



1 **Trends and abrupt changes in 104-years of ice cover and**
2 **water temperature in a dimictic lake in response to air**
3 **temperature, wind speed, and water clarity drivers**

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5 **M. R. Magee¹, C. H. Wu¹, D. M. Robertson², R. C. Lathrop³, and D. P. Hamilton⁴**

6 [1]{Civil and Environmental Engineering, University of Wisconsin-Madison, Madison,
7 Wisconsin, USA}

8 [2]{Wisconsin Water Science Center, U.S. Geological Survey, Middleton, Wisconsin, USA}

9 [3]{Center for Limnology, University of Wisconsin-Madison, Madison, Wisconsin, USA}

10 [4]{Environmental Research Institute, University of Waikato, Hamilton, New Zealand}

11 Correspondence to: C.H. Wu (chinwu@engr.wisc.edu)

12

13 **Abstract**

14 The one-dimensional hydrodynamic-ice model, DYRESM-WQ-I, was modified to simulate
15 ice cover and thermal structure of dimictic Lake Mendota, WI, USA, over a continuous 104-
16 year period (1911-2014). The model results were then used to examine the drivers of changes
17 in ice cover and water temperature, focusing on the responses to shifts in air temperature,
18 wind speed, and water clarity at multi-year time scales. Observations of the drivers include a
19 change in the trend of warming air temperatures from 0.081 °C per decade before 1981 to
20 0.334 °C per decade thereafter, as well as a shift in mean wind speed from 4.44 m s⁻¹ to 3.74
21 m s⁻¹ in 1994. Observations show that Lake Mendota has experienced significant changes in
22 ice cover: later ice on (9.0 days later per century), earlier ice-off (12.3 days per century),
23 decreasing ice cover duration (21.3 days per century), while model simulations indicate a
24 change in maximum ice thickness (12.7 cm decrease per century). Model simulations also
25 show changes in the lake thermal regime of: earlier stratification onset (12.3 days per
26 century), later fall turnover (14.6 days per century), longer stratification duration (26.8 days
27 per century), and decreasing summer hypolimnetic temperatures (-1.4 °C per century).
28 Correlation analysis of lake variables and driving variables revealed ice cover variables,



1 stratification onset, epilimnetic temperature, and hypolimnetic temperature were most closely
2 correlated with air temperature, whereas freeze-over water temperature, hypolimnetic heating,
3 and fall turnover date were more closely correlated with wind speed. Each lake variable (i.e.,
4 ice-on and ice-off dates, ice cover duration, maximum ice thickness, freeze-over water
5 temperature, stratification onset, fall turnover date, stratification duration, epilimnion
6 temperature, hypolimnion temperature, and hypolimnetic heating) was averaged for the three
7 periods (1911-1980, 1981-1993 and 1994-2014) delineated by abrupt changes in air
8 temperature and wind speed. Average summer hypolimnetic temperature and fall turnover
9 date exhibit significant differences between the third period and the first two periods.
10 Changes in ice cover (ice-on and ice-off dates, ice cover duration, and maximum ice
11 thickness) exhibit an abrupt change after 1994 which was related in part to the warm El Niño
12 winter of 1997–1998. Under-ice water temperature, freeze-over water temperature,
13 hypolimnetic temperature, fall turnover date, and stratification duration demonstrate a
14 significant difference in the third period (1994–2014), when air temperature was warmest and
15 wind speeds decreased rather abruptly. The trends in ice cover and water temperature
16 demonstrate responses to both long-term and abrupt changes in meteorological conditions that
17 can be complemented with numerical modelling to better understand how these variables will
18 respond in a future climate.

19

20 **1 Introduction**

21 Many studies have shown that lake temperatures and ice cover can strongly affect water
22 chemistry, individual organism physiology, population abundance, community structure, and
23 food-web dynamics (Weyhenmeyer et al. 1999; Straile, 2000; Gerten and Adrian 2000;
24 Arhonditsis et al. 2004b). Air temperature, wind speed, and water clarity are important factors
25 driving these lake ecosystem properties. Understanding how lakes respond to changes in these
26 drivers is of great interest to predict how lakes may change in the future (Robertson 1989;
27 Magnuson et al. 1997; Fang and Stefan 2009). The long-term response of lake ice and water
28 temperature to changing air temperature and wind speed is integral to assessment of the
29 potential impacts of climate change on water quality and ecology of lakes.

30 Over the past 100 years, climate has been changing and will continue to change (IPCC 2013).
31 Globally-averaged combined land and ocean surface temperature data show a linear warming
32 trend of 0.85 °C from 1880–2012 (IPCC 2013). This warming was most pronounced from



1 1979–2012, greater than 0.25 °C per decade (Hartmann et al., 2013) Increases in air
2 temperature alter the ice cover of lakes (Robertson et al. 1992; Magnuson et al. 2000; Butcher
3 et al. 2015) and affect their thermal structures (Robertson and Ragotzkie 1990), evidenced by
4 increasing epilimnetic temperatures (Schindler et al. 1990; Arhonditsis et al. 2004a; Dobiesz
5 and Lester 2009), warming of the lake surface temperature (Schneider and Hook, 2010;
6 Shimoda et al. 2011), increasing temperature gradient across the thermocline (King et al.
7 1997; Livingstone 2003), changing thermocline depth (Schindler et al. 1990; King et al.
8 1997), advancing the onset of summer stratification (Austin and Colman 2007), delaying fall
9 turnover (King et al. 1997), increasing the strength of thermal stratification (Rempfer et al,
10 2010), and prolonging the stratified period (Robertson and Ragotzkie 1990; Wilhelm and
11 Adrian 2008).

12 Trends in wind speed over the last 30–50 years have been reported in several studies that have
13 analyzed historical wind speed records across the globe (Jiang et al. 2010; Wan et al. 2010).
14 Klink (2002) examined 22- to 35-year records (ranging between 1959–1995) of wind speed at
15 seven stations in and around Minnesota and found decreasing annual wind speeds at five of
16 the seven stations. Pryor et al. (2009) reported that the 50th and 90th percentile annual wind
17 speeds over the period 1973–2005 across most of the U.S. have also decreased. Decreased
18 wind speeds increase thermal stratification and can reduce whole-lake average temperature
19 (Tanentzap et al. 2008). Interestingly, an opposing trend (increasing wind speed) has been
20 observed in Lake Superior, North America, where the lake surface temperatures have been
21 warming faster than air temperatures (Austin and Colman 2007). Desai et al. (2009) suggest
22 that the larger increase in water temperatures than air temperatures reduced the air-water
23 temperature gradient and destabilized the atmospheric surface layer above Lake Superior,
24 which resulted in increasing wind speed at a rate of nearly 5% per decade. Differences in
25 wind-driven mixing may explain different temperature responses of hypolimnetic waters in
26 large and small lakes (Winslow et al. 2015). While the importance of wind in lake heat
27 transfer (Fu et al. 2009; Read et al. 2012), mixing, and thermal structure (Schindler et al.
28 1990; Desai et al. 2009) has been recognized, studies on the effects of wind speed alterations
29 on seasonal ice cover and thermal structure of lakes are still rare.

30 Water clarity, which controls the amount of solar radiation penetrating into a lake, plays an
31 important role in heat budgets in lakes. Increased water clarity can result in warmer deep
32 waters (Stefan et al, 1996), while reduced light penetration can result in decreased mixing



1 depth and cooler deep waters (Hocking and Straskraba, 1999; Tanentzap et al., 2008).
2 Previous studies have also shown that increased water clarity was correlated with deeper
3 mixed-layer depth (Mazumder and Taylor 1994; Fee et al. 1996; Schindler et al. 1996) and
4 increased hypolimnetic heating rate (Yan 1983). Water clarity has important consequences for
5 photosynthesis and vertical distribution of biota. Accompanied by a warming climate,
6 increased evaporation and changes in precipitation patterns can alter the inputs of nutrients
7 and dissolved organic carbon (DOC) into lakes, resulting in changes in water clarity
8 (Schindler et al. 1990; Schindler et al. 1996).

9 Abrupt shifts, rather than linear changes, can occur in both climate and lake variables over
10 long time scales. In the past, the majority of referenced studies used some form of linear
11 regression analysis to look for changes, which assumes that lakes undergo monotonic changes
12 over time (Van Cleave et al., 2014). This assumption can mask the occurrence of step changes
13 (Liu et al., 2013; North et al., 2013) and can hide the underlying mechanisms that are
14 responsible for the changes (North et al., 2014). As noted in other studies (Mueller et al.,
15 2009; Van Cleave, 2014), long-term changes in lake thermodynamic variables can be
16 associated with a pronounced, nonlinear step change that indicates a rapid switch between
17 stable states or regimes (Scheffer et al., 2001; Rodionov, 2004; North et al., 2013). Identifying
18 this type abrupt change in climate drivers and corresponding abrupt or gradual change in lake
19 variables may shed light on the role of changing climate on the corresponding ecosystem.

20 The purpose of this study is to investigate how long-term changes in air temperature, wind
21 speed, and water clarity affect the ice cover and thermal structure of a dimictic Lake Mendota,
22 Wisconsin, USA during the past century using a long-term 104-year simulation model. We
23 hypothesize that changes in lake ice cover (ice-on and ice-off dates, ice cover duration, and
24 maximum ice thickness) and thermal structure (stratification onset, fall turnover date,
25 stratification duration, water temperatures, hypolimnetic heating) variables may be
26 characterized by periods of abrupt change rather than gradual trends, based on observations of
27 rapid change in the climate drivers of air temperature and wind speed. To address this, a one-
28 dimensional hydrodynamic model with ice cover is first validated using a long-term (104-
29 year) observational dataset. The model is then employed to simulate long-term (1911-2014)
30 ice cover and water temperature in the lake. With the knowledge of lake responses to these
31 past conditions, we aim to reveal how lakes might respond to future changes in air
32 temperature, wind speed, and water clarity.



1 **2 Methods**

2 **2.1 Hydrodynamic model**

3 An ice and snow model was added to the DYRESM-WQ model (Hamilton and Schladow
4 1997), a physically-based one-dimensional hydrodynamic model for simulating vertical
5 mixing and advective transport in lakes and reservoirs. The resulting model, DYRESM-WQ-I,
6 is validated and employed to simulate daily changes in the vertical distribution of water
7 temperature and ice cover in Lake Mendota from 1911-2014. In this model, the lake is
8 represented by a series of Lagrangian horizontal layers with uniform properties that may
9 change in elevation and thickness in response to inflows/outflows and surface mass fluxes
10 (evaporation and precipitation). Layer thickness is updated using an algorithm to give
11 appropriate vertical density resolution at each time step. Mixing in the model is represented
12 by merging the layers when the sum of available turbulent kinetic energy (TKE) produced by
13 wind stirring, convective turnover, and shear stress exceeds the potential energy required to
14 mix the adjacent layer below. Hypolimnetic mixing is modeled with an eddy diffusivity
15 coefficient, which is a function of the dissipation of TKE and strength of stratification. More
16 detailed descriptions of the simulation of water temperature and mixing are provided by
17 Imberger and Patterson (1981).

