Thermal Mediation in a Natural Littoral Wetland: Measurements and Modeling

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Thermal mediation in a natural littoral wetland: Measurements and modeling

Hrund Ó. Andradóttir and Heidi M. Nepf
Ralph M. Parsons Laboratory, Department of Civil and Environmental Engineering
Massachusetts Institute of Technology, Cambridge

Abstract. As a river flows through shallow littoral regions such as wetlands, forebays, and side arms, the temperature of the water is modified through atmospheric heat exchange. This process, which we call thermal mediation, can control the initial fate of river-borne nutrient and contaminant fluxes within a lake or reservoir. This paper presents temperature observations that demonstrate the occurrence of thermal mediation and directly support the theoretical results derived by Andradóttir and Nepf [2000]. The measurements show that the wetland warms the river inflow by approximately 1–3°C during summer and fall nonstorm conditions. Less thermal mediation occurs during storms, both because the residence time is significantly reduced and because the wetland circulation shifts from laterally well mixed (low flows) to short-circuiting (storms). The dead-zone model can simulate both these regimes and the transition between the regimes and is therefore a good choice for wetland modeling.

1. Introduction

Wetlands can play an important role in improving downstream water quality. Numerous mass balance studies have shown that suspended sediments, nutrients, metals, and anthropogenic chemicals are efficiently removed in natural and constructed wetlands through a variety of sink mechanisms, such as bacterial conversion, sorption, sedimentation, natural decay, volatilization, and chemical reactions [Tchobanoglous, 1993]. Yet other studies have shown that wetlands may also act as a temporal source of nutrients and pollutants as they release stored materials [Mitsch and Gosselink, 1993, p. 157]. To date, extensive research has been conducted to understand the chemical, biological, and physical processes underlying the sink/source potential of these complex ecosystems. However, one process that so far has received little attention is thermal mediation, i.e., the temperature modification of the water flowing through the wetland. For littoral wetlands the outflow temperature determines the lake intrusion depth which, in turn, affects the initial fate of nutrients/contaminants in the lake, as well as the residence time and mixing dynamics within the lake. Using a dead-zone model, Andradóttir and Nepf [2000] showed that wetland thermal mediation can profoundly affect intrusion depth. For example, this process can shift the timescale of lake intrusion depth variability from predominantly seasonal to synoptic and diurnal. Moreover, wetlands can prolong surface intrusions during summer, which can lead to increased human exposure to river-borne contaminants. Similarly, the increased nutrient supply to the epilimnion during the growing season can accelerate lake eutrophication [Carmack et al., 1986; Metropolitan Council, 1997]. On the other hand, more surface intrusions lead to quicker flushing, potentially reducing the long-term deposits of nutrients and contaminants in the lake. Wetland thermal mediation can therefore have complex short- and long-term effects on lake water quality.

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This paper presents detailed observations of thermal mediation in a natural wetland that receives unregulated river inflow. The major contributions of the paper are to (1) demonstrate through field observation that wetland thermal mediation occurs in a small watershed and how it is affected by flow conditions, (2) compare wetland thermal structure and circulation during low flows and storms, (3) validate the use of the dead-zone model in wetlands, both by confirming the analytical dead-zone model results derived by Andradóttir and Nepf [2000] and by showing that the model simulates well wetland thermal behavior during variable flow conditions, and (4) demonstrate the effect of wetland thermal mediation on lake intrusion dynamics. This paper is the observational counterpart to the theory presented by Andradóttir and Nepf [2000].

2. Theoretical Background

Thermal mediation is a well-known process in cooling ponds, in which atmospheric heat exchange is exploited to attenuate the waste heat from power plants. Andradóttir and Nepf [2000] showed that thermal mediation is also an important process in littoral wetlands, when the river inflow follows a different seasonal temperature cycle than the wetland water. This condition is met in small or forested watersheds, where the river follows a damped seasonal cycle because of groundwater recharge [Gu et al., 1996] and/or sun shading [Sinokrot and Stefan, 1993]. Thermal mediation is less significant in larger watersheds, in which the river has enough time to equilibrate with the atmosphere and in which sun shading and wind sheltering are less prominent.

