light areas inshore along the eastern coast. The eastern front no longer defined as a line, but was strongly perturbed by eddies, giving it a remarkable braided rope-like appearance. Does that suggest the presence of transient inertial waves, generated during a geostrophic adjustment forced by the upwelling? In ocean fronts, Kunze and Sanford (1984) have demonstrated that inertial waves, generated and trapped in frontal regions, are energetic enough to play an important role in the dynamics of those fronts.

When the wind stress is sustained over the basin, the whole upper layer begins to drift, to the right when looking downwind. Ekman's (1905) simplified model of that process is illustrated in Fig. 48. That model is clearly an over-simplification for a large stratified lake, because it assumes that the mixing (eddy diffusion) coefficient is constant throughout the water column. More elaborate models of Ekman flow in Lake Michigan were constructed by Birchfield (1973) and Birchfield and Hickie (1977). But Fig. 48 serves to illustrate the important general conclusion that, in the N hemisphere, the average drift of the upper layer is directed to the right when looking downwind.

(iii) Response 2: nearshore upwelling and downwelling

Along the righthand shore, looking downwind, the shoreward Ekman drift of the upper layer causes the thermocline to be depressed in a nearshore strip of width one or two Rossby radii. Along the opposite shore, there is a strip of comparable width, in which there is a compensatory elevation of the thermocline (upwelling). An example of that downwelling/upwelling asymmetry appears, after strong sustained south-directed wind, in Fig. 32.

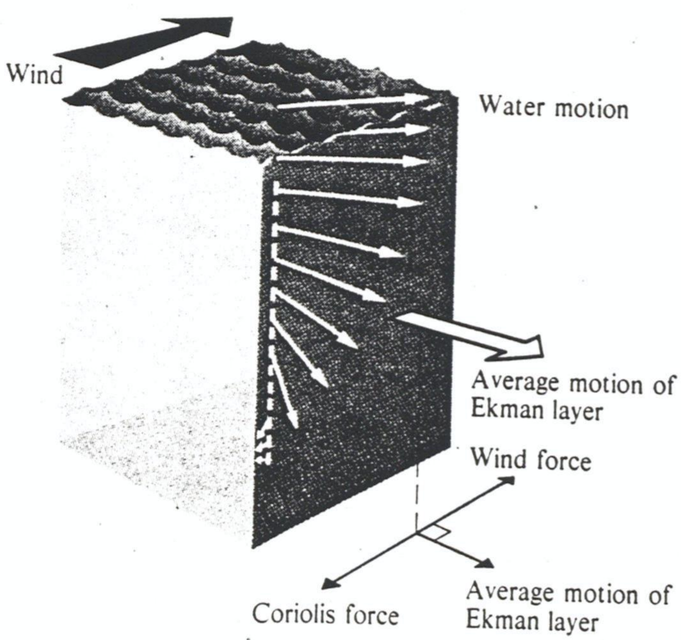
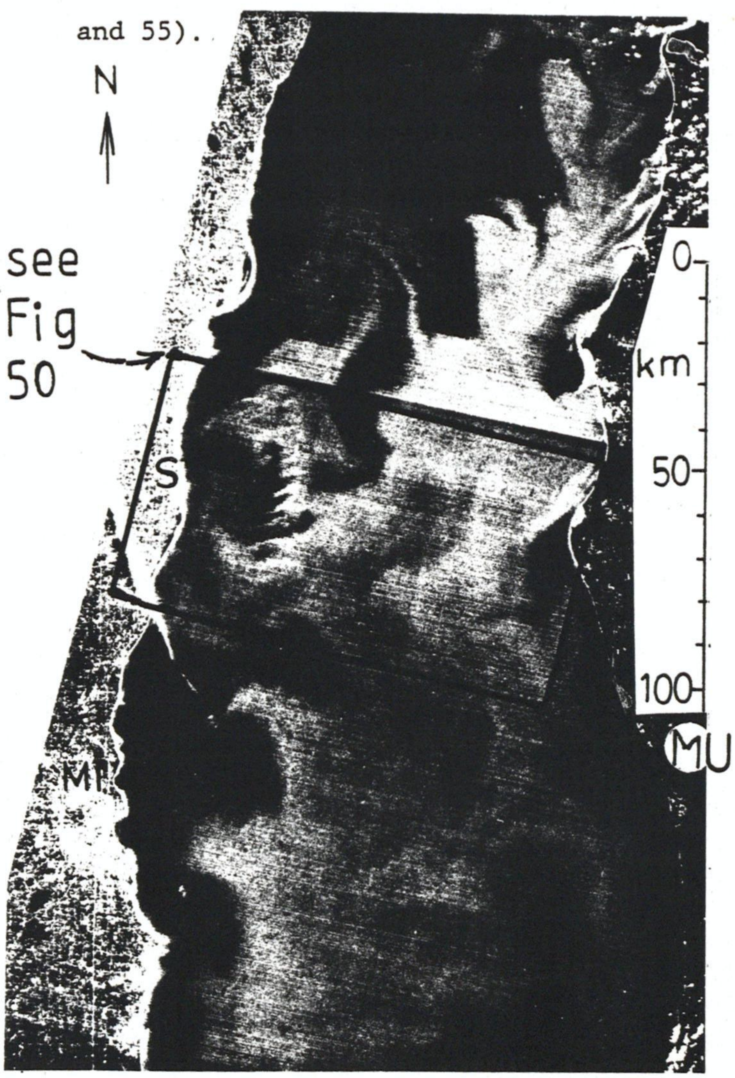


Fig. 48. Model of drift currents in an Ekman layer responding to sustained wind stress (from Perry and Walker, 1977).

If the alongshore component of the wind stress is strong enough, the upwelling thermocline is raised to the lake surface, i.e. Csanady's

(1977) "full" upwelling. With a two-layered model and examples from Lake Ontario, Csanady demonstrated that the magnitude of the minimum wind impulse, required to produce full upwelling, is "close to the product of top layer depth and propagation velocity of long waves on the thermocline." When the wind impulse exceeds that minimum, the upwelled thermocline front moves further offshore, and simple theory no longer applies. The offshore wind component produces little effect, but adjustment to continued or subsequent alongshore wind stress "produces a more complex thermocline shape and under certain conditions a frontal countercurrent". Csanady cites a Lake Ontario example; and a similar process may explain thermocline shapes observed during sustained upwelling along the eastern shore of Lake Michigan (later Figs. 54 and 55).



Some features of the upwelled thermocline front -- large eddies of order 20 km diameter -- are illustrated in Fig. 49. Rao and Doughty (1981) view those features as evidence of baroclinic instability. How long those eddies persist after removal of the stress, and what effect they have on horizontal mixing and trans-thermocline exchange, has yet to be determined.

Fig. 49. LANDSAT view of a fully upwelled thermocline front in Lake Michigan, 21 Aug 1973, rendered visible by $CaCO_3$ precipitation ("whiting") in the upper layer (Strong 1978). Sub-thermocline water, which has risen to the surface along the W shore, remains dark. The box at S is enlarged in Fig. 50.

(iv) <u>Response 3: geostrophic readjustment after nearshore downwelling</u>

This response, which comes into play as soon as the wind stress is removed, is initiated by the downwelling/upwelling asymmetry. If, in the two-layered model, the thickness of the upper layer is much less than that of the lower layer, and if the wind stress has been strong enough, surge-like features moving offshore from the downwelled region dominate the response, as in Fig. 38. The initial downwelling may be represented as a two-part set of periodic waves, comprising (i) Poincaré waves of near-inertial and higher frequencies and (ii) low-frequency Kelvin waves, both released to propagate when the wind stops. Sub-set (ii) remains trapped to propagate alongshore at group velocity c_i. Sub-set (i), which propagates offshore, is dispersive, i.e. the group velocity is wavelength-dependent. The group of shortest waves, which may form a front, travel fastest at speed c_i in the simplest model (Fig. 39). Observations (Figs. 38 and 42) disclosed higher group (surge) speeds, reproduced in Simon's (1978) model, when the nonlinear accelerations and advection of momentum normal to the shore were taken into account.