18 The ice model is based upon a quasi-steady state assumption that the time scale for heat
19 conduction through the ice is short relative to the time scale of meteorological forcing
20 (Patterson and Hamblin, 1988; Rogers et al., 1995). This assumption is valid under a Stefan
21 Number < 0.1 (Hill and Kucera, 1983). The ice module is applied when the simulated surface
22 water temperature first drops below $0\text{ }^{\circ}\text{C}$; the initial ice thickness is set to a value of 5 cm
23 (Patterson and Hamblin, 1988; Vavrus et al., 1996). Upward conductive heat flux between
24 ice/snow cover and the atmosphere, q_0 , is determined by numerically solving the quasi-
25 steady state heat conduction equations (Rogers et al., 1995) and assigning appropriate
26 boundary conditions to the water, ice, and atmospheric interfaces. At the ice (or snow)
27 surface, a heat flux balance provides the condition for surface melting, and accretion or
28 ablation of ice is determined through the heat flux at the ice-water interface. Imbalance
29 between heat conduction through ice and the heat flux from the water to the ice gives the rate
30 of change of ice thickness at the ice-water interface. Snow conductivity is estimated from its
31 density using an empirical equation (Ashton, 1986), and snow compaction is based on an



1 exponential decay formula (McKay, 1968), with snow compaction parameters based on air
 2 temperature and snowfall/rainfall (Rogers et al., 1995). Snow (white) ice is generated in
 3 response to flooding, when the mass of snow that can be supported by the ice cover is
 4 exceeded. The uppermost solid layer of ice or snow is adjusted in thickness at the end of each
 5 1-hr model time step according to the balance of the heat budget. When ice thickness
 6 decreases to less than 5 cm, open water conditions are restored.

7 Sediment heat flux, the main external source of heat after freezing, is important to water
 8 temperatures beneath ice cover (Ellis et al., 1991). Sediment heat flux (Fang and Stefan,
 9 1996) is included as a source/sink term for each Lagrangian layer to closely simulate under-
 10 ice water temperatures. A simple diffusion relation (Rogers et al., 1995) is used to estimate
 11 heat transfer from the sediments to the water column, q_{sed} :

$$12 \quad q_{sed} = K_{sed} \frac{dT}{dz}. \quad (1)$$

13 where K_{sed} is the sediment conductivity ($=1.2 \text{ W m}^{-1} \text{ }^\circ\text{C}^{-1}$). dT/dz , the temperature gradient
 14 across the sediment-water interface, is estimated as:

$$15 \quad \frac{dT}{dz} \approx \frac{T_s - T_w}{z_{sed}}. \quad (2)$$

16 where T_s is the sediment temperature, T_w is the water temperature adjacent to the sediment
 17 surface, which varies hourly and with depth, and z_{sed} is the distance beneath the water-
 18 sediment interface at which the sediment temperature becomes largely invariant. From data
 19 collected at four locations on Lake Mendota (Birge et al., 1927), it was found that sediment
 20 temperatures varied little at 5 m depth below the sediment-water interface, so z_{sed} is set to be 5
 21 m. In addition, data from Birge et al. (1927) is used to fit a curve to describe the seasonal
 22 variation of T_s

$$23 \quad T_s = 9.7 + 2.7 \sin \left[\frac{2\pi(D-151)}{TD} \right] \quad (3)$$

24 where D is the number of days from the start of the year and TD is the total number of days
 25 for the year of interest (365 or 366). The vertical transfer of heat in the water column beneath
 26 the ice is regulated by an assigned thermal diffusion coefficient. The formulation is the same
 27 as that originally used in DYRESM-WQ (Hamilton and Schladow, 1997) to simulate heat



1 transfer throughout the open water period. This produced diffusivities that were within the
2 range of measurements by Ellis et al. (1991) of 1-3 times greater than molecular values.
3 Input for the model includes lake morphometry (lake volume and surface area as a function of
4 elevation), initial vertical profiles for water temperature and salinity, Secchi depth,
5 meteorological variables, and inflows/outflows. The model calculates the surface heat fluxes
6 using meteorological variables: total daily shortwave radiation, daily cloud cover, air vapor
7 pressure, daily average wind speed, air temperature, and precipitation. During the simulation,
8 all parameters/coefficients in the model are kept constant. The time step in the model for
9 calculating water temperature, water budget, and ice thickness was set to 1 hr. Snow ice
10 compaction, snowfall and rainfall components are updated at a daily time step, corresponding
11 to the frequency of meteorological data input. Cloud cover, air pressure, wind speed, and air
12 temperature are assumed to be constant throughout the day, and precipitation is assumed
13 uniformly distributed. Shortwave radiation distribution throughout the day is computed based
14 on the lake latitude and the day or year. The DYRESM-WQ-I model is calibrated using
15 measured lake variables including water level, temperature profiles, ice thickness, and ice on
16 and ice off dates. The overall simulation period was 104 years, starting with an isothermal
17 (measured) water column temperature of 3.1 °C on 7 April 1911 and ending on 31 October
18 2014.

19 **2.2 Piecewise regression algorithm**

20 Breakpoints in the air temperature trend over the study period were determined using a
21 piecewise linear regression (PLR) method (Toms and Lesperance, 2003; Tome and Miranda,
22 2004, Ying et al, 2015) that assumes continuity in the trendlines across the breakpoint. The
23 piecewise linear regression model finds the breakpoint, B , that minimizes the residual sum of
24 squares (RSS_{PLR}) of the model between the two phases (Ying et al, 2015)

$$25 \quad RSS_{PLR} = \sum_{i=1}^n [y_i - a - k_1 x_i - (k_2 - k_1) \max(x_i - B, 0)]^2 \quad (4)$$

26 where x_i and y_i are the time and air temperature data corresponding to the i -th data point,
27 respectively, and k_1 and k_2 are the slopes of the linear fit to data before and after the
28 breakpoint B . The parameter a is the intercept of the linear fit to data below the estimated
29 breakpoint B , and n is the series length.



1 An F -test was used to compare the residual sum of squares of an ordinary linear regression
2 model with the piecewise linear regression model with a level of $\alpha = 0.05$ used to indicate
3 significance, and the breakpoint year was determined to pass with $p = 0.0016$.

4 **2.3 Sequential t-test analysis of abrupt changes**

5 Abrupt changes in mean annual wind speeds and lake ice cover and temperature variables
6 were detected using the sequential t-test STARS (Rodionov, 2004), which can automatically
7 detect multiple change points. A shift occurs when a statistically significant difference exists
8 between the mean value of the variable before and after a certain point based on the t-test. The
9 variables were tested using a threshold significance level $p = 0.05$, a Huber weight parameter,
10 $h = 1$ (Rodionov, 2006; North et al, 2013), and a cut-off length $L = 15$ years (North et al,
11 2013).

12

13 **3 Model input data**

14 **3.1 Bathymetry**

15 Lake Mendota (43°40'N, 89°24'W) has a surface area of 39.4 km² and maximum fetch of 9.8
16 km (Kitchell, 1992). The mean depth is 12.7 m, and maximum depth is 25.3 m. For detailed
17 bathymetry refer to Kamarainen et al. (2009).

18 **3.2 Meteorological variables**

19 Daily meteorological data required to run the DYRESM-WQ-I model include solar radiation,
20 air temperature, vapor pressure, wind speed, cloud cover, rainfall, and snowfall. Daily air
21 temperatures are computed as the average of daily maximum and minimum temperatures.
22 Vapor pressure, wind speed, and cloud cover are also entered as daily average values while
23 solar radiation, rainfall, and snowfall are daily accumulation values.

24 Meteorological data for the Madison area have been continuously recorded since 1869,
25 however, the station and techniques have changed several times. Robertson (1989)
26 constructed a continuous, homogeneous daily meteorological dataset from 1884 to 1988 by
27 adjusting for changes in site location and observation time, and resultant changes in the
28 surface roughness (e.g. height of surrounding trees and buildings). These data were appended
29 with the most recent weather station of the National Climate Data Center (NCDC, NOAA)



1 located in Madison (MSN) Dane County Regional Airport (Truax Field), the same site as that
2 used in 1988. Since Robertson (1989) adjusted all historical data to that collected in 1988, no
3 adjustments are applied to the recent data except for wind. In 1996, a discontinuity in the
4 wind record was caused by change in observational techniques and sensor locations (McKee
5 et al. 2000). To address the non-climatic changes in wind speed, data from MSN are carefully
6 compared with those collected from the tower of the Atmospheric and Oceanic Science
7 Building at the University of Wisconsin-Madison
8 (<http://ginsea.aos.wisc.edu/labs/mendota/index.htm>). Hourly data from both sites ($U_{\text{MSN,hourly}}$
9 and $U_{\text{AOS,hourly}}$) during 2003–2010 were used to form a 4×12 (four components of wind
10 direction \times 12 months) matrix ($K_{4,12}$) of wind correction factors, yielding $U_{\text{AOS,daily}} =$
11 $K_{i,j} \times U_{\text{MSN,daily}}$. A comparison of results indicated that the MSN weather station demonstrated a
12 higher magnitude in winds out of the east by 5% and lower magnitude in winds out of the
13 west and south by 30% and 10%, respectively. The adjusted wind data ($=K_{i,j} \times U_{\text{MSN,daily}}$) are
14 employed and used in the model simulation. Overall the adjusted wind data show a decline in
15 mean wind velocities of 16% from 1988–93 to 1994–2014) compared to 7% at a nearby
16 weather station with no known observational changes (St. Charles, Illinois; 150 km southeast
17 of Lake Mendota).

18 3.3 Light extinction

19 Seasonal Secchi depths can be used to determine the light extinction coefficient for
20 DYRESM-WQ-I. Lathrop et al. (1996) compiled Secchi depth data for Lake Mendota
21 between 1900 and 1993 (1,701 daily Secchi depth readings from 70 calendar years), and
22 summarized the data for six seasonal periods: winter (ice-on to ice-out), spring turnover (ice-
23 out to 10 May), early stratification (11 May to 29 June), summer (30 June to 2 September),
24 destratification (3 September to 12 October), and fall turnover (13 October to ice-on). After
25 1993, Secchi depths were obtained from the North Temperate Lakes – Long Term Ecological
26 Research (NTL-LTER) (<https://portal.lternet.edu/nis/home.jsp#>). For years with no Secchi
27 data, the long-term mean seasonal Secchi depth was used to estimate light extinction. Light
28 extinction coefficients are calculated as a function of Secchi depth using the equation,
29 $k = 1.1 / z_s^{0.73}$ (Williams et al. 1980), where k is the light extinction coefficient and z_s is the
30 measured Secchi depth, in meters.



1 **3.4 River inflow and outflow**

2 Daily inflow measurements have been made on selected Lake Mendota tributaries since 1974,
3 and daily outflow from the lake has been measured since 1975. These measurements were
4 used to calculate total daily inflow and outflow. Total daily inflow and outflow from 1930 to
5 1974 were estimated from measurements made at the gauging station downstream of the lake
6 using the drainage-area ratio method (Maidment 1993). Prior to 1930, total inflow and
7 outflow were calculated using a water budget approach, i.e., the balance of inflow/outflow,
8 precipitation, evaporation, and lake level changes. The inflow/outflow is the residual
9 unknown term of the water balance where other terms include evaporation, rainfall, and water
10 level. The calculation is performed at the interval of water level measurements and the
11 residual term is then distributed evenly across the number of days between water level
12 measurements. Water level of Lake Mendota has been recorded since 1916 at the Yahara
13 River inlet in the north of the lake. Prior to 1916, the long-term mean lake level was assumed
14 in water budget calculations. River water temperature has been recorded at the Yahara River
15 inlet since 2002 (U.S. Geological Survey, <http://waterdata.usgs.gov/nwis>). To estimate daily
16 river temperatures prior to 2002, river temperatures were estimated from air temperatures. A
17 linear regression was used to correlate river temperatures and weekly average air temperature
18 data during 2002–2009 for air temperatures above the freezing temperature ($r^2 = 0.86$), and a
19 separate polynomial regression analysis was performed for air temperatures below the
20 freezing temperature ($r^2 = 0.68$).