While the inflow condition sets the potential for wetland thermal mediation, the degree of thermal mediation actually occurring is governed by the residence time distribution (RTD) within the wetland and the thermal inertia of the water $t_{\text{heat}}$. The thermal inertia represents the heating timescale of the water column and is defined as the ratio of the water depth $H$, and the surface heat transfer coefficient $K$. The ratio of the nominal residence time $t$ to the thermal inertia, also called thermal capacity,
Table 1. Vegetation Drag Estimates in the Upper Forebay

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*Bed drag decreases significantly at high Reynolds numbers, $Re$, due to protraction of vegetation stem.*

*Estimates are adapted from (1) Kadlec and Knight [1996, p. 201]; (2) Chen [1976]; (3) Dunn et al. [1996, p. 54]; and (4) Andradóttir [1997, pp. 69, 74].

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is a major factor controlling how much thermal mediation occurs. For “ideal” plug flow, $r = 0$ produces no thermal mediation, whereas $r > 3$ produces over 90% of the maximum thermal mediation possible in the system [e.g., Jirka and Watanabe, 1980]. Wetland flow regimes, however, are rarely ideal [Kadlec, 1994], and water circulation, which governs both the skewness and variance of the RTD, will generally modify how efficiently the wetland mediates the water temperatures.

Wetland water circulation depends on three meteorological forces: wind, which scales on the mean wind shear stress and wetland length as $\tau_{w} L$; river buoyancy, which scales on the density anomaly between the inflow and wetland water and water depth as $\Delta \rho g H^2$; and river momentum per unit width, $\rho U_0^2 H$, where $U_0$ is the inflow velocity and $\rho$ is the water density. In addition, the water circulation is influenced by vegetation drag, which scales on the friction factor as $\frac{1}{2} \rho C_f U_0^2 L$. The relative importance of vegetation drag, wind, and river buoyancy to the river inertia can be summarized by the ratio of scales:

Vegetation drag parameter [DePaoli, 1999]

$$\Lambda = \frac{C_f L}{2 U_0^2}.$$  

Wind parameter

$$\Omega = \frac{\tau_{w} L}{\rho U_0^2 H},$$

(Internal Froude number)$^{-2}$

$$Fi^{-2} = \frac{\Delta \rho g H^{2}}{U_0^2}.$$  

Notice that since both $\Omega$ and $Fi^{-2} \propto U_0^{-2}$, then the influence of wind and buoyancy on water circulation changes with flow conditions. Similarly, $\Lambda$ is indirectly related to $U_0$, through the friction factor $C_f$, which depends strongly on water depth and velocity, as well as vegetation type and density [Petryk and Bosmajian III, 1975; Chen, 1976; Shih and Rahi, 1982]. In particular, $\Lambda$ can decrease drastically during storms, as a result of increased water depths and stem pronation. This is shown by the characteristic values of $C_f$ given in Table 1. Water circulation is therefore expected to change between low flows and storms [DePaoli, 1999; Andradóttir, 1997]: During low flows, vegetation drag, wind, and buoyancy dominate the water circulation ($\Lambda$, $\Omega$, and $Fi^{-2} \gg 1$). Uniform vegetation drag will help distribute the river flow over the width of the wetland [Wu and Tsanis, 1994], and wind will enhance lateral and vertical mixing. Consequently, the wetland behaves as a partially well mixed reactor where the RTD has a large variance around the mean nominal residence time $i$ (Figure 1a) [Kadlec, 1994]. During high flows, however, the circulation becomes inertia- or jet-dominated ($\Lambda$, $\Omega$, and $Fi^{-2} < 1$), and substantial short circuiting occurs; that is, a large portion of the river flow exits the wetland in much less time than the nominal residence time $i$ (Figure 1b). Both these regimes, and especially the short-circuiting regime, produce less thermal mediation than the ideal plug flow [Andradóttir and Nepf, 2000].

3. Methods

3.1. Site Description

The Aberjona watershed is a medium-sized watershed (65 km² surface area) located in suburban Boston, Massachusetts. Because of long term industrial activities the watershed is heavily contaminated with heavy metals such as arsenic and lead which are routinely transported downstream by the Aberjona River until they are finally deposited in the Upper Mystic Lake [Solo-Gabriele, 1995; Aurilio et al., 1994]. Before entering the lake, the river flows through two littoral wetlands, first the larger upper forebay and then the smaller lower forebay (Figure 2a). Both wetlands are vegetated with water lilies and submerged coontail, and their mean water depth ranges between 1.3 and 2.0 m depending upon season. In comparison, the lake epilimnion is approximately 5 m deep, and the thermocline is 5 m wide during the summer.