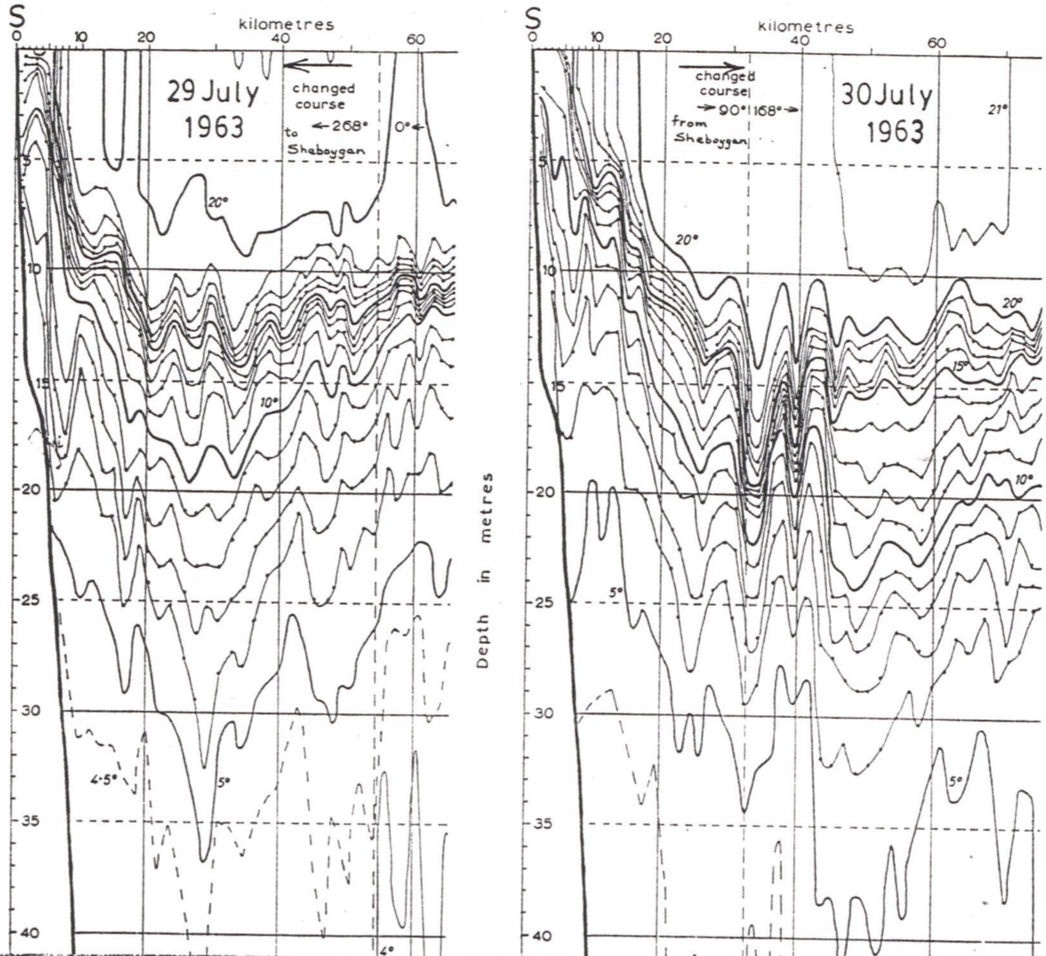
Let us examine a relatively simple situation in which a strong sustained wind impulse gives rise to offshore inertial waltzing and inshore downwelling. When calm is restored, the Poincaré wave ensemble moves away from the downwelling region, with the waves of shortest length leading, followed by a trailing wake of longer wavelengths. The ensemble advances into regions previously occupied by inertial (Response 1) waves only. The frequency of the leaders of the Poincaré ensemble is $>f$ and falls closer to f in the trailers. Thus, at a given offshore location, the frequency sequence will be: f before and $>f$ after the Poincaré leaders arrive, followed by a progressive decrease toward the f limit as the trailers go by. That speculation awaits testing, but it may explain the range of periods, seen in water intake temperature records (Fig. 28) on the rare occasions when the thermocline hovered near Lake Michigan water intakes for several days. (It was those, at first, puzzling findings which triggered my 1963 paper and subsequent engagement with that Lake.)

As Simons (1980) pointed out, the propagation speed of the Poincaré ensemble is similar to the speed of the inertial currents after strong wind in the offshore region into which the ensemble advances. Therefore,

> "the nonlinear acceleration terms in the equations of motion can no longer be neglected. This leads to two results ... [see Fig. 38]. In the first place, the propagation speed varies periodically in a frame fixed to the basin, because the wave advances into a medium that rotates with the inertial frequency. Secondly, the leading sides of the waves may steepen and form internal surges. For a basin with ratio of upper- to lower-layer depth less than unity, the latter effect is known to be limited to the case of depression waves. As the thermocline at the shore oscillates with near-inertial period, it is in principle possible for a series of such surges to emanate from the downwelling shore."

This conclusion is confirmed for Lake Ontario examples in Figs. 38 and 42 and provisionally also by later examination of Lake Michigan cross-sections after storms (Figs. 54,55); but several questions remain. What forms do the re-adjustment processes take at upwelled fronts? We may recall that it was also shown earlier that, in small lakes when thermocline depth is less than basin depth, thermocline downswings generate surges, but upswings do not.

On rare occasions, group or packets of internal waves of wavelength-order 5 km have been observed in Lake Michigan. One such example, perhaps moving offshore from an upwelled region, is illustrated in Fig. 50. Two isotherm-depth sections were run into and out of Sheboygan (S) during the afternoon and early morning of 29 and 30 July, 1963, respectively (Fig. 50a). The vessel tracks are sketched in Fig. 50(b). The latter figure is an enlargement of the box at S in Fig. 49, in which $CaCO_3$ precipitation had produced "whiting" in surface layers. The banding seen in (b) may arise



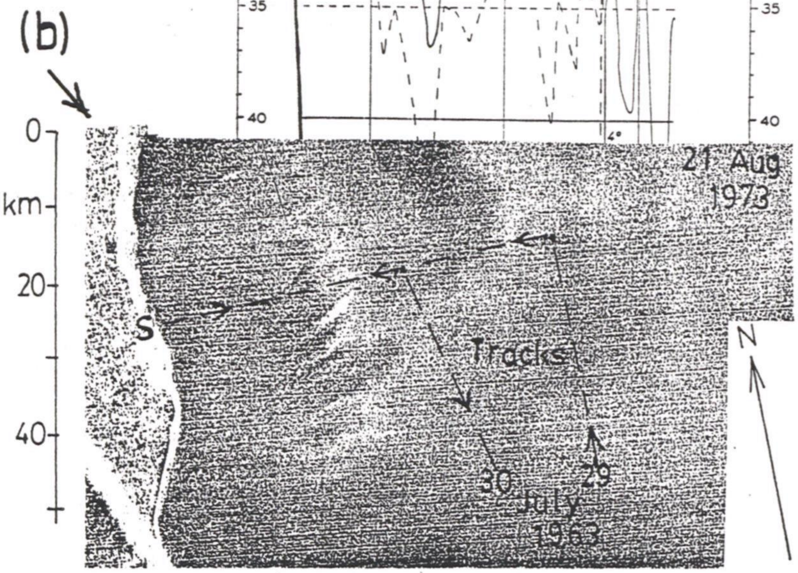


Fig. 50. Signals (?) from internal wave packets in Lake Michigan seen in:
(a) isotherm depths along tracks into and out of Sheboygan, Wis. (S) on 29 and 30 July 1963; and
(b) Landsat (green band) image near S on 21 Aug. 1973 (see Fig. 49 and text).

because the milky layer is thinner above the crests and thicker above the troughs of internal waves. The wave packet in Fig. 50(a) appears to have moved offshore by about 15 km and increased in amplitude in the approximately 11 h interval between the vertical dashed (turning point) lines. The corresponding propagation speed, ~1.4 km/h, is close to the local value of c_i. However, the "crests" of the postulated wave in Fig. 50(b) suggest an alongshore progress. How are such wave packets generated; and what is their influence, if any, on cross-thermocline exchanges?

Compared with the immediacy of Response 1, the geostrophic readjustment, Response 3, is relatively long-drawn-out. This is particularly true if one considers the shore-trapped Kelvin waves, left behind after the Poincaré ensemble has departed into offshore regions. At typical c_i speeds near 1.5 km/h, even the fastest wave groups in the Poincaré ensemble, emanating from a downwelling region near Milwaukee, would require 85 h to reach the opposite shore. Subsequent reflection to set up a standing wave pattern would take even longer.

(v) Response 4: changes in cross-basin thermocline structure, including transverse seiches

It follows from the foregoing paragraph that a very long post-storm calm would have to be maintained before a transverse (standing wave) seiche could be generated directly by Response 3. That was the justification for Fennel's (1989) contention that my Lake Michigan observations, in Fig. 34 for example, should be interpreted as an interpenetration of progressive Poincaré wave transients emanating from both sides of the basin and not (as in Mortimer 1971) as a set of transverse seiche modes. However, it will later be demonstrated (Fig. 55) that an evident transverse seiche does, in fact, appear within two days of the end of a major storm. In the meantime it is instructive to introduce Fennel's model (Fig. 51) -- a long, stratified channel of uniform width 100 km, depth 50 m, rotating with inertial frequency $f = 1.2 \times 10^{-4}$/s (inertial period $T_i = 14.5$ h). Wind at 12m/s, was applied at time $t = 0$ to the whole channel in an alongshore direction, "as a body force evenly distributed within a pre-existing mixed upper layer of thickness" 15 m. The continuous vertical density profile corresponded to a constant Brunt-Väsälä frequency $N = 0.01$/s. The initial responses in Fennel's model to the suddenly-applied wind stress were twofold: (i) "inertial oscillations"; and