21 **3.5 Lake ice cover and water temperature**

22 Observed data from 1911–2014 (available at NTL-LTER website:
23 <https://portal.lternet.edu/nis/home.jsp#>), showed that Lake Mendota, on average, freezes on
24 22 December (ice-on date), breaks up on 31 March (ice-off date), and has an ice duration of
25 98 days. Ice thickness and snow depth during the study period were compiled from various
26 sources: 1911–1916 (unpublished data from E. Birge, University of Wisconsin); daily ice
27 thickness during 1961–1962, 1962–1963, and for part of the winter 1963–1964 (Stewart
28 1965); 1975–1995 (unpublished data, D. Lathrop, Wisconsin Department of Natural
29 Resources); after 1995, ice thickness and snow depth were sampled once or twice every
30 winter by the NTL-LTER Program (<https://portal.lternet.edu/nis/home.jsp#>). In addition,
31 during winters of 2008–2009 and 2009–2010, blue ice, snow ice, total ice (blue ice + snow
32 ice), and snow cover depth were measured weekly at multiple locations on the lake (Hsieh,



1 2012). In total, there are 251 measurements of total ice, 21 for blue ice, 21 for snow ice, and
2 49 for snow depth during 1911 to 2014.

3 Long-term water temperature records were obtained from Robertson (1989) and the NTL-
4 LTER dataset (<https://portal.lternet.edu/nis/home.jsp#>). The frequency of water temperature
5 data varied widely from only one or two profiles per year to several profiles for a given day,
6 while there were no data collected in some years between 1931 and 1970. The vertical
7 resolution of the water profiles varied from 0.5 m to 2 m and sometimes 5 m when the water
8 column was weakly stratified.

9

10 **4 Results**

11 **4.1 Shifts in air temperature, wind speed, and water clarity**

12 Annual air temperature (Fig. 1a) had a relatively small increase from 1910 until 1980, but has
13 increased dramatically since 1981. Based on a piecewise linear regression algorithm, there
14 was a small warming trend of 0.081°C per decade during 1911–1980, followed by a dramatic
15 change (a warming trend of .334°C per decade) from 1981–2014. Figure 1b shows that mean
16 annual wind speed was 4.44 m s⁻¹ until 1994, when a significant shift occurred to 3.74 m s⁻¹
17 (15% reduction) based on the sequential t-test STARS method (Rodionov, 2004). Figure 1c
18 shows Secchi depth for Lake Mendota; however, no statistical trend was obtained due to the
19 incompleteness of the data.

20 Combining the statistically significant breakpoint in air temperature trend that occurred in
21 1981 and the shift in wind speed in 1994, the Madison climate may be broken into three
22 different periods. The first, from 1911–1980, was a relatively cool period and had an average
23 wind speed of 4.44 m s⁻¹. The second period (1981–1993) occurred after the breakpoint in the
24 air temperature trend and had a warmer air temperature and a wind speed of 4.44 m s⁻¹. The
25 third period (1994–2014) occurred after the shift in wind speed from 4.44 m s⁻¹ to 3.74 m s⁻¹
26 and had even warmer air temperatures.

27 **4.2 Model validation and long-term simulations**

28 Here we provide results of model validity comparisons followed by results of long-term
29 changes in both observed (where available) and simulated conditions for: snow and ice



1 thickness, ice-on and ice-off dates, lake water temperature, summer stratification, and
2 thermocline depth.

3 4.2.1 Ice and snow thickness

4 Figure 2 shows the comparison of the long-term (from 1911 to 2014) simulation and
5 measured data for blue ice, white ice, and total ice thickness and snow depth. Mean absolute
6 differences between simulated and measured thickness of total ice, blue ice, snow ice, and
7 snow cover are 7.8 cm (n=251; RMSE=7.9 cm), 5.5 cm (n=21; RMSE=3.3 cm), 1.9 cm
8 (n=21; RMSE= 2.3 cm), 4.1 cm (n=49; RMSE= 3.7 cm), respectively. Some discrepancy
9 between the model results and measurements may be due to the 1-D structure of the model.
10 The DYRESM-WQ-I model provides only the lakewide average ice and snow depths, hence
11 no horizontal variation. For a medium-to-large lake, such as Lake Mendota, spatial variations
12 in ice and snow depths should be expected (Bengtsson 1986). In the winters of 2008–2009
13 and 2009–2010, ice measurements were made at multiple locations on Lake Mendota (Hsieh
14 2012). Data in 2008–2009 showed variations in ice thickness. Total ice thickness in the
15 middle of the lake was approximately 12 cm greater than that at a littoral location (water
16 depth=2 m) in late winter. In 2009–2010, ice cover was fairly spatially uniform through the
17 entire sampling period. Consequently, discrepancy between model results and measurements
18 is expected in those years when ice thickness varied spatially. The accuracy of predictions
19 from DYRESM-WQ-I was compared with other models: MINLAKE (Fang and Stefan 1996)
20 produced standard errors of 11 cm for total ice thickness and 6 cm for snow cover for Thrush
21 Lake, Minnesota, and 12 cm for total ice and 7 cm for snow cover for the north basin of Little
22 Rock Lake, Wisconsin. Other models including LIMNOS (Vavrus et al. 1996) on Lake
23 Mendota, Wisconsin; MLI (Rogers et al. 1995) on Harmon Lake, British Columbia; and
24 CLIMo (Duguay et al. 2003) on lakes in Barrow, Alaska; Poker Flat, Alaska; and Churchill,
25 Manitoba produced similar errors to Lake Mendota between modeled and observed ice
26 thickness and snow cover. In general DYRESM-WQ-I predicts the ice and snow depths fairly
27 accurately. Figure 3a shows the evolution of simulated ice and snow thickness with multiple
28 ice and snow thickness measurements taken during the winter of 2009-2010. Once ice
29 develops, water beneath the ice continues to freeze, adding to the ice thickness, as heat is
30 conducted from water to the air. In this year, lake ice grew fast initially and then slowed
31 substantially. Eventually, the ice reaches its maximum thickness and stops growing (Ashton
32 1986). Once air temperatures begin to warm up, ice quickly thins until ice out. Overall, the



1 model is able to capture the growing ice cover as well as the temporally variable snow cover
2 found on the lake.

3 Figure 3b shows the long-term (1911–2014) simulated annual maximum ice thickness during
4 each winter. The annual maximum thickness varied widely from 20.1 cm (1997–1998) to 72.0
5 cm (1911–1912) during the 1911–2014 period. The timing when ice reached its maximum
6 thickness also varied considerably, from 17 January (1973) to 7 April (1923). In 73 out of 103
7 winters, ice reached its maximum thickness in March. Linear regression analysis on simulated
8 maximum ice thicknesses indicates that it has decreased at a rate of 12.7 cm per century
9 during 1911–2014. A t-test of the mean values shows a statistically significant ($p < 0.05$)
10 difference in the mean annual maximum ice thickness between period 1 (1911–1980) and
11 period 3 (1994–2014). Period 2 was not statistically different from either of the other two
12 periods.

13 4.2.2 Ice-on and ice-off dates

14 Figure 4 shows measured and simulated ice-on date, ice-off date, and ice duration on the lake.
15 The measured ice-on date is defined as the first day when the lake becomes fully ice-covered,
16 and the ice-off date is the last day of the ice breakup before the open water season. Simulated
17 results and observations are in good agreement, with a mean absolute difference of 2.3 days
18 (RMSE = 2.4 days) for ice-on date and 5.7 days (RMSE = 4.8 days) for ice-off date. It is noted
19 that both the mean error and RMSE are much smaller than the interannual variability in the
20 observed ice-on dates (standard deviation SD = 11.2 days) and ice-off dates (SD = 10.6 days).
21 Interannual variations in ice cover on Lake Mendota were large in the past century. Observed
22 ice-on dates ranged from 3 December (1929) to 30 January (1931): a range of 58 days. Ice-off
23 dates ranged from 27 February (1998) to 20 April (1923): a range of 51 days. The model
24 successfully captures the interannual variations of ice-on and ice-off dates. For example, it
25 reproduces the unusual late ice-on date in 1930–1931, and several noticeable early ice
26 breakups associated with intense El Niño -Southern Oscillation (ENSO) events, i.e., 1965,
27 1972, 1982, and 1997 (Anderson et al. 1996; Magnuson et al. 2000; Robertson et al. 2002).
28 Figure 4b shows the ice duration, defined as the period between ice-on and ice-off dates. The
29 mean absolute difference in ice duration between the model results and observations is 6.6
30 days (RMSE = 6.1 days), compared to the observed SD of 17.9 days. Overall, we consider
31 the model performs well in simulating ice on and off dates and ice cover duration.



1 Long-term trends in ice formation are examined by applying linear regression to the model
2 results and observed data. Figure 4 clearly shows progressively later freezing, earlier breakup,
3 and shorter duration in Lake Mendota from 1911 to 2014. Based on model results, ice-on
4 dates became later by 9.0 days, ice-off dates became earlier by 12.3 days, and ice duration
5 shortened by 21.3 days. Model results are in good agreement with those obtained from the
6 observed data (7.4 days later ice-on dates, 9.3 days earlier for ice-off dates, and 18.0 days
7 shorter duration). All linear trends from the observed and simulated data are statistically
8 significant ($p < 0.05$). Mean values of ice-on, ice-off, and ice duration for the three selected
9 periods, i.e., 1911–1980, 1981–1993, and 1994–2010, (see Table 1 and Fig. 4) show a
10 statistically significant ($p < 0.05$, based upon t-test) difference between period 1 (1911–1980)
11 and period 3 (1994–2014) for all ice cover variables and a statistically significant ($p < 0.05$)
12 difference between period 1 and period 2 (1981–1993) for ice cover duration.

13 4.2.3 Water temperature

14 The performance of the model in simulating water temperatures in the lake is presented in
15 Figures 5 and 6. For temperatures at the near-surface, we use the simulated volume-weighted
16 mean water temperatures between depths of 0 and 10m. The simulated epilimnetic
17 temperatures compare well with those estimated from the measured data; the annual mean
18 absolute error is 0.69 °C ($n=3,239$; $RMSE=0.30$ °C), demonstrating the ability of the model to
19 simulate the heat budget in the upper mixed layer of the lake. Near-surface temperatures range
20 from about 0 °C in winter to 26.1 °C in summer, and strongly respond to the net surface heat
21 flux, as illustrated by the 2–3 °C variations that occur in both simulated and measured data,
22 e.g. in the summer of 1921, 1960, 1994, and 2009 (Fig. 5). Overall the model captures
23 interannual variations in epilimnetic temperatures even during extreme years; for example, the
24 relatively high maximum epilimnetic temperatures (> 25 °C) in the summers of 1949, 1955,
25 1963, 1987, 1988, and 1999, and relatively cold annual maximum epilimnetic temperatures (\sim
26 22 °C) in the summers of 1915, 1924, and 1942.