In this paper, we focus predominantly on the thermal mediation occurring in the upper forebay shown on Figure 2b. This wetland provides most of the thermal mediation in this system, first, because the river discharges into it, second, because its larger size (thus longer residence time) allows more thermal alterations to occur, and, third, because it receives limited return flow from the lake. In contrast, the lower forebay exchanges significantly with the lake through the 60 m wide channel connecting this wetland to the lake as illustrated on Figure 2a. The lower forebay can thus be considered to be an
extension of the lake surface layer, and the outflow temperature of the upper forebay can be taken as representative of the temperature of the lake inflow.

3.2. Field Observations

Water temperatures were monitored every 5 min at different locations in the Upper Mystic Lake system from July to November in 1997 and 1998. Temperature measurements were made 10–20 cm below the surface and 10–20 cm above the bed at each site in the upper forebay shown on Figure 2b using Onset temperature loggers with 0.2°C resolution. In 1997 the temperature loggers at the outlet of the wetland were repeatedly stolen, so the following year, the measurements were repeated with additional loggers deployed at the outlet and in the shallow vegetated zone. Lake water temperatures were measured in 1997 at 1–1.5 m depth intervals within the surface mixed layer and thermocline (1.8–10.9 m depth) at the locations shown on Figure 2a using thermistor chains with 0.1°C resolution [Fricker and Nepf, 2000]. Lake intrusion depth with and without the wetlands was estimated by matching the average of the near-surface and near-bed wetland channel temperature and the river bed temperature, respectively, to the lake thermistor data. Surface intrusions within the top 2 m in the lake were not resolved because no thermistor was located there. Changes in suspended sediment levels did not significantly affect the water density and were neglected in the calculations.

Wind speed and direction were measured 10 m above ground at the southern end of the lake at 10 min intervals in 1997 and 1998 (Figure 2a). Hourly Aberjona River flow was measured by the U.S. Geological Survey approximately 800 m upstream of the inlet to the upper forebay. In addition, air temperature, relative humidity, and solar radiation were monitored every 5 min during 1997 (Figure 2a). Cloud cover was measured at 3 hour intervals at the Boston Logan Airport located 15 km east of the study site.

3.3. Dead-Zone Model Application

3.3.1. Model formulation. The dead-zone model was described in detail by Andradottir and Nepf [2000]. In short, the wetland area is divided into two zones, a flow zone (or channel) with cross-sectional area \( A_c \) and a stationary dead zone with cross-sectional area \( A_d \). As shown on Figure 2b, while the river traverses the wetland channel (shaded area), it exchanges continuously with two stationary dead zones on either side of the channel. Note that the southern dead zone is smaller, shallower, and more vegetated than the northern dead zone. The circulation regime in the wetland is described by the areal ratio, \( q = A_c / (A_c + A_d) \), and the nondimensional lateral exchange coefficient \( \alpha^* = \Delta Q / Q_r \), where \( \Delta Q \) is the total lateral exchange rate between the channel and dead zones and \( Q_r \) is the inflow flow rate. Neglecting longitudinal dispersion, the simplified governing equations for the depth-averaged water temperature in the channel, \( T_c(x, t) \), and dead zone, \( T_d(x, t) \), are

\[
\frac{\partial T_c}{\partial t} + u \frac{\partial T_c}{\partial x} = \alpha^* \frac{u}{L} \left( T_d - T_c \right) + \frac{\phi}{\rho C_p H_c}, \quad (5)
\]

\[
\frac{\partial T_d}{\partial t} = - \alpha^* \frac{u}{L} \left( 1 - q \right) \left( T_d - T_c \right) + \frac{\phi}{\rho C_p H_d}. \quad (6)
\]

The boundary conditions at the inlet \( (x = 0) \) and outlet \( (x = L) \) are

\[
\frac{\partial T_c}{\partial x} \bigg|_{x=L} = 0. \quad (8)
\]

Here \( u = Q_r / A_c \) is the average speed in the channel, \( L \) is the length of the wetland, \( T_0 \) is the wetland inflow temperature, \( \phi \) is the net surface heat flux per surface area, and \( C_p \) is the specific heat of water.