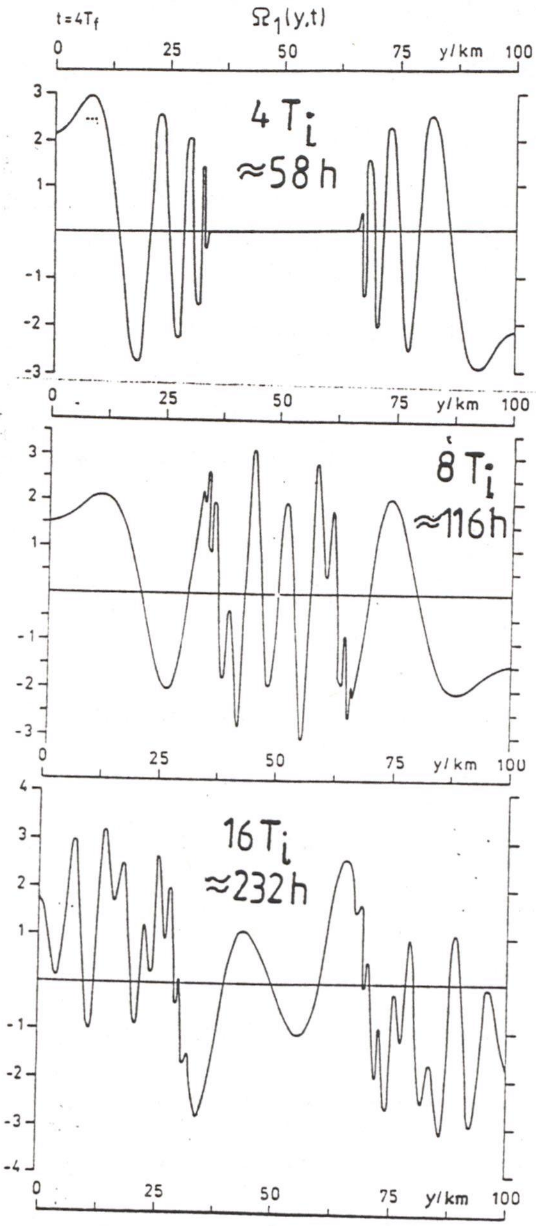


Fig. 51. Fennel's (1989) model of Poincaré wave ensembles moving dispersively offshore from both sides of a 100 km-wide channel. Distribution of the quantity Ω, proportional to the vertical velocity component, is depicted at intervals of 4, 8, and 16 inertial periods (T_i, with corresponding approximate time in days added for a Lake Michigan equivalent). At $4T_i$ the two fronts are advancing toward mid-channel into a region of pure inertial oscillations with no vertical component. The fronts are followed by a wake of longer waves. At $8T_i$ the wave ensembles have begun to interpenetrate. At $16T_i$ each ensemble has reached, and is being reflected from, the opposite shore. Note the symmetry about an apparent mid-channel "nodal point".

(ii) "inertial waves". The quotation marks indicate Fennel's terminology, whereby (i) corresponds to Response 1, in which the upper layer of the whole channel moves bodily in an inertial circle at exact frequency *f*, with no vertical component of motion. His model response (ii) is an ensemble of Response 3 dispersive Poincaré waves (Fennel's "Poincaré sums") radiating from both shores with wavelength distributions determined by the form and dimensions of the initial upwelling/downwelling disturbances. The progress and merging of the two ensembles is illustrated in Fig. 51, with explanatory details in the caption.

The rate of frontal progress in Fig. 51 (ca. 0.6 km/h) is less than half the speed, which would be seen in a 2-layered model fitted to Lake Michigan parameters. Whereas it takes 9 days or more in Fennel's model before the effects of side reflection become evident, the presumed standing wave pattern in Figs. 34 and 55 was in place less than two days after the storm on 16/17

August. The question is: was it a standing wave pattern? Fennel thinks not, and points to the potentially misleading aliasing effect of measurement from a moving ferry. But, when that effect is removed, as in Fig. 35 and Figs. 71-74 in Mortimer 1971, an apparent mid-lake standing wave pattern emerges.

(vi) Evidence of Response 4 extracted from repeated scans of cross-basin thermocline structure in Lake Michigan

I shall now attempt to resolve this debate by a closer scrutiny of the changes in cross-basin shape of the thermocline as displayed by the 1963 ferry transects in comparison with wind records from the FWPCA buoys at moorings 18 and 20 (see map in Fig. 29). In the following Figs. 52, 54, and 55, 10°C isotherms are selected to represent the thermocline interface in the Milwaukee (MI) to Muskegon (MU) section from 17 July to 23 August (Mortimer 1968, with transects here re-numbered in sequence -- details in Table 2 and in the caption of Fig. 52). The isotherms, displayed in pairs to reveal changes in thermocline topography, exhibit a variety of features: (a) slow changes extending over many transects; (b) traveling internal surges; (c) progressive and standing Poincaré waves; and (d) interactions of (b) and (c) with the ferry's progress. To enable the effect of (d) to be judged, the elapsed time (Δt) between each pair of transects is entered in italics in the figures at 20, 40, 60, and 80 km from MI.

Fig. 52 covers two weeks of light variable winds. The last week of July was calm, except for a short pulse of wind toward SW on the 21st and weak sustained wind toward N, 26th to 28th. Those weak stresses imposed no obvious change on the general persistent shape of the thermocline, i.e. a shallow dish with weak upwelling near MI and a relatively stationary trough and ridge features at 100 and 110 km from MI, respectively. Speculation, that those

Table 2. Particulars of selected 1963 ferry runs to and fro across Lake Michigan from Milwaukee (MI) to Muskegon (MU) yielding temperature distribution transects (Mortimer, 1968) from which 10°C isotherm depths were extracted in Fig. 33. Dates (of transect start), direction, and times of passage through points 10 and 120 km from MI are indicated. Also listed (as an index of onshore up- or downwelling) are the temperatures, at those times, in the municipal water intakes at MI (18 m depth) and MU (8 m).

in Fig.	Transect Nos.	in Mortimer 1968	Dates July	Direction towards	Times (CST) at points from MI 10 km	120 km	Water intake temperature, °C MI (18 m)	MU (8 m)
47	4	161	17	W	1530	1200	7.4	18.6
	5	163	18	W	2035	1700	6.7	20.0
	7	165	19	W	1745	1410	4.9	21.9
	9	167	20	W	0920	0550	4.9	21.9
	11	169	20	W	0145	2045	5.6	17.2
	13	172	21	E	2330	0430	5.1	15.6
	14	173	22	W	1845	1400	6.5	17.5
	15	174	22	E	2215	0300	6.2	16.7
	16	177	24	W	0500	0020	8.8	18.9
	17	178	24	E	0900	1400	9.2	20.6
	18	179	24	W	2200	1715	10.1	20.0
	19	180	25	E	0100	0545	9.9	20.0
	20	181	25	W	1720	1240	11.2	20.6
	21	182	25	E	2050	0140	10.4	20.6
	22	183	26	W	0920	0435	10.6	22.2
	23	184	26	E	1220	1700	7.1	22.5
	24	185	26	W	--	2050	5.8	22.2
	25	186	27	E	1130	1500	5.3	24.5
	26	Cisco	29	E	0550	--	--	--
Fig. 48	27	494	30	E	1110	1425	4.2	23.9
	29	496	31	E	0450	0805	3.9	24.2
	30	Cisco	31	W	1245	--	--	--
	32	498	31	E	2040	0000	3.7	24.2
	32a	Cisco	Aug.1	E	0405	--	--	--
	34	500	1	E	1135	1505	3.9	22.5
	36	Cisco	2	E	6400	--	--	--
	37	504	2	E	2215	0140	4.0	23.3
	39	505	3	W	1710	1355	3.3	23.9
	40	506	3	E	2110	0025	3.6	23.9
	41	507	4	W	1440	1120	4.7	23.9
	42	508	5	E	0010	0325	5.6	21.1
	42a	Cisco	5	E	--	1210	--	--
	43	191	6	W	1535	1205	4.2	23.6
	44	192	6	E	2000	0030	4.2	24.2
	45	193	7	W	1215	0740	4.2	24.2
	48	197	8	W	0045	2000	4.4	17.8
Storm	49	198	9	E	0400	0845	4.4	20.6
	50	201	10	W	1515	1040	9.3	7.2
	51	203	11	W	1540	1100	6.6	8.9
Storm	54	206	12	E	2300	0340	8.5	16.7
	54a	209	14	W	1220	--	12.8	8.1
	55	210	14	E	1530	2000	13.2	7.2
Fig. 49	56	211	14	W	0400	2330	13.3	7.2
	57	212	15	E	0845	1330	13.1	7.2
	58	213	15	W	2045	1615	13.7	7.2
	59	214	16	E	0030	0500	13.8	8.7
Storm	60	215	16	W	1200	0730	14.0	9.4
	61	221	18	W	1415	0945	15.2	7.2
	62	222	18	E	1645	2145	15.2	6.7
	63	224	19	E	0930	1400	15.0	6.9
	64	225	19	W	2220	1730	15.1	6.9
	65	226	20	E	0145	0620	15.1	8.9
	66	227	20	W	1400	0915	15.0	9.7
	67	228	20	E	1700	2145	15.2	13.3
	68	229	21	W	0530	0100	15.0	15.6
	69	230	21	E	0845	1335	15.0	14.4
	70	231	21	W	2115	1640	14.8	15.0
	71	232	22	E	0040	0515	14.8	15.6
	72	234	22	E	1545	1910	14.8	18.3
	73	235	23	W	0250	2320	14.8	18.3