27 Figure 6 shows the comparison between the simulated volume-weighted near-bottom (within
28 the hypolimnion) temperatures over the depths between 20–25 m and those estimated from
29 the measured data. The mean absolute error is 1.04 °C ($n=3,239$; $RMSE=0.53$ °C). Near-
30 bottom temperatures range from 0.2°C in winter to 19.1°C in autumn, which is approximately
31 10 °C less than the variation in near-surface temperatures. Near-bottom temperatures are
32 lowest when the lake first freezes, and then slowly increase because of heat released from the



1 bottom sediment although we cannot discount that cold inflows could also penetrate into the
2 lake and affect temperature at the selected depths (20-25 m). After ice break-up and just prior
3 to seasonal stratification, temperature in near-bottom waters is usually identical to surface
4 waters and the whole water column undergoes a period of sustained increase in temperature.
5 Hypolimnetic temperatures then stay relatively constant during the stratified period with
6 limited heat exchange associated with strong temperature gradients across the thermocline.

7 To further examine the model's capability of simulating the extreme differences in
8 temperature, we compare measured and modeled temperatures under years with cold (1924)
9 and warm (1963) epilimnion temperatures and cold (1972) and warm (1926) hypolimnion
10 temperatures. Figure 7 shows that the model accurately captures the extreme conditions in
11 those four years. For plotting, measurement values are assumed to remain constant for the
12 duration of time between measurements. Measurements are generally taken at 1 m intervals
13 over the depth of the water column, but in instances with irregular measurement intervals,
14 temperatures are interpolated at intervals of 1 m for the depth of the lake. Slight variation in
15 temperatures at each depth may be attributed to differences between the averaged seasonal
16 Secchi depth values used to drive the model and the observed Secchi depth values. To
17 demonstrate the model's ability to simulate changes in ice conditions and water temperature
18 under the ice, measured and simulated conditions for these extreme years are presented in Fig.
19 8. In this figure, the time frame and temperature scales are adjusted to highlight the subtle
20 changes in winter and ice conditions and modeled ice and snow thickness are added to the top
21 of the plots. While some discrepancies in temperature are recognized in the model
22 comparison, the model generally captures the inverse stratification under the ice cover.
23 Updated measurements of sediment temperature during winter may provide a better estimate
24 of the sediment heat flux during ice-cover conditions, and improve under-ice temperature
25 predictions. Overall, the model can reliably simulate water temperature in open water and ice
26 seasons for extreme cold and warm years.

27 Mid-summer (averaged between 16 July and 15 August when the thermocline was well-
28 established) epilimnetic (0–10m) and hypolimnetic (20–25m) simulated temperatures from
29 1911 to 2014 are shown in Fig. 9a. Epilimnetic temperatures range from 19.7 °C to 24.8 °C.
30 There is no statistically significant temporal trend in the mid-summer epilimnetic
31 temperatures, which may be partly explained by the lack of a warming trend for the mid-
32 summer air temperature in the Madison area based on the measured meteorological data



1 (Kucharik et al. 2010). In contrast, hypolimnetic temperatures range from 7.8 °C to 16.2 °C.
2 The interannual variation in hypolimnetic temperatures ($SD = 1.74^{\circ}C$) is greater than that in
3 epilimnetic temperatures ($SD = 1.0^{\circ}C$). Simulated hypolimnetic temperatures show a
4 decreasing trend of $-1.4^{\circ}C$ per century ($p < 0.05$). Figure 9b shows that the decreased
5 hypolimnetic temperatures resulted in the epilimnion-hypolimnion temperature difference
6 increasing by $2.0^{\circ}C$ per century ($p < 0.05$). Mid-summer hypolimnetic temperature and
7 epilimnion (0-10 m)-hypolimnion (20-25 m) temperature difference both exhibit significant
8 ($p < 0.05$) mean differences between period 1 (1911–1980) and period 3 (1994–2014), as
9 shown in Table 1.

10 Figure 10 shows the simulated long-term (104-year) mean under-ice water temperature
11 (volume-weighted average of all depths) ranging from $0.78^{\circ}C$ to $3.07^{\circ}C$ (mean = $1.74^{\circ}C$;
12 $SD = 0.54^{\circ}C$). Based on model results, under-ice temperatures reveal a significant increasing
13 trend ($p < 0.05$) of $0.48^{\circ}C$ per century during 1911–2014, suggesting that air temperature and
14 wind speed may affect under-ice water temperature. For the selected regimes, the under-ice
15 (bold lines) and freeze-over (dashed lines; volume-weighted average of all depths) water
16 temperatures exhibit interesting features. The difference between the water temperature over
17 the entire ice-covered period and the freeze-over water temperature has decreased through
18 time, indicating less heat gain during the winter, which may relate to a shorter time to transfer
19 heat and a smaller gradient in temperatures between the lake and the bottom sediments. The
20 under-water ice temperatures and freeze-over water temperatures both show significant
21 ($p < 0.05$) differences between regime 1 (1911-1981) and regime 3 (1994-2014), as shown in
22 Table 1.

23 4.2.4 Summer stratification period and thermocline depth

24 We characterize summer stratification by the dates of the onset of stratification and fall
25 turnover, and the total duration of stratification. Accurately describing these conditions
26 requires high frequency observations, which is challenging for long-term datasets. As a result,
27 we used modeled data to determine these variables. The dates of stratification onset/fall
28 turnover are defined as the day when the surface-to-bottom temperature difference is
29 greater/less than $2^{\circ}C$ (Robertson and Ragotzkie 1990). Figure 11 shows the dates of onset of
30 stratification breakdown (fall turnover), and the total duration of stratification. The onset dates
31 varied between 17 April (1977) and 30 June (1981), and the date of fall turnover varied from
32 02 August (1965) to 19 October (2013). Over 1911–2014, the onset of thermal stratification



1 has become earlier by 12.3 days ($p < 0.05$), and fall turnover has become later by 14.6 days
2 ($p < 0.05$), resulting in the stratification period increasing by 26.8 days ($p < 0.05$). Stratification
3 onset date shows no significant difference among the three regimes (see Table 1); however,
4 fall turnover date is significantly ($p < 0.05$) different between periods 2 and 3 and between
5 period 1 and 3, and the duration of stratification is significantly ($p < 0.05$) different between
6 periods 1 and 3, as shown in Table 1.

7 The thermocline depth, defined as the depth of maximum temperature gradient, is determined
8 from the model results (not shown in figure for brevity). During 1911–2014, the mid-summer
9 thermocline depth in Lake Mendota varied from 8.3 m to 12.4 m (mean = 10.7 m; SD = 0.84
10 m) with no significant long-term change.

11

12 **5 Discussion**

13 **5.1 Significance of lake drivers**

14 Changes in long-term simulated ice cover and thermal variables in Lake Mendota from
15 1911–2014 appear to occur over short time frames within this period, synchronous with rapid
16 changes in measured drivers (i.e., air temperature, wind speed, and water clarity). The
17 variables examined included ice on/off dates, maximum ice thickness, freeze-over water
18 temperature, mid-summer epilimnetic and hypolimnetic temperatures, summer hypolimnetic
19 heating, and dates of stratification onset and fall turnover. To identify the the relationship
20 between the drivers and each of these variables, we employed Pearson correlation analysis on
21 the detrended driver and simulation data. While using this method does not allow us to
22 directly determine causality, we may identify possible related drivers to changes in lake
23 variables and the relative importance of relationship of the three drivers (i.e. air temperature,
24 wind speed, and water clarity) to the lake variables. To calculate correlation coefficients, each
25 lake driver was averaged over a fixed period (e.g. April-May or November-December) and
26 then paired with each of the simulated lake variables from the same period. The averaging
27 period for air temperature and wind speed was chosen based on the fixed period that yielded
28 the best correlation. We chose to employ fixed periods because thorough testing indicated that
29 using a dynamic period introduced seasonal meteorological variations into the analysis. For
30 instance, the correlation coefficient between onset date and the mean air temperature averaged
31 over the prior month is expected to be high simply due to the seasonal variations over the



1 course of the year, i.e., the later in the spring, the warmer air temperature. Seasonal Secchi
2 depths were compiled for six seasonal periods: winter, spring turnover, early stratification,
3 summer, destratification, and fall turnover (Lathrop et al. 1996), and the period closest to the
4 averaging period used for air temperature and wind speed was used for correlation analysis.
5 The averaging periods used for calculating correlation coefficients of each pair of variables
6 are listed in Table 2. We discuss below the correlations for each pair of variables, shown in
7 Figure 12.

8 5.1.1 Ice cover variables

9 Ice-on and ice-off dates have been shown to be sensitive to climate conditions (Robertson et
10 al. 1992; Livingstone 1997; Magnuson et al. 1997). For both air temperature and wind speed,
11 the averages from 1 November to 31 December (ND) give the highest correlation with ice-on
12 date; in agreement with the results of Assel and Robertson (1995) and similar to Gao and
13 Stefan (1999). Averaged air temperature and wind speed over the period from 1 February to
14 31 March (FM) provide the highest correlation with ice-off date, consistent with the study by
15 Gao and Stefan (1999), who found that ice-off dates for ten Minnesota lakes correlated
16 significantly with February to April air temperatures. The strong positive correlation ($r =$
17 0.82) between ice-on date and N-D air temperature indicates a relationship between warmer
18 N-D air temperatures and later freezing dates. Strong negative correlation between ice-off
19 date and F-M air temperature ($r = -0.70$) indicates a relationship between earlier ice breakup
20 and warmer F-M air temperatures. In contrast to air temperature, ice cover is only weakly
21 related to the seasonal average wind speed. Nevertheless the correlation between ice-on date
22 and wind speed ($r = -0.26$) is significant, indicating a relationship between decreasing wind
23 speed and earlier ice-on. Since large wind events prevent ice formation by breaking up skim
24 ice, decreased wind speeds allow ice cover to form slightly earlier in the year. In the Madison
25 area, the decreasing wind speeds may act to slightly mitigate the effects of increasing air
26 temperature in later ice freezing. Correlation between ice dates and Secchi depths is low ($r =$
27 0.24 for ice on and $r = -0.22$ for ice off), indicating that water clarity has only a minor
28 relationship with ice dates, in agreement with the simulation results of Fang and Stefan
29 (1997). For both ice-on and ice-off dates, correlations with wind speed and Secchi depth are
30 much less significant than with air temperature, indicating that air temperature has a more
31 significant relationship with ice-on and ice-off dates and may be a more significant indicator
32 of ice cover dates and ultimately ice cover duration.



1 Annual maximum ice thickness is negatively correlated with the 1 January – 31 March (J-F-
2 M) averaged air temperature ($r = -0.77$) and weakly correlated with J-F-M wind speed ($r =$
3 0.21), indicating a better relationship between annual maximum ice thickness and air
4 temperature than for wind speed. The relationship of air temperatures on ice phenology is
5 consistent with previous findings for lakes in Northern Wisconsin, Canada, and Maine (Gao
6 and Stefan 2004).

7 The strong correlations between air temperature and ice cover variables agree with several
8 earlier studies (Anderson et al. 1996; Goa and Stefan 2004; Vavrus et al. 1996; Livingstone
9 1997; Williams et al. 2004). Snowfall has also been shown to be related to ice-off date
10 (Jensen et al. 2007) and ice thickness (Vavrus et al. 1996; Duguay et al. 2003) because
11 snowfall changes surface albedo and isolates ice cover from the atmosphere. We found
12 simulated number of snow days (number of days when the lake has snow cover) is strongly
13 correlated with ice-off date ($r = 0.70$) and maximum ice thickness ($r = 0.75$), indicating
14 snowfall is also significantly related to ice cover.