The parameters \( q \) and \( \alpha^* \) are site-dependent and are typically evaluated by conducting dye experiments [e.g., Bencala and Walters, 1983]. For preliminary design and assessment, however, it is useful to derive expressions based upon the inflow and site geometry alone. First, consider the river-dominated circulation regime, when the river flows through the wetland like a jet (Figure 1b). The laboratory experiments of Chu and Baines [1989] suggest that the jet spreads linearly from its initial width \( W_0 \), with spreading angle \( \beta \approx 0.1 \), such that the channel width is \( W_c = W_0 + 2\beta x \). This yields the mean wetland areal ratio of

\[
q = \left( \frac{W_0 + \beta L}{W} \right) \frac{H_c}{H} \approx \frac{\beta L}{W}. \quad (9)
\]
In water depth as $A_{H_2}$; and river momentum per unit width, wetland length as $wL$; river buoyancy, which scales on the buoyancy to the river inertia can be summarized by the ratio of forces: wind, which scales on the mean wind shear stress and vegetation drag parameter $[DePoli, 1999]$, and wind will enhance lateral and vertical vegetation drag, wind, and buoyancy dominate the water circulation changes with flow conditions. Similarly, $A$ is indirectly related to $U_0$ through the friction factor $C_p$, which depends strongly on water depth and velocity, as well as vegetation type and density $[Petlyk and Bosmajian III, 1975; Chen, 1976; Shih and Rahi, 1982]$. In particular, $A$ can decrease drastically during storms, as a result of increased water depths and stem pronation. This is shown by the characteristic values of $C_p$ given in Table 1. Water circulation is therefore expected to change between low flows and storms $[DePoli, 1999; Andrad6ttir, 1997]$: During low flows, vegetation drag, wind, and buoyancy dominate the water circulation $([A, \Omega, \text{and } F_i^{-2} < 1])$, and substantial short circuiting occurs; that is, a large portion of the river flow exits the wetland in much less time than the nominal residence time $\bar{t}$ (Figure 1b). Both these regimes, and especially the short-circuiting regime, produce less thermal mediation than the ideal plug flow $[Andrad6ttir and NePF, 2000]$. 

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![Figure 1. Bimodal circulation in free water surface wetlands. (a) Under low flows, drag, wind, and/or buoyancy dominate the circulation $([A, \Omega, \text{and } F_i^{-2} >> 1])$; the wetland is laterally well mixed producing a relatively symmetric residence time distribution (RTD) around the nominal residence time $\bar{t}$. (b) During high flows the circulation is river-dominated $([A, \Omega, \text{and } F_i^{-2} < 1])$; short circuiting occurs producing a skewed RTD with much of the flow exiting the wetland in much shorter time than $\bar{t}$.](image-url)

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(7)

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$$

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$$
q = \left( \frac{W_0 + \beta L}{W} \right) \frac{H_c}{H} \approx \beta \frac{L}{W}.
$$

(9)
Figure 4. Water temperatures in the wetland channel $T_c$ and the northern and southern dead zones, $T_\alpha$ and $T_\beta$, respectively, during (a) early and (b) late fall 1998. The bold line depicts the near-surface temperatures, and the light lines depict near-bed temperatures. Cross-hatched areas denote large storm occurrences when $Q_r > 1$ m$^3$/s. Wetland stratification and diurnal surface temperature fluctuations decrease progressing into the fall because of convective cooling and increasing winds. Observation resolution is ±0.2°C.

10 to 1000. Under these conditions the wetland can exhibit a complex, transient, three-dimensional flow behavior. Since $\Lambda$ ranges from about 30 to 300 (Table 1), vegetation drag is also expected to play an important role in distributing the river laterally [DePauw, 1999]. Dye experiments show that the integrated effect of all these processes over the residence time (>3 weeks) is that the wetland behaves as a partially well mixed reactor [Andradóttir, 1997]. This supports the use of a simple one-dimensional (1-D) model such as the dead-zone model for low-flow conditions, even though the instantaneous circulation may be far from 1-D.