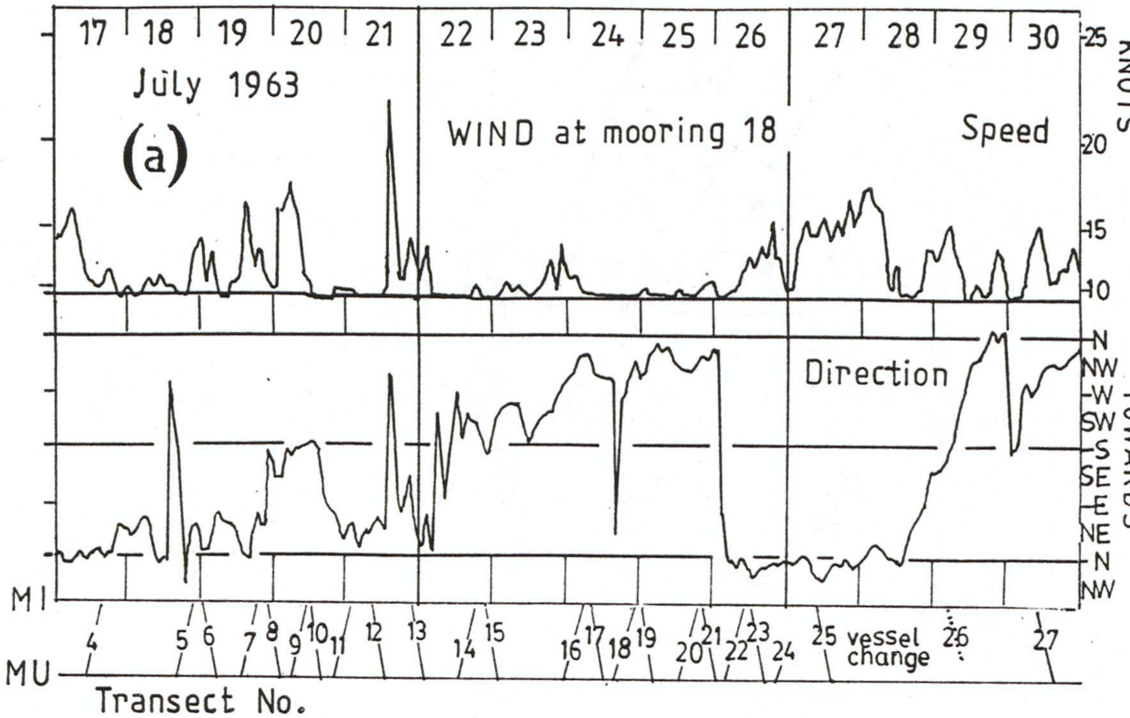


Fig. 52. Lake Michigan, 17 to 30 July 1963: (a) wind speed and direction (towards) at FWPCA mooring 18 (map in Fig. 29); (b) mid-thermocline (10°C) isotherms in Milwaukee (MI) to Muskegon (MU) transects Nos. 4 to 27, from Mortimer (1968), renumbered in sequence (see Table 2) and here arranged in pairs to display changes. For each pair, the dashed line denotes the earlier member (often the immediate predecessor); and italic numbers record Δt, the hours elapsed between transect passages at 20, 40, 60, 80 and 100 km from MI. For transect pair 25/26, Δt was -5h; for pair 25/27 it was 72 h, because of a vessel change. Wind speeds (plotted on a knots squared scale, approximately proportional to stress,) are taken as the mid-points between maxima and minima, computed by FWPCA (1967) for successive two-hourly intervals.

features are in geostrophic near-equilibrium and are therefore accompanied by near-steady shore-parallel currents, awaits testing. Also to be investigated is whether the dish shape of the remainder of the thermocline illustrates Bennett's (1978) conjecture that wind impulses from random directions produce a net upwelling around the perimeter of a large stratified basin with a corresponding thermocline depression in the middle. The average thermocline

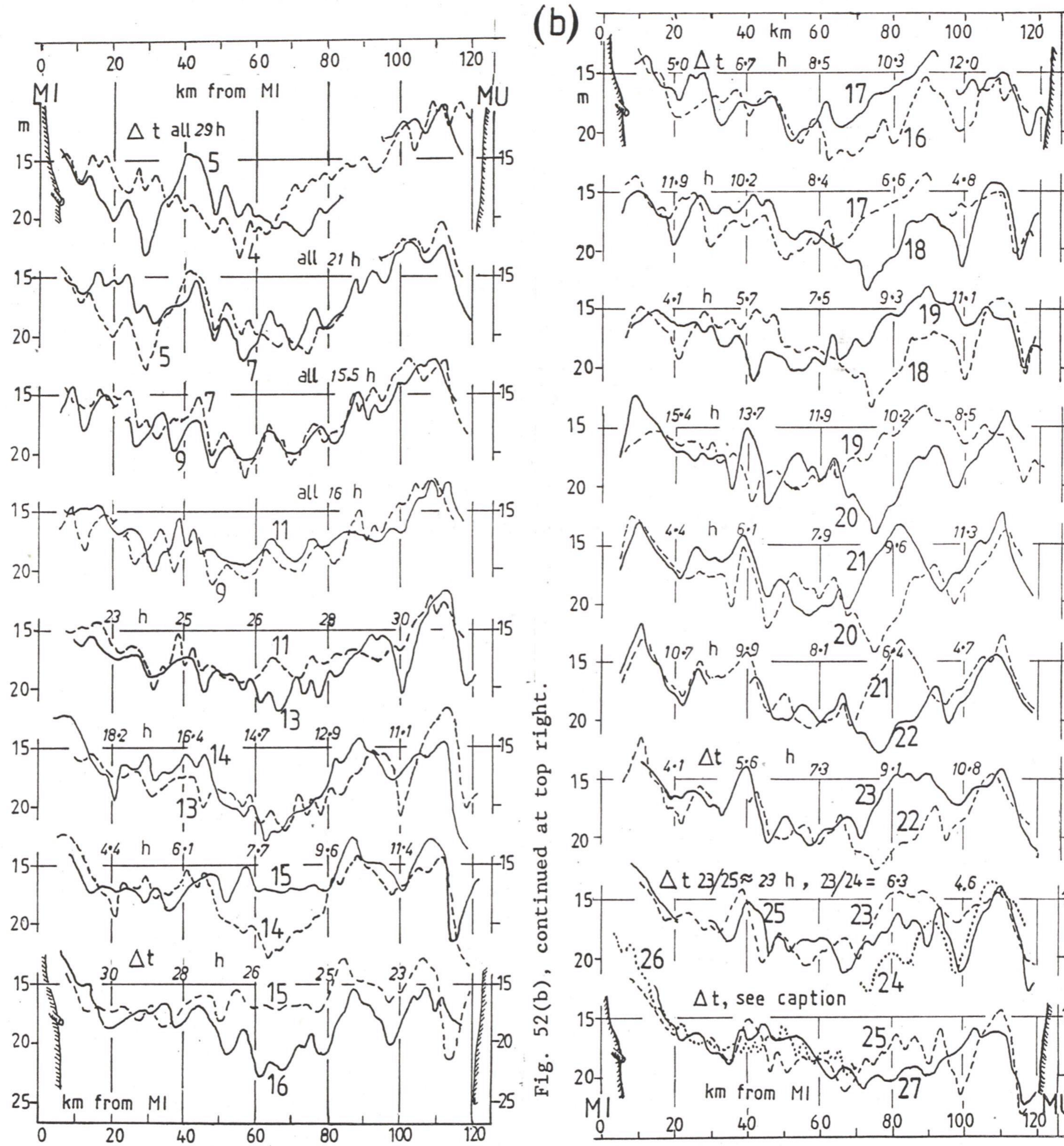


Fig. 52(b), continued at top right.

shape in Fig. 52 is decorated by smaller, non-persistent features. When the most prominent of those surge-like features were plotted on a distance/time diagram, no unambiguous patterns of motion appeared.

Also unexplained, and entirely unexpected in the absence of previous wind, are the appearances in Transects 17 to 22 (Fig. 52b) of: weak downwelling at MI; an amplitude increase in isotherm-depth fluctuations near MI, including a solitary and nearly stationary ridge near 40 km; and an apparently regular oscillation in thermocline depth in the 70 to 90 km region. The latter is illustrated (in Fig. 53) by the close fit of isotherm depth at 80 km to a 17 h sinusoid. That fit is remarkable, but unexplained, as is also the apparent absence of such an oscillation elsewhere along the transect. The between-transect elapsed times, Δt, suggest that this unexplained asymmetry cannot have been a consequence of the ferry timetable. Notable also -- but not discussed further here -- is the near-phase relationship in Fig. 53 between the oscillation at 80 km and the oscillation of thermocline depth observed at a mid-lake anchor station M2, continued in Fig. 30. The latter trace contains evidence of a superimposed oscillation at double frequency, $2f$, discussed earlier. Fig. 53 is consistent, either with a standing wave pattern with a probable node west of M2, or a co-incidental in-phase relationship between locally-generated inertial oscillations at M2 and 80 km.

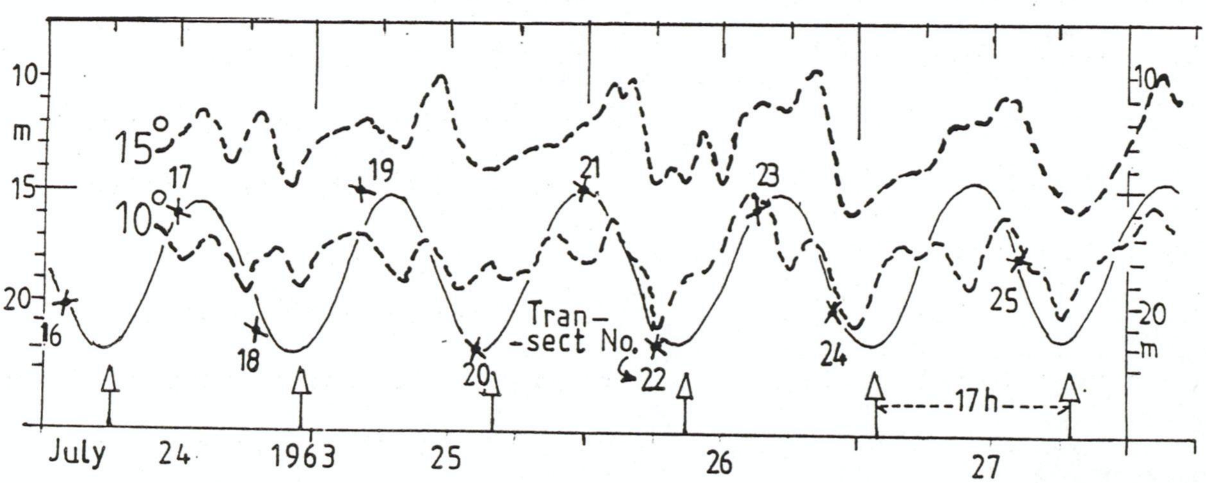


Fig. 53. Crosses indicate the depth of the 10° (mid-thermocline) isotherm as observed at 80 km from MI in Transects 16 to 25, 24 to 27 July 1963. A sinusoid of 17 h period and ±3.5 m amplitude is fitted to the crosses by eye. The dashed lines indicate depth of the 10° and 15° isotherms at a mid-transect anchor station, 64 km from MI.