15 5.1.2 Water temperature and stratification

16 Figure 10 shows that the simulated water temperatures in Lake Mendota at the time of
17 freezing have been increasing since 1911. The correlation between freeze-over water
18 temperature and lake drivers during 1 November and 31 December (N-D) is significant but
19 with low r values for wind speed ($r = -0.37$) and air temperature ($r = -0.24$) and insignificant
20 for Secchi depth ($r = 0.18$). This result indicates that wind mixing is related to cooling of the
21 water column before Lake Mendota freezes.

22 The simulated onset dates of stratification are negatively correlated with April-May (A-M) air
23 temperature ($r = -0.55$) and positively correlated with A-M wind speed ($r = 0.35$). In Lake
24 Mendota, the combination of increasing air temperatures and decreasing wind speeds are both
25 related to earlier onset of stratification. Austin and Colman (2007) suggested that the
26 declining ice cover combined with higher air temperatures cause the earlier onset of
27 stratification in Lake Superior at a rate of 0.5 day yr^{-1} . However, no correlation between ice-
28 off date and the onset of stratification was found for Lake Mendota in this study.

29 Mid-summer (16 July – 15 August) epilimnetic temperatures are most strongly correlated
30 with the air temperature and wind speed averaged over the corresponding period. Epilimnetic
31 temperature is highly correlated with air temperature ($r = 0.77$) and not significantly



1 correlated with wind speed averaged over the corresponding period ($r = -0.11$) and Secchi
2 depth ($r = 0.09$). Mid-summer hypolimnetic temperatures depend on the timing of
3 stratification (how long the near-bottom water is heated before the onset of stratification) and
4 the amount of heat mixed into hypolimnion before and during the stratification period.
5 Hypolimnetic temperatures are weakly correlated with air temperatures ($r = -0.23$) and not
6 significantly correlated with wind speed ($r = 0.11$) and Secchi depth ($r = -0.02$). Summer
7 hypolimnetic heating (temperature change between 1 July and 31 August, J-A) is most
8 strongly correlated with J-A wind speed ($r = 0.49$), moderately correlated with Secchi depth
9 averaged over 30 June – 2 September ($r = 0.35$), and not significantly correlated with J-A air
10 temperature ($r = -0.02$). This indicates a relationship between stronger winds and clearer
11 water and deeper heat mixing.

12 Fall turnover occurs when the lake water cools, driven largely by colder air temperature and
13 wind-induced mixing. The date of fall turnover is significantly negatively correlated with
14 September wind speed ($r = -0.43$) so that higher wind speed is related to earlier fall turnover.
15 Air temperature ($r = 0.11$) and Secchi depth ($r = -0.15$) are not significantly correlated with
16 the timing of fall mixing. Interestingly, the mid-summer epilimnion-hypolimnion temperature
17 difference has a higher correlation with turnover date ($r = 0.62$) than any of the three lake
18 drivers. This finding suggests that the influences of spring conditions may be transmitted to
19 the following summer and fall seasons. In other words, the cooler hypolimnion and the greater
20 epilimnion-hypolimnion temperature difference after a warmer and less windy spring may be
21 related to later fall turnover and a longer stratification period.

22

23 **5.2 Abrupt changes in lake variables**

24 To investigate the effects of abrupt changes in air temperature and wind speed on lake ice
25 cover and water temperatures, we used the hydrodynamic model DYRESM-WQ-I to describe
26 changes in several lake variables during 1911–2014. Simulation results were used to examine
27 differences in mean values of these lake variables between specific periods (see Sect. 4.1).
28 For each period of the selected periods, mean lake variables were calculated and the
29 differences between periods were analysed with t-tests to determine if they were significantly
30 different. Table 1 lists the mean values and differences for the nine lake variables during the
31 three selected periods. Comparison of period 1 (1911–1980) to period 2 (1981–1993) of lake



1 variables shows a shift to warmer air temperature; period 2 to period 3 (1994-2014) represents
2 an abrupt change to lower wind speed; and period 1 to period 3 represents a shift in to warmer
3 air temperature combined with an abrupt change to lower wind speeds.

4 5.2.1 Ice cover

5 Three simulated ice cover variables (maximum ice thickness, ice-on date, and ice-off date)
6 show no significant difference in means between period 1 and 2. In other words, the abrupt
7 change in air temperature trend does not result in a different ice regime even though the ice
8 cover variables are all highly correlated with air temperature ($r > 0.70$). This may be because
9 the change in air temperatures was not of sufficient magnitude to cause a particularly large
10 change in ice cover or it may signify that other drivers are contributing to changes in ice cover
11 variables. Additionally, no significant difference is observed between period 2 and period 3
12 for the ice cover variables since the wind speed and ice cover variables are only weakly
13 correlated. The ice variables do show a statistically significant difference in mean values
14 between periods 1 and 3, indicating that a significant shift in these variables occurs only after
15 a sufficiently large increase in air temperature and an abrupt shift in wind speed within the
16 time between periods 1 and 3. In other words, air temperature need to increase sufficiently to
17 observe a statistically significant difference in ice cover. Ice cover duration, however, shows a
18 significant difference in the mean between all regimes (1-2, 2-3, and 1-3), indicating that
19 distinct differences in ice cover duration can be affected by both trends in air temperature
20 (i.e., there was a large enough change in air temperature between each period) and an abrupt
21 shift in the wind speed. The combined effects of slightly later ice-on dates and earlier ice-off
22 dates during each of the three periods resulted in statistically significant difference in mean
23 ice cover duration values between each of the three periods.

24 Analysis of simulated maximum ice thickness, ice-on, ice-off, and ice cover duration using
25 the method of Rodionov (2004) shows that the most statistically significant timing of the shift
26 in these ice cover variables occurs in the winter of 1997-1998, but a major shift in the air
27 temperature or wind speed data was not observed at that time. The unusual winter of 1997-
28 1998 strongly drove the statistically significant difference in mean values between periods 1
29 and 3 rather than the abrupt shift in wind speed in 1994. Interestingly, similar results have
30 been reported in Lake Superior, where statistically significant step changes were found in
31 winter ice duration and maximum wintertime ice extent; these step changes account for most
32 of the long-term trends in ice cover for the lake (Van Cleave et al, 2014). This timing of this



1 may be attributed to a combination of the longer term changes in meteorological conditions
2 and the short-term annual change occurring in the warm El Niño winter of 1997-1998 (Van
3 Cleave et al. 2014). Mueller et al. (2009) found that a similar climate shift between 1997 and
4 1998 initiated a change in lake ice phenology from infrequent to frequent summer loss in
5 several high-Arctic lakes.

6 5.2.2 Water temperature and stratification

7 Means of five simulated lake variables (under-ice water temperature, freeze-over water
8 temperature, epilimnion-hypolimnion temperature difference (indicative of strength of
9 stratification), and duration of stratification) over the three periods have significant ($p < 0.05$)
10 differences only between period 1 and period 3. This change likely occurs because of the
11 combined effects of large changes in air temperature and a change in wind speed. Both air
12 temperature and wind speed are significantly correlated with these five lake variables. Each
13 driver alone may not be strong enough to cause a major shift in the lake variables, but their
14 combined effects may reinforce the drivers of abrupt change in ice and thermal phenology.
15 Further work is required to examine how the major drivers may either reinforce or dampen
16 lake ice and temperature responses, particularly in relation to directional shifts predicted
17 under climate change.

18 Fall turnover date, highly correlated with wind speed, exhibits a significant ($p < 0.05$) shift in
19 the mean value in 1994, corresponding with the abrupt shift in the wind speed. Interestingly,
20 hypolimnetic water temperatures, which are not significantly correlated with wind speed, but
21 are correlated with air temperatures, also show a significant ($p < 0.05$) shift in the mean value
22 in 1994. Hypolimnetic heating, significantly correlated with wind speed ($r = 0.49$), does not
23 exhibit a significant breakpoint, nor are any of the mean differences among the three periods
24 significant. Given the high correlation between wind speed and hypolimnetic heating, it is
25 hypothesized that there should be a shift in hypolimnetic heating caused by the abrupt shift in
26 wind speed in 1994. The lack of statistically significant step change may be explained by the
27 simultaneous high correlation between Secchi depth and hypolimnetic heating ($r = 0.35$),
28 indicating that water clarity may act to inhibit heating regardless of changes in wind speed, or
29 it may be acting to filter or mitigate the effects of the wind speed shift. Finally, mean onset
30 date of stratification and mid-summer epilimnetic temperature exhibit no difference among
31 the three periods. This may be due to two processes: (i) the climate signal is being filtered out



1 the lake; or (ii) the external perturbation of the system is not yet strong enough to trigger a
2 major shift in the system's internal dynamics.

3 **5.3 Ecological significance of long-term changes in lake variables**

4 Changes in lake ice cover and thermal structure are of great ecological significance. Summer
5 stratification inhibits the vertical transport of oxygen and nutrients. The vertical temperature
6 gradient in Lake Mendota increased (Fig. 9b), which should decrease the depth of vertical
7 mixing and oxygen penetration and decrease the transport of nutrients from the hypolimnion
8 to the epilimnion (Fee et al. 1994). Thus, increases in air temperature may increase the
9 epilimnetic-hypolimnetic temperature gradient and decrease the extent of summer algal
10 blooms in the Lake Mendota where internal loading of phosphorus has been shown to be
11 significant to its summer phosphorus budget (Soranno et al. 1997; Lathrop et al. 1998).
12 Additionally, phosphorus entrainment predominately occurs during periods of high wind
13 (Stauffer and Armstrong 1986; Kamarainen et al. 2009). Algal blooms have been associated
14 with increases in epilimnetic phosphorus due to these high-wind episodes (Stauffer and
15 Lee 1973; Soranno et al. 1997; Robarts et al. 1998). Changes in the frequency of high wind
16 episodic events may further impact the growth of algal blooms within Lake Mendota (Kara et
17 al. 2012). Elevated water temperatures can affect plankton community composition and
18 abundance (Elliot et al. 2005; Findlay et al. 2001; Francis et al. 2014; Rice, 2015) and fish
19 populations (Magnuson et al. 1990; Carpenter et al. 1992; Destasio et al. 1996; Gunn 2002).
20 High water temperatures have been also shown to enhance the dominance of cyanobacteria
21 (Huber et al. 2008; Johnk et al. 2008). Such changes in algal blooms, zooplankton
22 populations, and fish populations may have drastic effects on the overall food-web within
23 Lake Mendota (Carpenter et al. 2007). Variations in ice-cover conditions have impacts on
24 ecosystems by dramatically changing habitat factors such as light and temperature (Adrian et
25 al. 1999; Quayle et al. 2002). Such changes impact the distribution, behavior, community
26 composition, reproduction, and evolutionary adaptations of organisms (Sala et al. 2000).
27 Previous studies have shown that climate change will likely affect the biodiversity of
28 freshwater ecosystems (Wrona et al. 2006; Heino et al. 2009; Mantyka-Pringle et al. 2014).
29 Climate change can bring asynchronies in biotic communities that may result in a
30 phenological decoupling of trophic relationships (Stenseth and Myrsetrud, 2002).