4.2.2. Storms. The cross-hatched areas on Figure 4 show that the diurnal thermal fluctuations are reduced or erased during storms (e.g., J.D. 229, 265, and 282) as the wetland heat budget is more strongly influenced by the river heat input than atmospheric fluxes. Other important storm features are demonstrated in the three large storms on Figure 5. First, consider the largest storm in 1997, portrayed on Figure 5a. As is typical for late fall storms, the river is a source of warm water to the wetland. Unlike low-flow conditions the water temperatures in the channel and dead zone deviate from one another. In particular, during the initial stage of the storm (J.D. 305.7-306) the water temperature in the channel increases abruptly, whereas the water temperature in the dead zone rises more gradually. This is consistent with short circuiting (Figure 1b) in which a large portion of the river advects rapidly through the wetland channel while the remainder mixes slowly with the dead zones. The time lag between the temperature increase in the river and wetland channel (1 hour) matches the advective timescale estimated for the river jet [Chu and Baines, 1989]. Also, note that the water column in the flow zone is well mixed during the rising limb of the hydrograph, as expected when river inertia dominates buoyancy and wind ($F_l^{-2}$ and $\Omega < 0.5$). However, this one-layer flow structure breaks down on the receding limb shortly after the onset of a southerly wind at J.D. 306.0. The colder water that appears at the channel bed is likely advected from the northern dead zone by the subsurface return flow associated with the wind setup.

Next consider the August storm illustrated on Figure 5b. In

Figure 5. River and wetland water temperatures at $x = 0.4L$ during large storms in (a) November 1997, (b) August 1997, and (c) October 1998. Bold lines depict near-surface temperatures, and light lines depict near-bed temperatures.
the beginning of the storm the river is a source of cold water to the wetland, the reverse of the previous storm. Short circuiting is again observed as the channel temperature drops more abruptly than the dead zone temperature around J.D. 234.0. Furthermore, the water column is well mixed on the rising limb of the storm but becomes stratified on the receding limb. Unlike the previous storm, this transition cannot be easily explained based upon the available data, because it does not coincide with wind setup nor a significant increase in buoyancy forcing. After J.D. 234.0 the colder river flows through the wetland as an underflow until around J.D. 234.4, when the river becomes warmer than the wetland water. The river then switches to an overflow that appears to be short-circuiting, indicated by the sudden drop in the near-surface temperature in the channel that is not mirrored in the dead zone around J.D. 234.7.

Finally, consider the largest storm in 1998, depicted on Figure 5c. The wetland thermal response is very complicated, both because this event consists of a series of storms and because it occurs on windy days (peak wind speed 7 m/s). In the following we focus solely on the second storm (J.D. 282.2), when the river is warmer than the wetland water and the wind is blowing from the north, producing less interference with the jet structure than in the first storm (J.D. 281.6). The temperature records on the rising limb of this storm show the warm river pulse moving through the wetland as an overflow, reaching the channel station more rapidly than both the northern and southern dead zones. This is in agreement with the conceptual picture of short circuiting portrayed on Figure 2b in which the channel cuts through the middle of the wetland. The river continues to short-circuit on the receding limb (J.D. >282.2), as seen by the fact that the water in the channel is significantly warmer than that in the northern dead zone. Furthermore, in contrast to the storms on Figures 5a and 5b, the flow is stratified on the rising limb of the storm but is well mixed on the receding limb.

To summarize, although the temperature records during storm events are quite complex, they consistently demonstrate that the river short-circuits through the wetland (Figure 1b). Our observations thus support the hypothesis that wetland circulation changes during storms from partially well mixed to short-circuiting. In addition, the upper forebay exhibits vertical structure both during low flows and storms. This suggests that the use of a depth-averaged dead-zone model is at times an oversimplification, especially during storms. This will be investigated in greater detail in section 4.3.

4.3. Dead-Zone Model Simulations

To further validate the application of the dead-zone model, we now consider how well the model simulates the thermal response within the upper forebay during both low flows and storms.

4.3.1. Low flows. Figure 6 illustrates model simulations and observations during a 10 day period in 1997, when the Aberjona River flow rate was 0.1 m³/s. Since the dead-zone model is 1-D, we compare the simulated depth-averaged temperatures (solid lines) to the observed near-surface and near-bed water temperatures (dotted lines), which define the upper and lower limit for an observed depth-averaged temperature. Figure 6a shows that the model simulates well low-flow conditions, producing values that lie within the observed temperature range, except on windy days (J.D. 246–248) when simulations are colder than observations. This overprediction of convective cooling can be explained by the fact that the model does not account for wind sheltering. During this low-flow period the residence time in the wetland is so long, r_w ~ 8, that the upper forebay approaches the limit of functioning like a stationary water body [Andradóttir and Nepf, 2000]. At this limit the thermal budget is predominantly governed by the surface heat flux and not the river heat input, and the detailed circulation regime in the wetland is not so important in predicting thermal mediation. This is demonstrated by the fact that both the stirred reactor and the dead-zone models predict very similar wetland outflow temperatures (Figure 6b). Note, however, the wetland circulation is still important for predicting chemical or solid removal which is unaffected by atmospheric exchange [Thackston et al., 1987].