With Fig. 54 we enter a windier period. There was a short, strong wind impulse toward NW on 2 August. During the preceding week, winds had been moderate and mainly northgoing; downwelling on the eastern shore had intensified; and the broad near-stationary ridge centered on 105 km from MI had become more prominent. The remainder of the thermocline had retained and deepened its dish shape; and upwelling at MI had strengthened. But the short wind pulse of 2 August had little effect on the overall shape of the thermocline; and the same was true for the moderate southgoing pulses on 7 and 9 August, except for an intensification of upwelling at MU (see Table 2). But the much stronger storm, veering from northgoing on 12 to southgoing on 13 August, had a dramatic effect, as shown by comparison of Transects 54, 54a (not completed because of rough seas) and 55. That storm brought about an overall 5 m deepening of the thermocline. It broadened the upwelling region off MU (compare Csanady, 1977) and reversed the previous upwelling at MI to downwelling close inshore (see Table 2). The amplitude of apparent surge features were also increased. Although their motion cannot be unambiguously demonstrated, progress of the presumed surge S in Transects 54(a), 55, and 56 is consistent with motion away from the MI downwelled region at a speed (1.5 to 2 km/h) comparable to that of the clearer examples of Lake Ontario internal surges illustrated in Fig. 38. Similar troughs also appear at the upwelled (MU) side of the section; but no progressive offshore motion was seen. This does not fit Fennel's (Fig. 51) model of Poincaré radiation from both sides; but it would be consistent with the conditions described and modeled for "full" upwelling along the N shore of Lake Ontario by Csanady (1977) and described on p. 82. It appears that, in Lake Michigan as in Lake Ontario, the internal surges emanate principally from the downwelled shore only, for the reasons advanced by Simons (1980) and quoted earlier on p. 84.

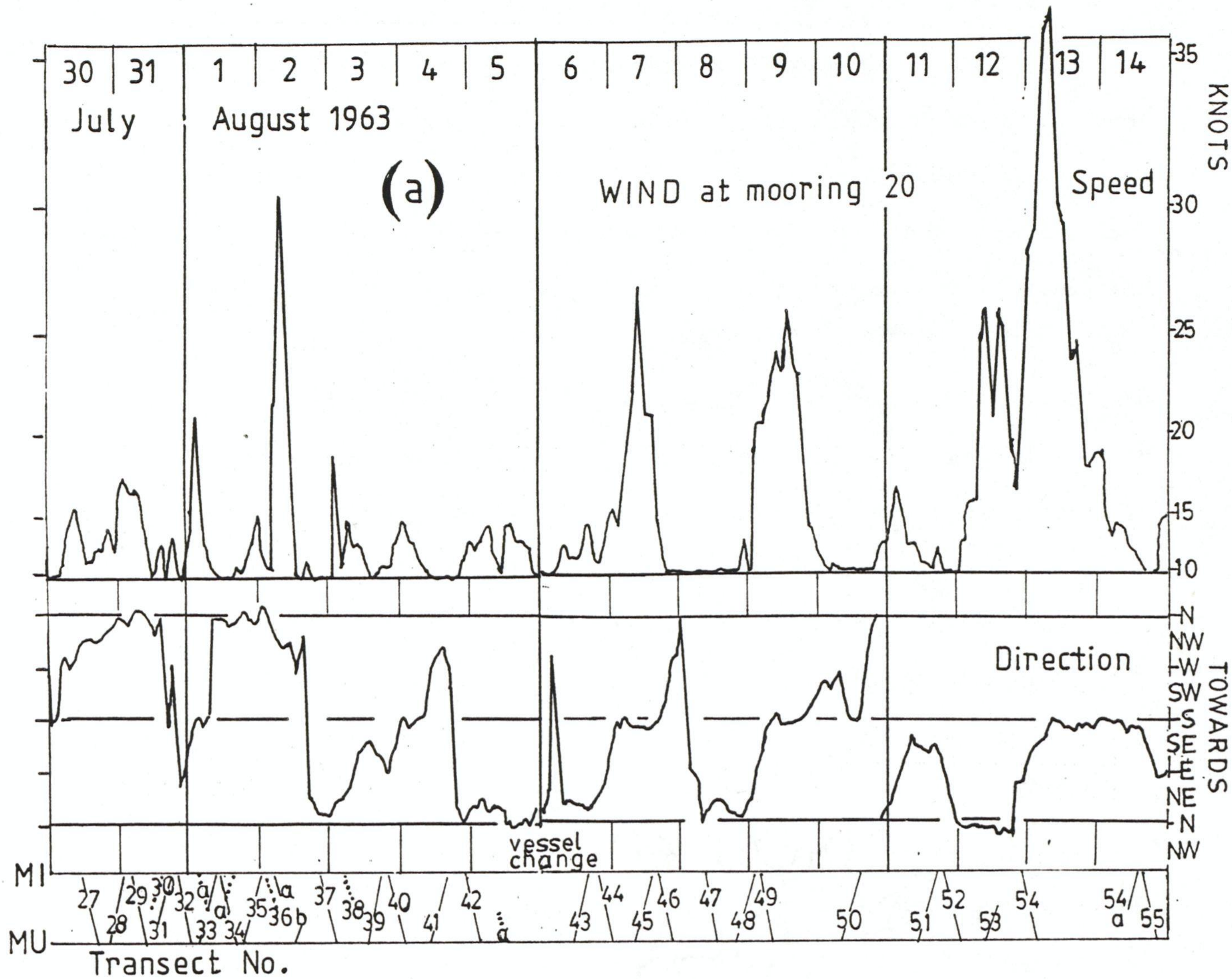


Fig. 54(a). Caption as for Fig. 52 but covering a windier interval, 30 July to 14 August 1963. The wind record (a) is from FWPCA mooring 20 and includes a storm on 13 Aug., in force between Transect Nos. 54 and 54(a).

During the interval of relative calm from noon on the 14th to noon on 16th August (Figs. 54 and 55) the thermocline undulations in the middle reach of the Section (in Transects 55 to 59) increased in amplitude and wavelength (Poincaré waves moving offshore?); but the direction and speed of their progress cannot be unambiguously determined from the transects alone. Meanwhile strong downwelling at MI and strong upwelling of MU were maintained.