31



1 6 Conclusions

2 The 1-D hydrodynamic-ice model, DYRESM-WQ-I, is developed and validated, then used to
3 simulate the ice cover and thermal structure of Lake Mendota, WI from 1911–2014. The
4 model reliably reproduces the interannual variations and long-term (104-year) changes in ice
5 cover and thermal structure. To our knowledge, this study presents the first attempt to
6 continuously model both ice cover and thermal structure of a lake over a period as long as a
7 century. Simulated ice cover over this period has changed dramatically: freezing later (9.0
8 days per century), breaking up earlier (12.3 days per century), resulting in shorter ice duration
9 (21.3 days per century), and thinner maximum ice thickness (12.7 cm per century). These
10 results agree well with the observed data and previous studies. For lake thermal structure,
11 results of the continuous model simulation indicate an earlier onset of thermal stratification
12 (12.3 days per century), later fall turnover (14.6 days per century), and a longer stratification
13 period (26.8 days per century) during this period. In addition, simulated mid-summer
14 hypolimnetic temperature decreased (-1.4°C per century), causing mid-summer epilimnetic-
15 hypolimnetic temperature differences (strength of stratification) to increase (2.0°C per
16 century). Interestingly, there is no significant trend in mid-summer epilimnetic temperatures.

17 Correlations comparing simulated lake conditions with seasonally averaged meteorological
18 conditions indicate that, among the three drivers, air temperature has the largest relationship
19 with the ice cover variables (ice-on and ice-off dates, and maximum ice thickness) and with
20 three stratification variables (date of the onset of stratification, epilimnetic temperatures, and
21 hypolimnetic temperatures). Freeze-over water temperature, summer hypolimnetic heating,
22 and data of fall turnover all have the largest correlation with wind speed. Both air temperature
23 and wind speeds are highly correlated with the onset of stratification. Secchi depth does not
24 appear to have a strong relationship with interannual variability of these variables for Lake
25 Mendota, but in combination with wind speed may be related to hypolimnetic heating.

26 Changes in meteorological factors over the past 104 years were examined to determine if
27 there have been abrupt shifts, rather than linear changes. Based on a change in the trend of air
28 temperature increase occurring in 1981 and a major shift in wind speed in 1994, the Madison
29 climate is divided into three distinct periods: 1911–1980, with relatively low air temperatures
30 and mean wind speeds of 4.44 m s^{-1} ; 1981–1993, with higher air temperatures and mean wind
31 speeds of 4.44 m s^{-1} ; and 1994–2014 with still higher air temperatures and mean wind speed
32 of 3.74 m s^{-1} . Ice cover duration exhibited a significant difference in the mean among all three



1 periods, while ice-on, ice-off, and maximum ice thickness only show a significant difference
2 between period one and three, indicating that only with a large change in air temperature and
3 an abrupt shift in wind speeds are change in the ice cover variables statistically different. Mid-
4 summer hypolimnetic temperature and fall turnover date both reveal significant ($p < 0.05$)
5 differences in the mean value in 1994, corresponding with the abrupt shift toward lower wind
6 speeds. Some lake variables (under-ice water temperature, freeze-over water temperature,
7 epilimnion-hypolimnion temperature difference, and stratification duration) may not be driven
8 by either the change in air temperature trend or the abrupt shift in wind speed alone, but a
9 shift in the mean of the lake variables does occur in 1994 when both the air temperatures are
10 warmest and the wind speed experienced an abrupt shift. The exact timing of shifts may be
11 difficult to define because of extreme changes in weather in specific years and it may mask
12 the longer term changes in meteorological conditions (i.e. abrupt shifts).

13 Linear trends from the extensive data collection may be misleading as they masks these
14 sudden shifts in the air temperature and wind speed drivers and lake ice cover and thermal
15 structure variables. Examining the mean differences in lake variables in response to shifts in
16 air temperature and abrupt shifts in wind speeds provide a better understanding of how the
17 Madison Area Wisconsin climate has changed and what impact those changes have on shifts
18 in the ice cover and water temperature of dimictic Lake Mendota. It is shown that air
19 temperature and wind speed changes have occurred in stages and lake variables have
20 responded in a non-linear way to these changes.

21

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30



- 1 Table 1. Mean values of climate drivers and lake variables of three hypothesized periods during 1911–2014. Asterisks (*) mark significant
 2 differences between two period ($p < 0.05$).

Driver/Variable	Unit	Period			Difference in Mean		
		(1)1911-1980	(2)1981-1993	(3)1994-2014	(1) and (2)	(2) and (3)	(1) and (3)
Lake driver							
Air temperature (slope)	°C decade ⁻¹	0.081	0.334	0.334			
Wind speed	m s ⁻¹	4.44	4.44	3.74			
Lake variables							
Maximum ice thickness	cm	49.8	44.9	40.7	-4.9	-4.2	-9.1*
Ice-on date (model)	Date	21-Dec	23-Dec	29-Dec	2 days	6 days	8 days*
(observation)	Date	21-Dec	24-Dec	29-Dec	3 days	5 days	8 days*
Ice-off date (model)	Date	9-Apr	2-Apr	30-Mar	-7 days	-3 days	-10 days*
(observation)	Date	3-Apr	27-Mar	26-Mar	-7 days	-1 days	-8 days*
Ice duration (model)	Days	108.7	99.5	91.1	-9.2*	-8.4*	-17.6*
(observation)	Days	103.2	92.9	85.6	-10.3*	-7.3*	-17.6*
Under-ice water temperature	°C	1.74	1.81	2.08	0.07	0.27	0.34*
Freeze-over water temperature	°C	1.03	1.14	1.66	0.11	0.52	0.63*
Stratification onset date		24-May	17-May	18-May	-7 days	-1 days	-6 days
Mid-summer epilimnetic temperature	°C	23.0	23.2	23.4	0.2	0.2	0.4
Mid-summer hypolimnetic temperature	°C	12.0	11.8	10.9	-0.2	-0.9*	-1.1*
Epilimnion-hypolimnion temp. difference	°C	10.9	11.4	12.5	0.5	1.1	1.6*
Hypolimnetic heating (1 July–31 August)	°C	0.699	0.688	0.583	-0.011	-0.105	-0.116
Turnover date	Date	20-Sept	21-Sept	3-Oct	1 day	12 days*	13 days*
Stratification Duration	Days	119.4	127.4	138.6	8.0	11.2	19.2*

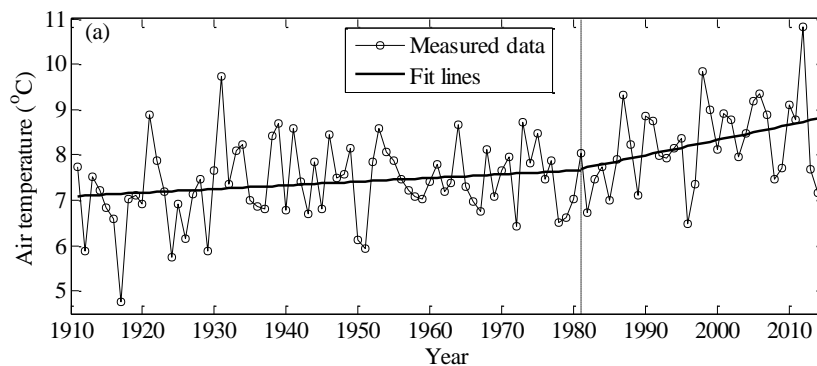


- 1 Table 2. Averaging periods for lake drivers used for calculating correlation coefficients of
2 each pair of variables.

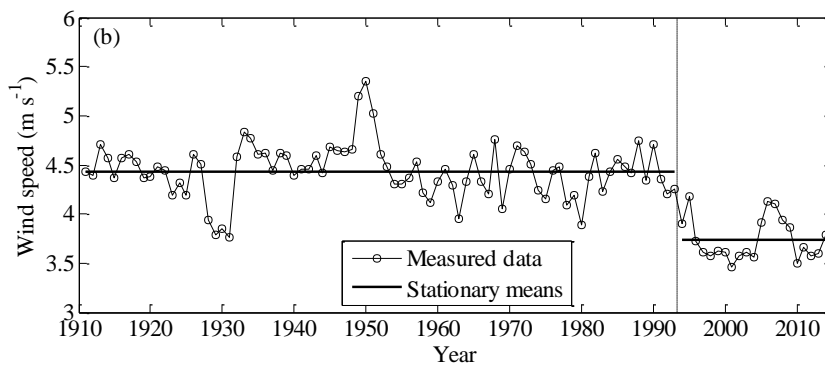
Lake variables	Air temperature/wind speed	Secchi depth ¹
Ice-on date	November – December	13 October – ice-on
Ice-off date	February – March	ice-on – ice-off
Max. ice thickness	January – March	ice-on – ice-off
Freeze-over water temperature	November – December	13 October – ice-on
Date of stratification onset	April – May	ice-out – 10 May
Epilimnetic temperature	16 July – 15 August	30 June – 2 September
Hypolimnetic temperature	May	ice-out – 30 June
Hypolimnetic heating	July – August	30 June – 2 September
Date of fall turnover	September	3 September – 12 October

¹Six seasonal periods defined in Lathrop et al. (1996)

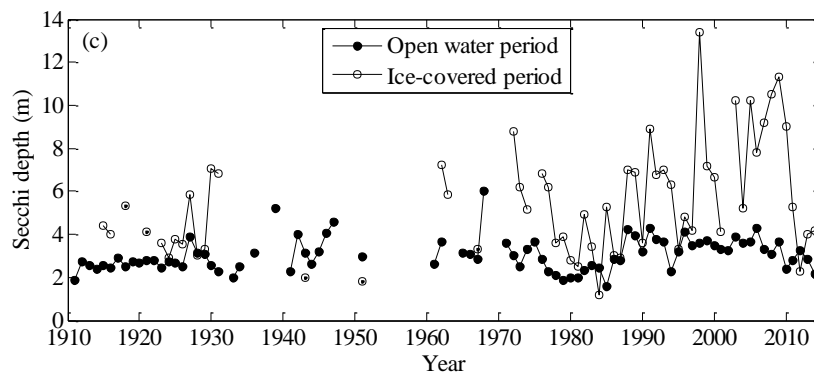
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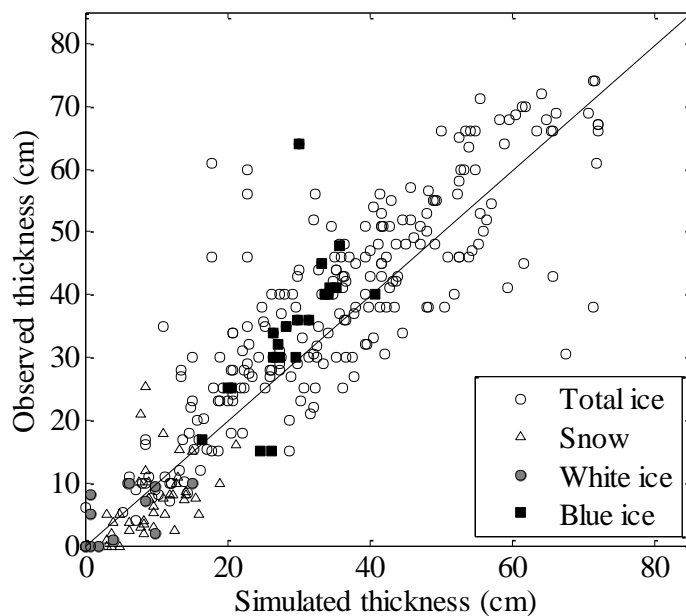
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4 Figure 1: Historical record of annual average (a) air temperature and (b) wind speed in
5 Madison, WI, USA, and (c) Secchi depth in Lake Mendota.