4.3.2. Storms. Figure 7 summarizes model simulations during the three large storms in 1997 and 1998. Recall from section 4.2 that these storms all exhibit short circuiting either as a one layer-flow or as an overflow/underflow. An important function of the model with respect to lake transport is to predict the temperature of this short-circuiting flow. Since the model is 1-D, this may involve simulating only the surface wetland temperature during overflows (or bed temperatures during underflows). First, consider the 1997 November storm, which has the strongest signature of short circuiting. Figure 7a shows that the dead-zone model predicts well the propagation of the warm short-circuiting river flow to the middle of the wetland (x = 0.4L), even after the flow is no longer 1-D (J.D.
Figure 7. Storm simulations for the (a) 1997 November, (b) 1997 August, and (c) 1998 October storms. Dead-zone model simulations (solid lines) are compared with observations (dotted lines) and stirred reactor simulations (dashed lines) at different locations within the wetland. The horizontal bars indicate when the flow is jet-dominated, i.e., $F_i < 2$ and $\Omega < 0.5$.

The model, however, slightly underpredicts the water temperature in the northern dead zone, probably because it does not account for lateral exchange generated by wind-driven circulation. For the 1997 August storm portrayed on Figure 7b, the model again simulates well the channel and dead-zone temperatures in the middle of the wetland ($x = 0.4L$). Similarly, Figure 7c shows that the dead-zone model does a good job predicting the movement of the thermal front along the wetland channel both at $x = 0.4L$ and $x = L$ during the 1998 October storm. Recall that the resolution of the thermistor probes is 0.2°C, and the difference between the observed and simulated water temperatures (0.2–0.3°C) is therefore not significant. These three simulations thus demonstrate that the 1-D, quasi-steady, dead-zone model gives reliable predictions of short circuiting and the transition between short circuiting and laterally well mixed flow during storms, even in systems such as the Upper Mystic Lake forebays which exhibit vertical structure (i.e., non-1-D behavior).

Finally, the comparison between the dead-zone and stirred reactor model simulations, given at the bottom on Figure 7, highlights the strength of the dead-zone model relative to the stirred reactor model. The dead-zone model captures the sharp rises/drops in water temperature associated with short circuiting, whereas the stirred reactor model predicts a diffused front because it assumes that the river mixes instantaneously throughout the wetland. This diffusion of frontal signatures can affect the prediction of lake intrusion depth. This can be seen by comparing the model simulations to the mean temperature in the lake epilimnion $T_{ep}$, as on Figure 7a. Notice that because of the larger receiving volume of the lake, the lake water temperature does not follow the same sharp temperature increase observed in the wetland. During this 1997 November storm the stirred reactor model predicts a 4 hour delay in surface intrusions ($T_{ep} > T_L$) during the peak flow rates. Since the highest concentrations typically occur during the rising limn of a storm, the stirred reactor model will underpredict the contaminant/nutrient transport to the lake surface, which can have important implications for human health and
reservoir management as will be discussed in greater detail in section 4.4.

4.4. Lake Intrusion Dynamics and Water Quality

Wetland thermal mediation is an important physical process in part because it can alter lake intrusion dynamics, which, in turn, impact downstream water quality. Figure 8 describes the expected intrusion depth in the Upper Mystic Lake with and without the upper forebay, based on the water temperature measurements in 1997, as discussed in section 3.2. The seasonal and synoptic trends presented in Figure 8a show that in the absence of the wetland, the colder river water would plunge when entering the lake throughout the fall except for a few days of short-term heating (e.g., J.D. 284 and 305-310). However, with the littoral wetland present, episodes of surface intrusion are prolonged and occur during 28 days or 40% of the monitoring period (e.g., J.D. 256-265, 279-291, and 306-312). Therefore substantially more river-borne fluxes are delivered into the surface water of the lake with the wetland present, which can have a high impact on the nutrient and chemical budgets of the lake. A close-up on diurnal timescales (Figure 8b) shows that the intrusion depth can also vary significantly over the course of a day, especially during synoptic heating periods when the inflow inserts into the lake epilimnion. During these times a small temperature variation can lead to a large change in intrusion depth because of the low thermal gradient in the lake surface layer. In the Upper Mystic Lake these variations can be as large as 4-6 m. Overall, the temperature measurements in the Upper Mystic Lake system suggest that the presence of a littoral wetland shifts the timescale of lake intrusion depth variability from predominantly seasonal (no wetland) to synoptic and diurnal timescales (with wetland). A similar result was predicted by the linearized dead-zone model [Andradottir and Nepf, 2000]. The shift in intrusion depth variability can impact lake water quality. For example, nutrients and contaminants may be more homogeneously distributed over depth as the river inflow covers a larger vertical domain in the lake. In addition, lateral and vertical transport processes in the lake can be enhanced. Both of these will affect the management of reservoir use and withdrawals.