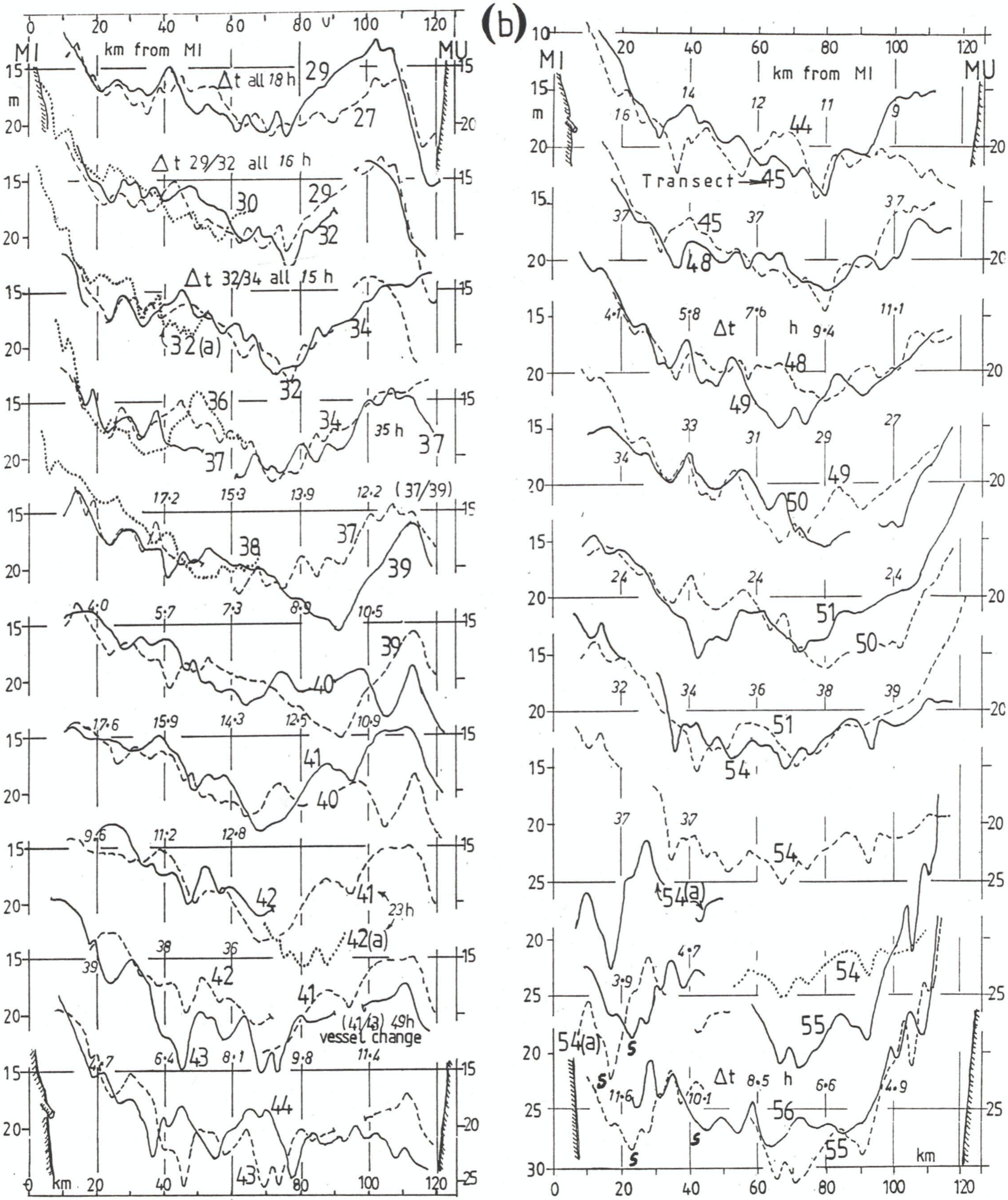
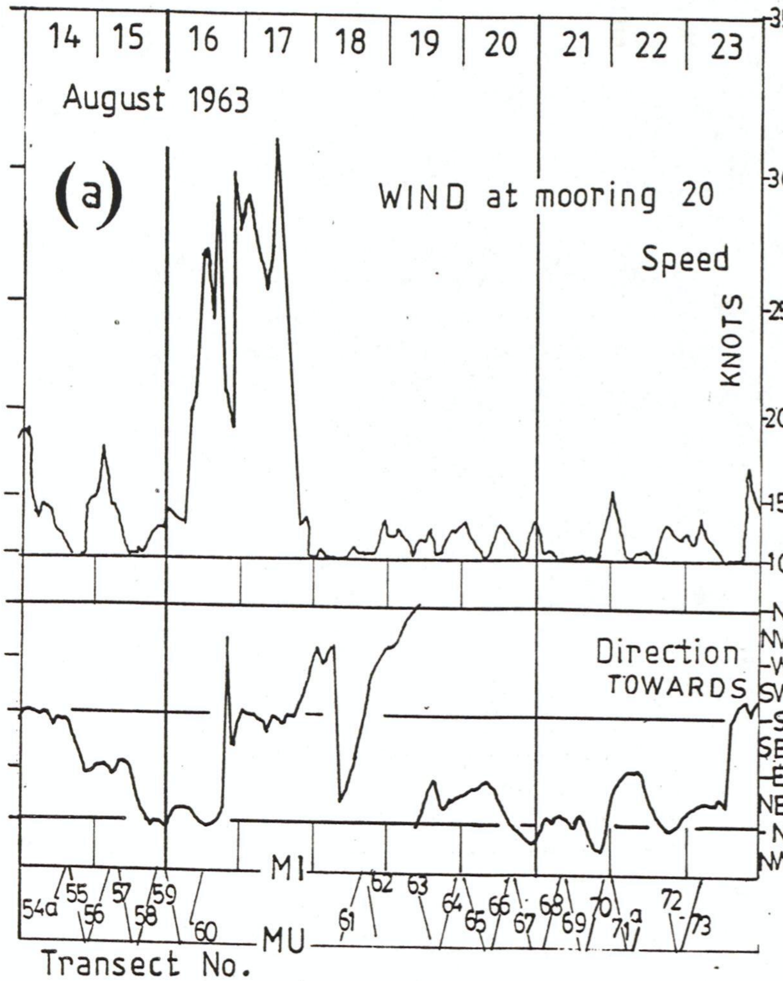


Fig. 54(b).



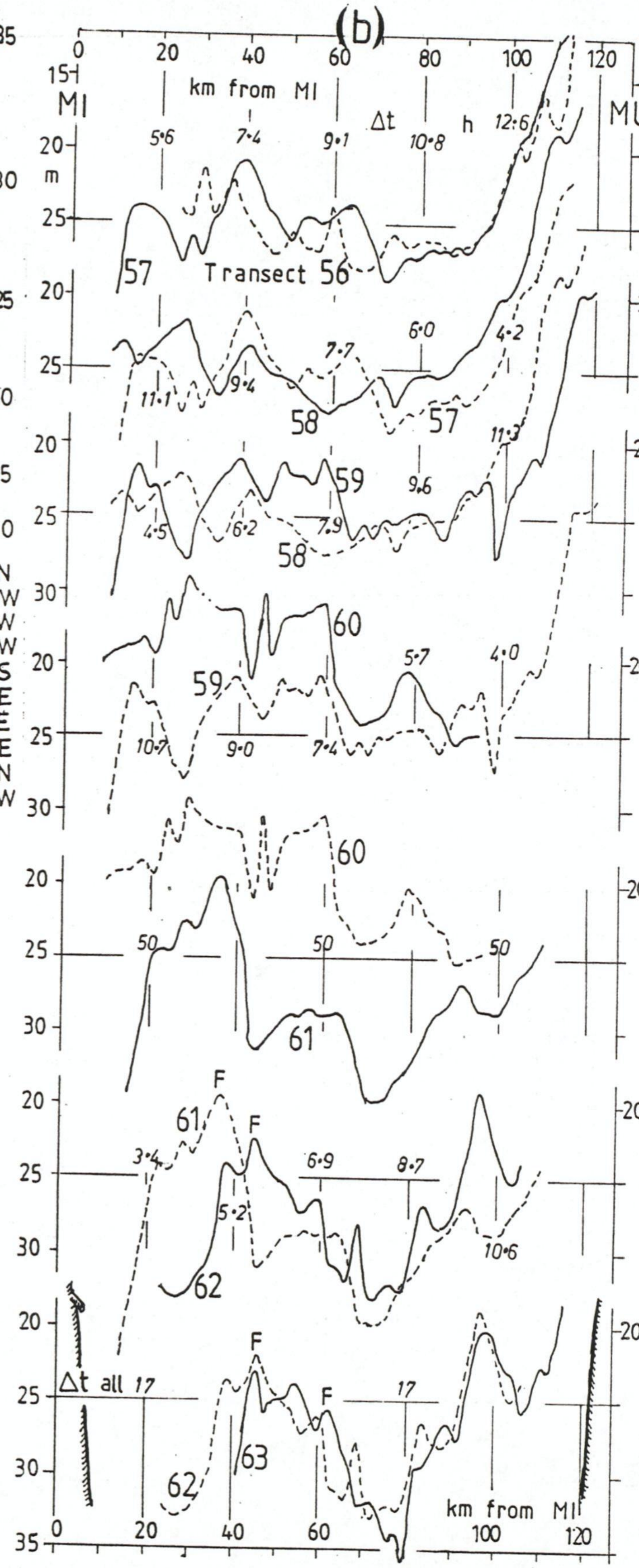


Fig. 55. Caption as for Fig. 52, but for 15 to 23 Aug. and Transect Nos. 56 to 72. The wind record (a) is from FWPCA mooring 20, the main event of which was the 16/17 Aug. storm, between Transects 60 and 61. Dashed lines indicate the earlier transect of each pair. In section (c), which is an expansion of the lower half of Fig. 33 with revised numbering of transects, the dashed-line transects are all E-going, as are the model dashed-line transects in (d), described in the text. Section (c) and models (d) and (e) are on the opposite page.

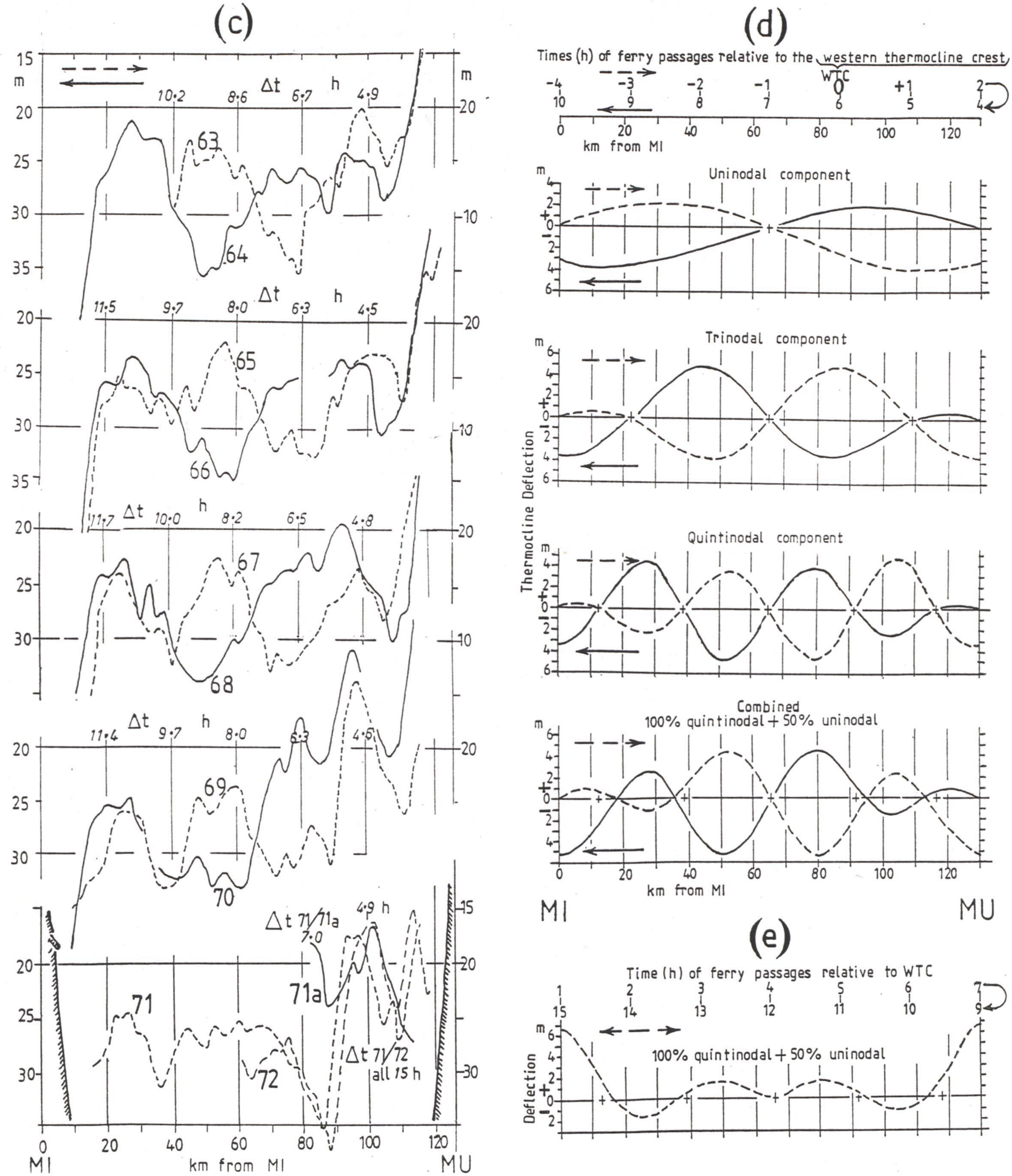


Fig. 55 c, and models d, and e (explained in the text).