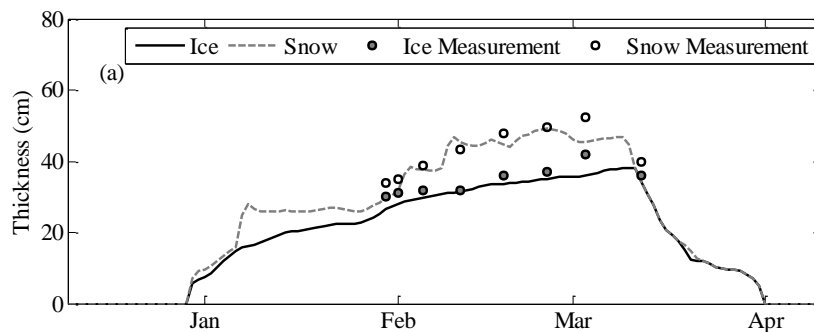
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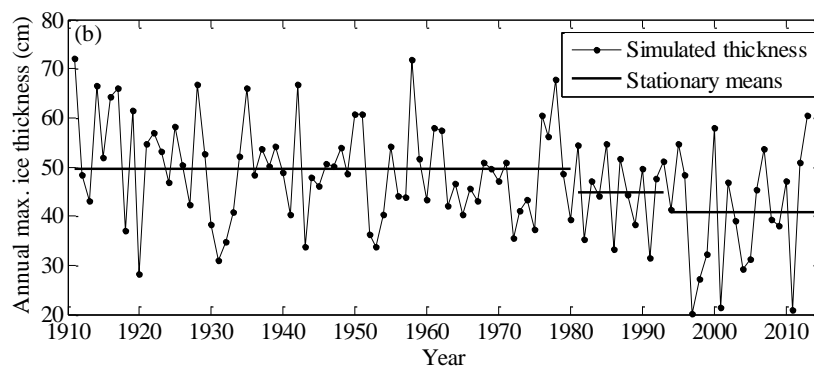
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2 Figure 2. Comparison between observations and simulated total ice, snow cover, white ice,
3 and blue ice

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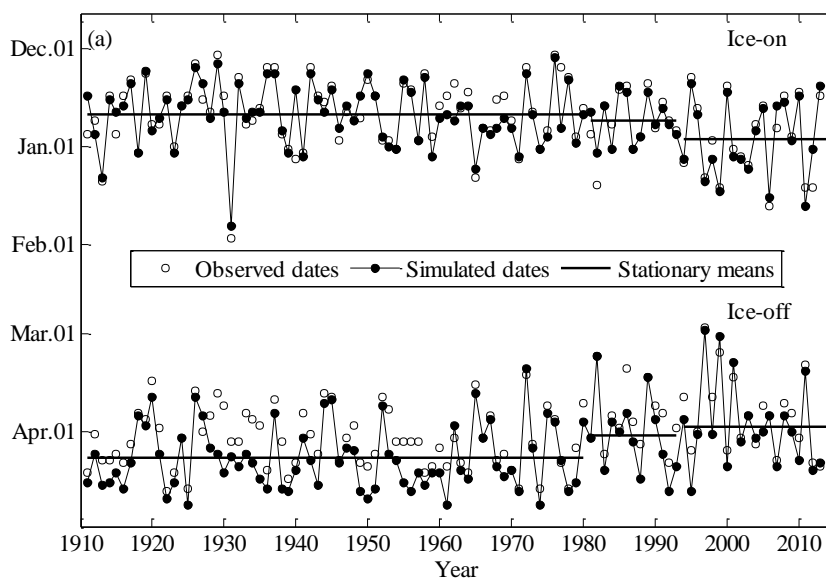
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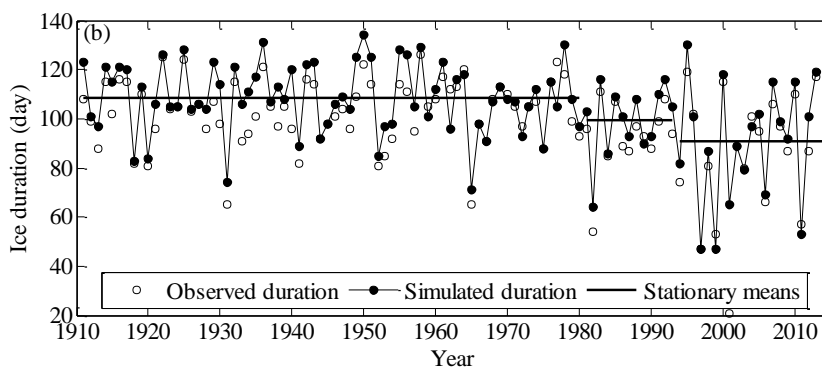
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3 Figure 3: (a) simulated ice and snow with measured values for the winter 2009-2010 and (b)
4 simulated annual maximum total ice thickness.

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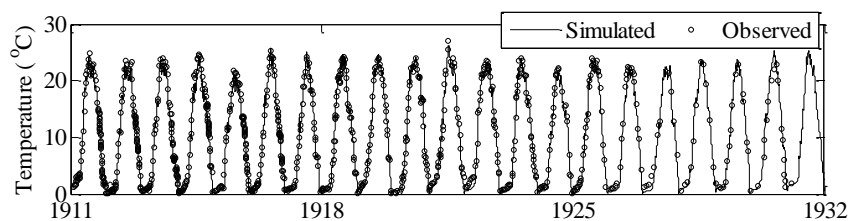
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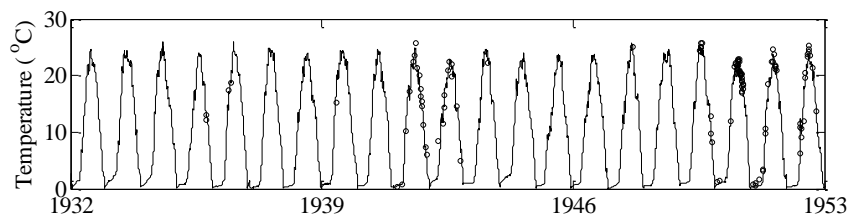
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3 Figure 4: Observed and simulated (a) ice-on and ice-off dates and (b) ice cover duration. The
4 annual ice cover was plotted in the first year of the winter, e.g., the winter 1911-1912 was
5 plotted in the year 1911. Stationary means for three selected periods are denoted by solid
6 lines.

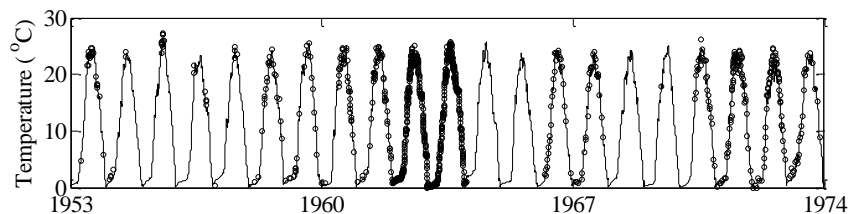
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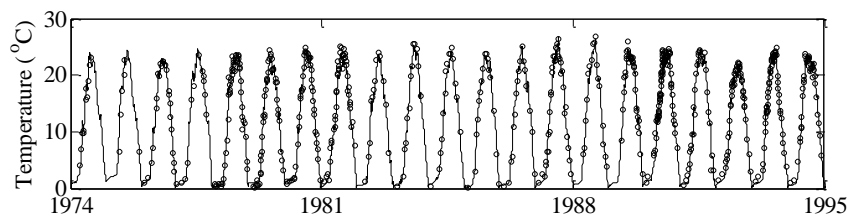
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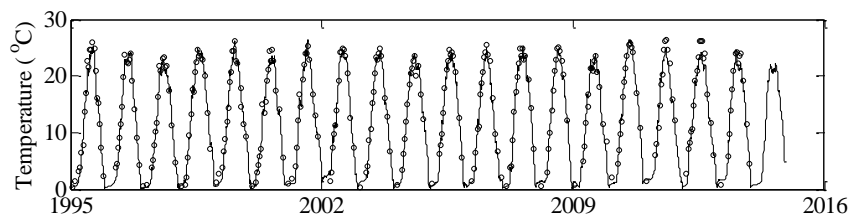
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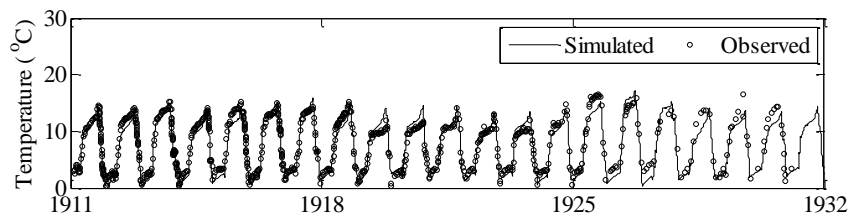
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6 Figure 5: Time series of the simulated and observed volume-weighted near-surface
7 (epilimnetic) temperature (0–10m) from 1911–2014.