The intrusion depth during storms is of particular importance for lake water quality. Because of surface runoff and remobilization of channel sediments, river nutrient and contaminant concentrations usually increase during storms. For example, storms account for 30-60% of the total annual fluxes of heavy metals such as arsenic, chromium, iron, and copper to the Upper Mystic Lake [Soto-Gabriele, 1995]. The arrows on Figure 8a depict the storm occurrences in the Upper Mystic Lake system during fall 1997. As expected from the previous discussion, the presence of a wetland does not significantly modify the lake intrusion depth during storms because of the low effective thermal capacity associated with the high flow rates. Figure 8a also shows that seven out of nine of the late summer and fall storms occur during periods of synoptic cooling and produce intrusions in the thermocline. These measurements thus suggest that elevated surface concentrations happen infrequently during summer and fall in the Upper Mystic Lake.

5. Conclusions

Thermal mediation in littoral wetlands is an important process that can alter the intrusion depth in lakes and thus can determine the downstream fate of land-borne nutrients and contaminants. Measurements in the Upper Mystic Lake system in Massachusetts during summer and fall 1997-1998 demonstrate that thermal mediation occurs in a natural littoral wetland located in a small-to-medium-sized watershed, where sun shading produces a different water temperature for the river than the wetland and lake. In addition, thermal mediation is affected by flow conditions. During low flows the wetland raises the temperature of the lake inflow by approximately 1-3°C, which is sufficient to change the lake intrusion depth by 1-6 m and shift the timescale of intrusion depth variability from seasonal (no wetland) to synoptic and diurnal (with wetland). In contrast, during storms almost no thermal mediation occurs. These observations are both in agreement with the theory presented by Andradottir and Nepf [2000]. The change in thermal response associated with flow conditions can partially be explained by a shift in wetland circulation: During low flows the river inflow fills the entire wetland volume, and the wetland behaves like a partially well mixed reactor. During storms, however, the temperature records demonstrate that the river short-circuits across the wetland, and thus only a fraction of the wetland volume is available to produce thermal mediation. A simple 1-D, quasi-steady, dead-zone model can...
give reliable predictions of thermal mediation during both regimes and predicts well the transition between short circuiting and laterally well mixed flow.

Notation

- $A_c, A_d$: cross-sectional area of wetland channel and dead zone, respectively.
- $A_r$: river cross-sectional area at wetland entrance.
- $C_r$: friction factor.
- $F_i$: inverse internal Froude number, equal to $|\Delta \rho|/\rho g H_u^2$.
- $g$: acceleration of gravity.
- $H$: mean wetland water depth.
- $H$, $H_d$: mean water depth in wetland channel and dead zone, respectively.
- $K$: surface heat transfer coefficient.
- $L$: wetland length.
- $\Delta Q$: lateral exchange flow rate.
- $Q_r$: river flow rate.
- $r$: thermal capacity, equal to $t_{\text{heat}}/t_{\text{heat}}$.
- $t_{\text{heat}}$: nominal residence time, equal to $HL/Q_r$.
- $T_c, T_a$: wetland channel and dead-zone temperature, respectively.
- $T_L$, $T_R$: lake and river temperature, respectively.
- $\alpha$: lateral exchange coefficient, equal to $\Delta Q/Q_r$.
- $\beta$: linear spreading coefficient.
- $\phi$: surface heat flux.
- $\Lambda$: drag parameter, equal to $\Lambda C_r L/H$.
- $\rho$: water density.
- $\Delta \rho$: density anomaly between river and wetland.
- $\tau_w$: mean wind shear stress.
- $\Omega$: wind parameter, equal to $\tau_w L/(\rho U_0^2 H)$.

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