The whole section was again strongly perturbed by a second major storm veering from north- to southgoing on 16/17 August. Comparison of the pre- and post-storm transects (60 and 61, Fig. 55b) again shows a large increase in thermocline depth and the development of large-amplitude undulations of the thermocline with "wavelengths" of order 20 km. The eastern ridge at about 97 km and trough at about 105 km from MI did not appear to move (cf. Csanady 1977 and compare 62, 63 and later transects); but there was apparent eastward (i.e. offshore) motion of the western ridge in transect sequence 61 to 63. But again, unambiguous identification of particular fronts (F, for example, in Fig. 55b) is not possible here, as it was in Lake Ontario.

By fortunate circumstance, the 16/17 August storm was followed by six days of near-calm; and the thirteen transects or part-transects run during that interval are presented in consecutive pairs in Fig. 55(b) and (c). The apparently moving (Poincaré?) wave patterns in Transects 61 to 63 (new numbering) merged into standing wave pattern revealed by Transects 63 to 71 in earlier Figs. 34.(lower half) and 35. The latter was a distance/time diagram designed to extract information from the post-storm transect pairs, 63/64, 65/66, 67/68 69/70, in a form not obscured by phase differences between the ferry's to-and-fro timetable and any internal waves which had been in progress. In fact, Fig. 35 reveals a standing wave pattern in the middle (40 to 90 km) part of the cross-section, with a distinct and persistent node at 66 km from MI, interpreted (Mortimer 1971) as an ensemble of odd-numbered standing Poincaré modes, with the quintinodal component dominant. That interpretation was questioned by Fennel (1989) and is re-examined here.

Where it first appears in Fig. 35, the thermocline wave crests at about 1300 h on 19 August on the western side of the mid-Lake node; and from then on the "western thermocline crest" (WTC) can be followed for four 16 h intervals.

The ferry departures 63, 65, 67 and 69 from MI were 3.8, 3.7, 4.4 and 4.4 h, respectively ahead of WTC. The respective turnaround intervals at MU were 2.6, 2.2, 2.6, and 2.3 h. That relative regular timetable makes possible a comparison between the ferry-observed thermocline depths in Fig. 55(c) with predictions of simplified models of thermocline seiches in a 130 km-wide channel, constructed (in Fig. 55d) as combinations of odd-numbered modes, each of 16 h period, ± 5 m elevation range and all in phase. In that model, the channel is scanned by a ferry which departs MI 4 h ahead of WTC, takes 6 h for the crossing and 2 h for the turnaround at MU. The results are illustrated in Fig. 55(d) for the uninodal, trinodal and quintinodal components separately and for a combination (100% quintinodal +50% uninodal) found by trial and error to provide a good fit with the observed thermocline structure in Fig. 55(c). (Addition of a smaller trinodal contribution improves the fit only marginally; and it is assumed that contributions of even-numbered modes are negligible.) In each transect pair, the E-going member is shown as a dashed line.

The comparison demonstrates that the quintinodal/uninodal combination satisfactorily reproduces the salient features of thermocline geometry in the 20 to 90 km region of the transect, but that strong and persistent downwelling off MI and upwelling off MU obscures the signal from the nearshore portions of the presumed multinodal seiche.

At other MI departure times (relative to WTC) the correspondence between observed and model transects is less close; and it was therefore a fortunate coincidence that the actual ferry timetable was nearly optimal to reveal the seiche which followed the 17 August storm. For example, a MI departure 1 h after WTC, in a 100% quintinodal +50% uninodal model (Fig. 55e) would have suppressed the mid-Lake seiche signal almost entirely. (In that model case, the thermocline elevation traces on the E-going and W-going transects are identical.)

In Figs. 55(d) and (e) the models are all symmetrical about the central node. Thus that model cannot explain the cross-basin asymmetry of the oscillation illustrated in Fig. 53. What happens when the modal components are not in phase remains to be explored; and it should not be forgotten that, in reality in Lake Michigan, the periods of the first three modes will be nearer 17, 16, and 15 h, respectively (Mortimer 1971). Therefore, even if they start off in phase, they will shift further out of phase as the oscillation progresses. Also, in the foregoing discussion, the experimental data have restricted us to two-dimensional views and models of central basin cross-sections. Although those models permit us to ignore the effects of basin ends, if sufficiently distant, Fig. 56 reminds us that responses in real channels and basins are three-dimensional and must eventually be modified by end effects. Therefore, the questions posed and the methods developed for future research in these fascinating inland seas will have to take that into account.

Retrospect and Prospect

The foregoing review -- one of the chapters in physical limnology -- has been a story of the interplay between observation and theory, spurred by the invention of a procession of new tools: reversing thermometers at the beginning of the century followed by the bathythermograph in 1942, the thermistor chain in 1950, remote thermometry by satellites, and acoustic Doppler current profilers in recent years. As in oceanography*, observations have, again and again, preceded and prompted theoretical modeling (Simons, 1980). This iterative, two-legged mode of progress will continue and accelerate during the next century, with the advent of instruments, some

*Stommel (1989) "The chief source of ideas in oceanography comes, I think, from observations...Most theories are about observations that have already been made."

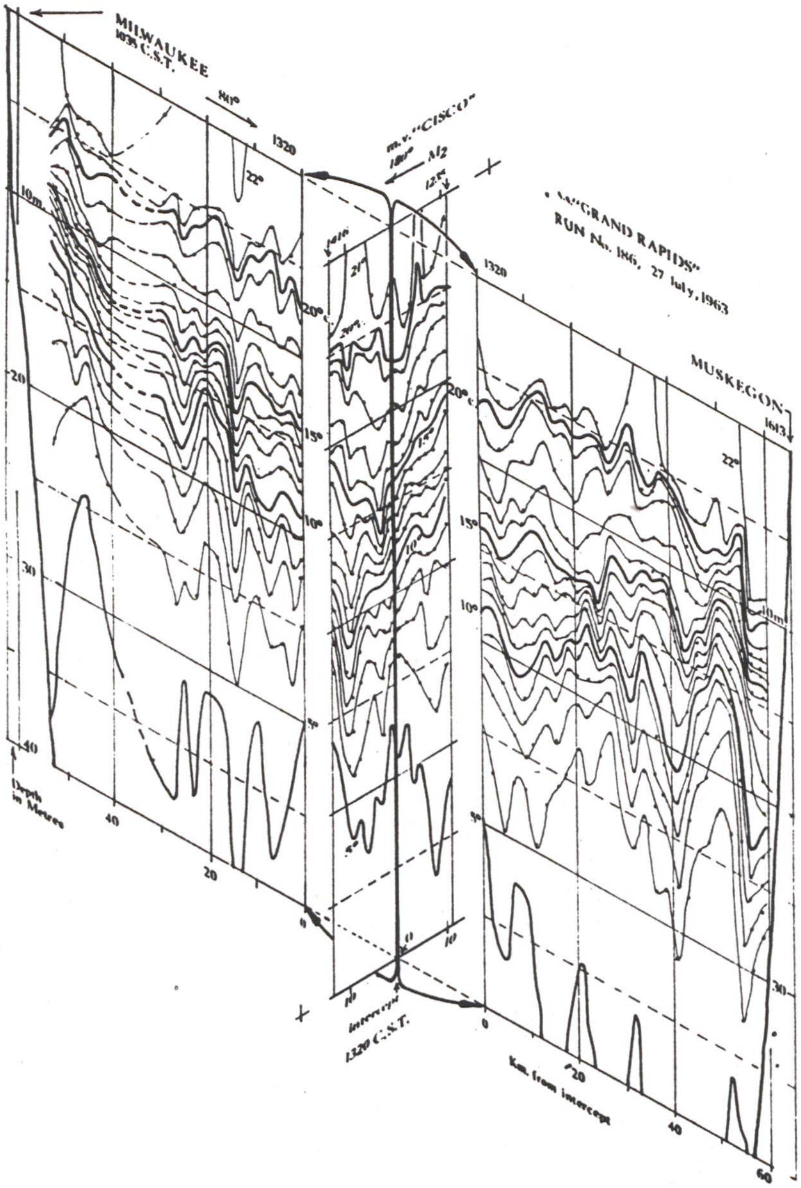


Fig. 56. Lake Michigan, 27 July 1963: Three-dimensional structure of isotherm distribution, illustrated by a W to E ferry Transect 25 (new numbering) and a short intersecting N to S transect near mid-Lake (station M2, from Mortimer, 1968).

satellite-interrogated, better able to reveal the forced and free responses of the system over relevant scales of time and space.

This review was focused principally on the free responses; but the dynamics of forcing and friction are also of fundamental importance (Heaps 1966, Hamblin 1978). For example, attention should be directed to those many lakes in which wind-forcing typically proceeds from one event to the next, with little intervening opportunity for free oscillations? Illuminating here are the upwelling and mixing studies of Monismith (1986).

Interpretation and planning of the measurement campaigns will be notably aided by the evolution of computers with greatly expanded capabilities of near-realistic simulation. In particular, we may expect the emergence of a clearer understanding of geostrophic readjustments and associated generation of internal waves and their vertical and horizontal modal structure. An important area of study will be the nonlinear dynamics of those waves in large basins, in which rotation and geostrophic readjustment effects are important.

It will be necessary, not only to distinguish between the ubiquitous response of inertia-dominated motions offshore and the nearshore responses -- Poincaré and Kelvin waves -- but also to understand the interactions of those components, including interactions with topographic Rossby waves. In all of this, the distinction between forced and free responses will require clarification, with the aim of achieving a deeper understanding of air/water interactions, helped by combined aerodynamic and hydrodynamic modeling. Internal waves, their generation and decay, promise to fascinate researchers in the next century, as they did in the last.

Synopsis

The sub-discipline of physical limnology contains four main cross-linked chapters. These are (1) properties and formation of water masses, (2) mixing and diffusion, (3) hydro-optics, and (4) oscillatory responses to forcing by wind. As part of the 4th chapter, this review covers a century of discovery and research related to long internal (i.e. sub-surface) gravity waves. It is a pictorial, non-mathematical, and somewhat autobiographical account from one who has fished in these particular waters for fifty years. It is an account which pays particular attention to "firsts".

The first milestone of the century was Murray's (1888) demonstration of isotherm-tilting forced by wind stress on stratified lakes and marine fjords in Scotland. Oceanography in the post-Challenger era was then languishing for lack of funds; and Scottish investigators (including Murray and Chrystal) turned from oceans to lakes. Starting with measurements in Loch Ness, the 1903-1913 decade encompassed, first, the discovery of the internal seiche as a standing wave response to wind-induced displacement of stratified layers, and second, the modification of Chrystal's seiche equations (1)* to predict periodicity and modal structure in real basins. The chief architect of that achievement was Wedderburn.

Three decades of relative neglect, by limnologists, of Wedderburn's discoveries and their widespread influence on transport and mixing came to an end when I demonstrated, not only large Wedderburn-type thermocline oscillations in Windermere, but also oscillations of lower frequency and larger amplitude in sub-thermocline layers. That drew attention to the importance of <u>vertical structure</u> (2). A comparison of extensive records of wind and water temperature fluctuations (recorded by moored thermistor chains 1950/51) confirmed Wedderburn's conjecture that internal seiches are <u>resonant responses</u>, initially forced by wind. Heaps model (3) of the forcing phase and of its transition to the free response satisfactorily reproduced the observed motions.

The thermistor chains, transferred to Loch Ness in 1952 to explore the two-dimensional, rotation-related structure of the internal seiche, also revealed conspicuous nonlinear (steep-fronted) features. That asymmetry was later studied by Thorpe and modeled as an <u>internal surge</u> (4). His observations with a profiling current meter (1977) revealed that the surge, when first formed, was accompanied by a short train of solitary internal waves (<u>solitons</u>), reminiscent of waves seen at the crest of internal surges in Seneca Lake (5). Such nonlinear features continue to be objects of recent research (6). They are to be anticipated wherever the seiche amplitude is large. The degree of nonlinear modification of the seiche ranges from

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*Numbers in this form indicate the principal interpretative models, referred to in footnote entries at the end of this synopsis.

moderate (when the surge is reflected at basin ends) to extreme, when a one-way-traveling surge becomes the sole feature (5,7).

The influence of the earth's rotation on lake motions was first revealed in 1950 in Lac Léman as a cyclonically-progressing internal seiche/surge. Similar responses in other lakes (8,9) and recent observations in Léman (10) have been modeled as modified internal Kelvin waves.

Records from nearshore water intakes around a much larger basin, Lake Michigan, also revealed occasional cyclonic progression of temperature waves traveling at internal Kelvin wave speeds and set in motion after strong wind had brought about extensive downwelling along one shore and upwelling along the other. Those waves displayed nonlinear features (11).

Occasionally the Lake Michigan intake records also signaled oscillations of the nearshore thermocline at a frequency near the local inertial frequency, f. Sometimes the frequency of those inertial waves was blue-shifted to as much as 15% above f. That unexpected blue-shift -- now explicable in part by modern theories of coast-modified inertial motion (12,13) and by the later-described geostrophic readjustment process (20, 1982) -- led me to borrow from tidal theory a two-layered, rotating wide-channel model (14) in which the thermocline responses to wind impulses included: (a) shore-trapped low-frequency Kelvin waves; (b) offshore inertial motion proper (Sverdrup waves here called prototype Poincaré waves); and (c) offshore-propagating Poincaré waves of near-inertial frequency which, if given sufficient undisturbed time, would produce a cross-basin standing wave pattern of transverse internal seiches.

Tested by extensive campaigns of current and temperature measurement in Lake Michigan (1963/64) and in Lake Ontario (IFYGL, 1972), the Kelvin-Poincaré rotating channel model (14, 17, 19) has yielded insights into several observed phenomena, but has proved to be oversimplified for others. From today's vantage point, four separate but interacting responses to applied wind stress can be recognized. They are here denoted as follows:

Response 1, locally-generated offshore inertial motion (here dubbed "waltzing"), is the immediate ubiquitous and intermittent response in offshore regions to change in wind stress (13) or other sudden changes. In its "pure" form it is solely expressed as currents rotating clockwise at inertial frequency f; but it often appears "contaminated" by thermocline-depth oscillations at main frequency near f, but with a small unexplained $2f$ component, first discovered in Lake Michigan. In Lake Erie, the only other basin in which extensive measurements have been made, inertial current rotations dominated with maximum amplitude above the seasonal thermocline. That response has been modeled (23) in terms of changes in layer structure. Transverse seiching has not yet been seen in that lake.

If wind stress is sustained, the offshore Response 1 includes Ekman drift of the upper layer to the right of the wind direction.

Response 2. The Ekman drift, approaching shore, brings about nearshore downwelling along the shore lying to the right of the wind direction, accompanied by upwelling along the opposite shore. If the upwelling is strong enough to bring the thermocline to the surface, continued wind stress can move the thermocline front further offshore (18, 1977).

Response 3, a geostrophic readjustment to nearshore up/downwelling, comes into play when wind stress is removed. Part of the energy thus released stays nearshore as geostrophic currents and shore-trapped Kelvin waves; and part radiates offshore as a dispersive ensemble of Poincaré waves, covering a range of wavelengths and frequencies. The shortest members (of frequency $>f$) lead the ensemble, followed in sequence by a trail of slower waves of increasing length and decreasing frequency (tending to the limit f). Those waves, with their corresponding thermocline undulations, advance offshore into regions previously dominated by Response 1; and this explains some of the amplitude and frequency changes observed. If large enough, the leading waves form internal surge fronts (16, 1978). The fronts, so far observed, propagate exclusively from downwelled shores (16, 1980).

In Lake Michigan the predicted shore-trapped Kelvin-type responses were seen, but only as a response to upwelling along the E shore, perhaps because a whole-basin response cannot be completed before destruction by new wind disturbance. In Lake Ontario, Kelvin responses were also seen, but were complicated by interaction with a barotropic topographic Rossby wave (15,22).

Response 4: The occasional generation of cross-basin standing waves (internal seiches) was demonstrated by repeated scans of cross-basin temperature structure in Lakes Michigan and Ontario. Response 3, it is believed, is too slow (21) to produce such patterns, except perhaps on rare occasions. But, in Lake Michigan, a standing wave pattern was developed within three days of a storm; and this is attributed to a temporary distortion of the equilibrium shape of the thermocline. The initial shape of that distortion determines the conspicuous contribution of particular odd-numbered higher modes to the ensuing seiche pattern. No standing wave pattern has yet been detected in Lake Erie (23).

The above four responses should be viewed as components of the combined response of the basin to change in wind stress. One component may dominate in one region and at one time, another at another time; some may be expressed in sequence; and some may be seen more frequently than others; but they remain combined and interacting.

The review closes with a list of questions unresolved for large lakes: inertial waves (generation, intermittency, blue-shift, and nonlinearity); internal surges and fronts (generation, propagation, modification or reflec tion at basin boundaries, and possible interactions with Rossby waves); cross-basin internal seiches and the apparently rare conditions under which they become conspicuous.

Bracketed numbers denote the principal interpretative models referred to above (see bibliography): (1) Wedderburn, 1912; (2) Longuet-Higgins in Mortimer, 1952; (3) Heaps, 1966; (4) Thorpe, 1972; (5) Hunkins and Fliegel, 1973; (6) Grimshaw, 1978; (7) Farmer, 1978; (8) Kanari, 1975; (9) Hamblin, 1978; (10) Beckers, 1989; (11) Bennett, 1973; (12) Millot and Crépon, 1980; (13) Kundu et al., 1983; (14) Mortimer, 1963; (15) Csanady, 1976; (16) Simons, 1978, 1980; (17) Mortimer, 1971; (18) Csanady, 1973, 1977; (19) Lee and Mysak, 1979; (20) Gill, 1976, 1982; (21) Fennel, 1988; (22) Simons and Schertzer, 1989 (23) Boyce and Chiocchio, 1987.

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