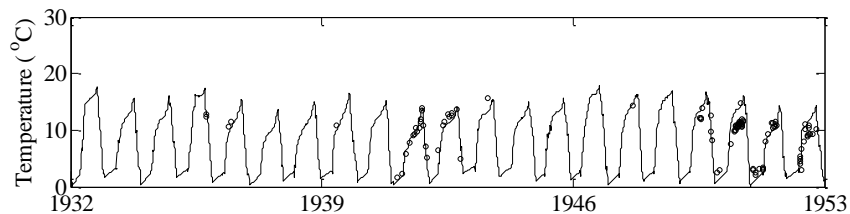
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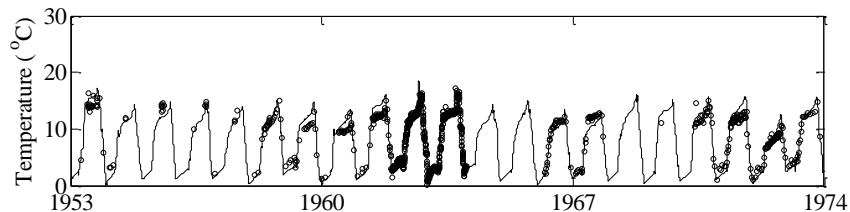
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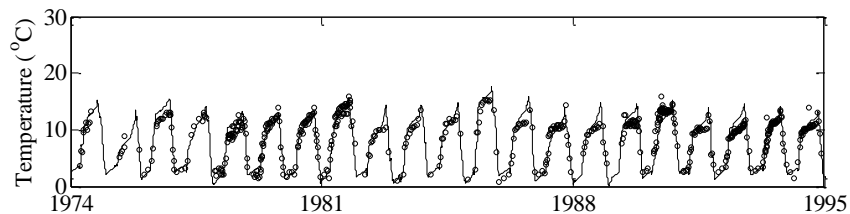
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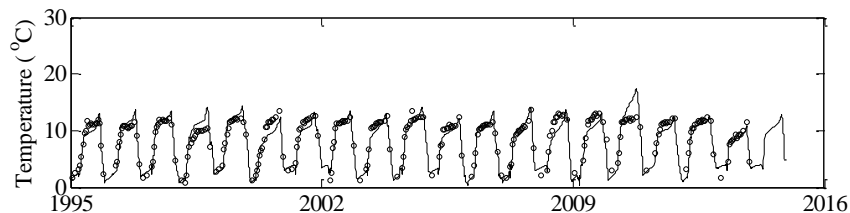
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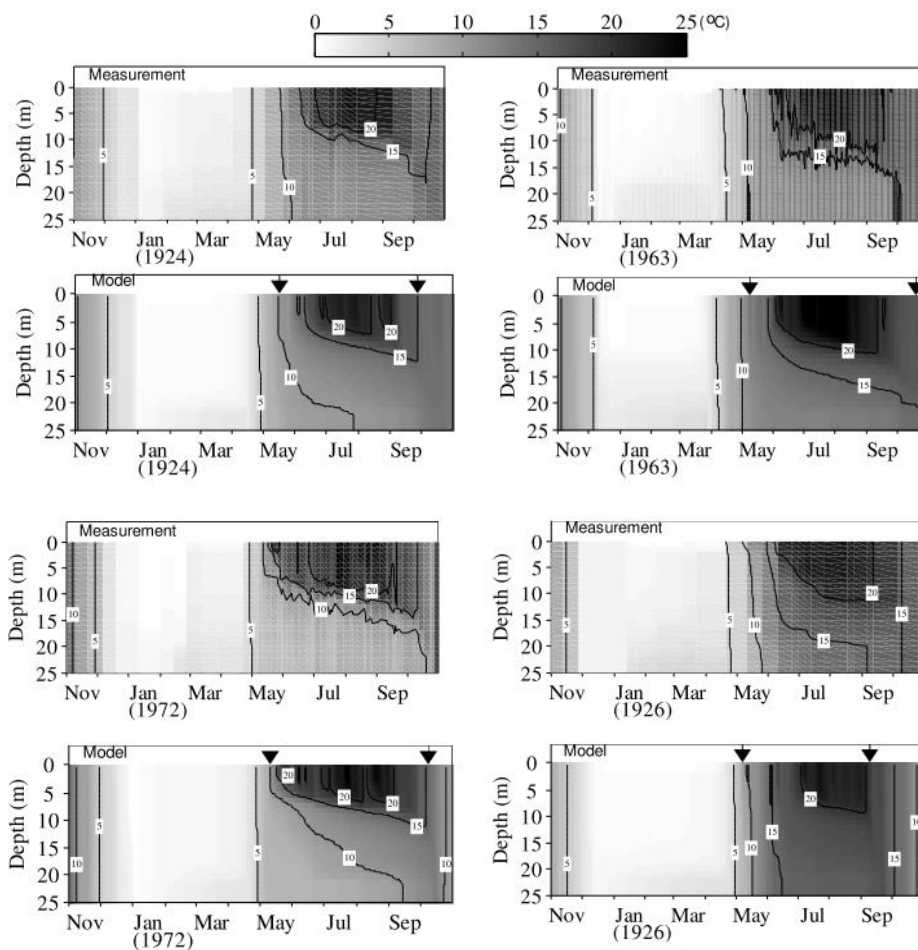


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6 Figure 6: Time series of the simulated and observed volume-weighted near-bottom
7 (hypolimnetic) temperatures (20–25 m) from 1911–2014.

8

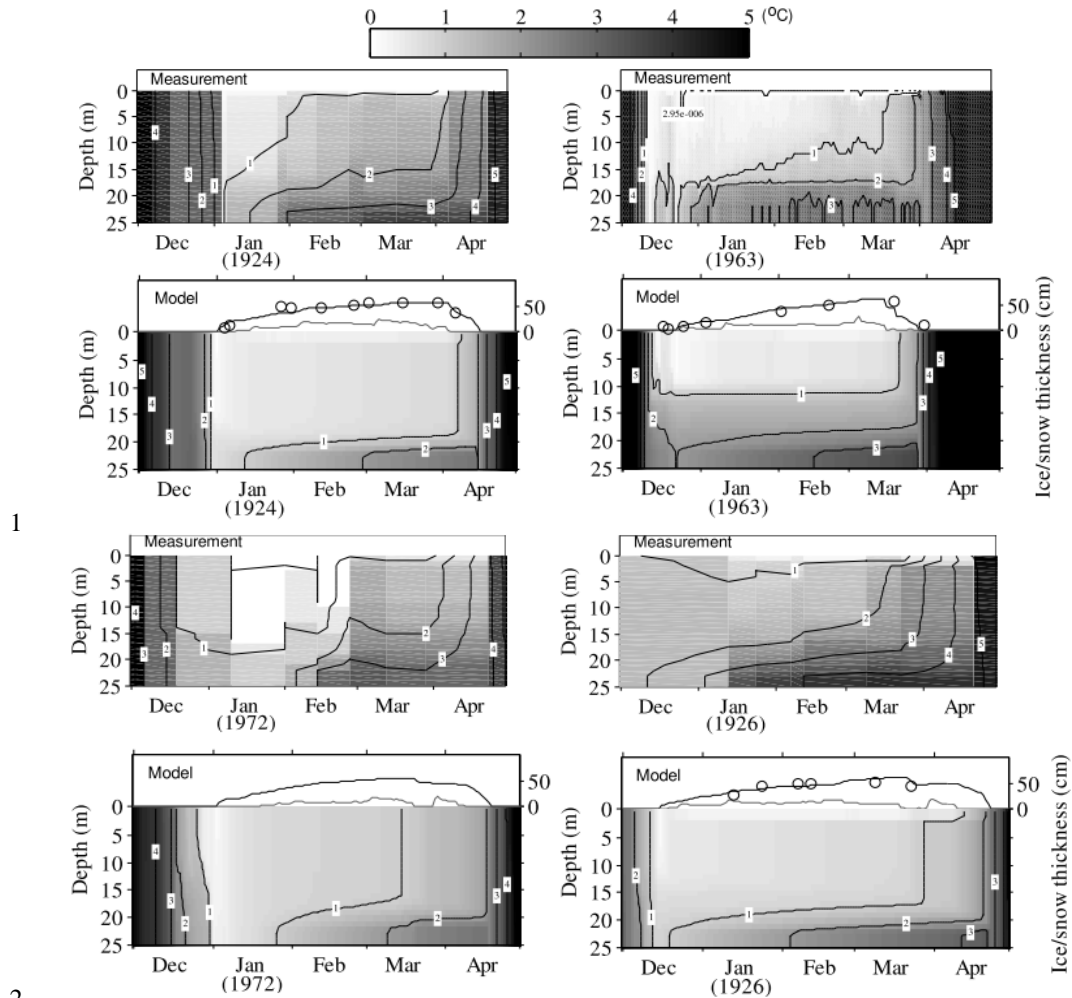


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3 Figure 7: Comparison between the measured and simulated water temperatures as a function
 4 of time and depth under the cold epilimnetic year (1924), warm epilimnetic year (1963), cold
 5 hypolimnetic year (1972), and warm hypolimnetic year (1926) in Lake Mendota. Arrows
 6 mark the dates of onset of stratification and fall turnover.

7

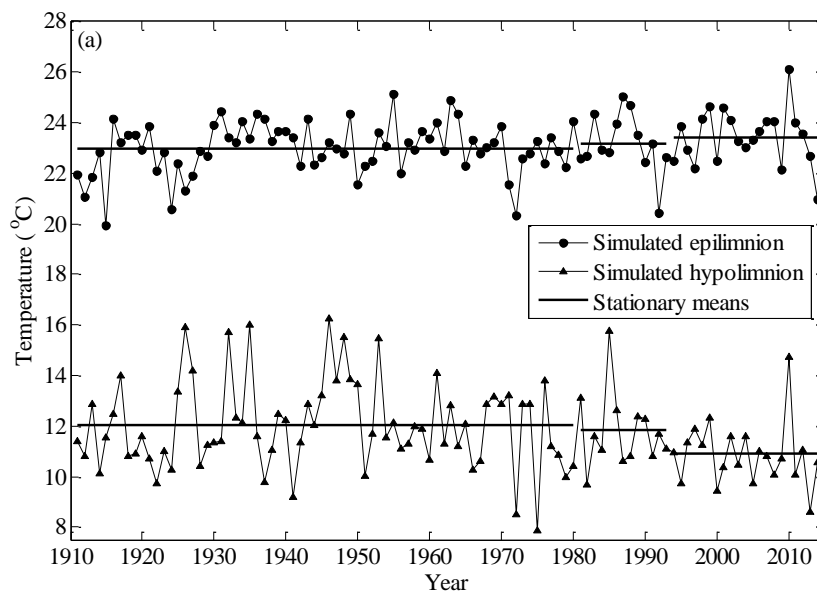


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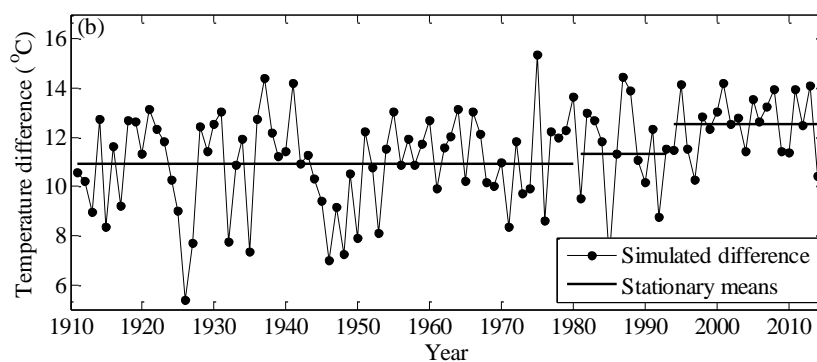
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3 Figure 8: Comparison between the measured and simulated water temperatures and ice/snow
 4 thickness under the cold epilimnetic year (1924), warm epilimnetic year (1963), cold
 5 hypolimnetic year (1972), and warm hypolimnetic year (1926) in Lake Mendota during the
 6 ice-covered period. Time series are simulated ice and snow thickness (solid lines) with
 7 measurements (open circles).

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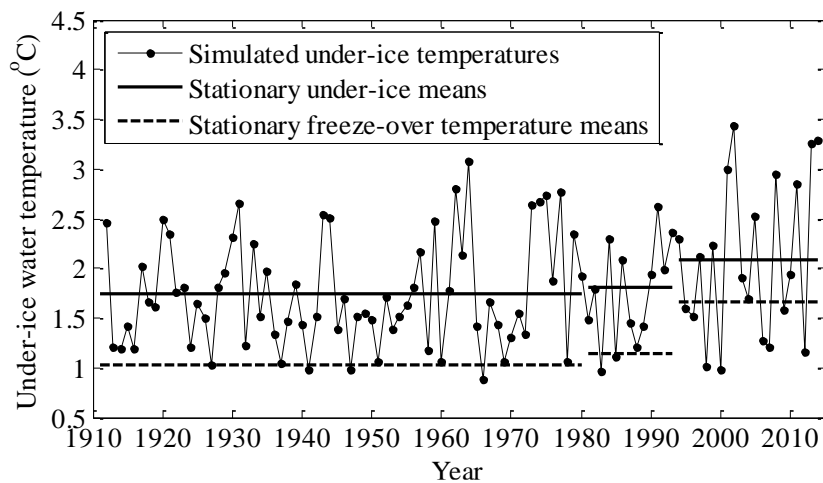
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3 Figure 9: (a) Simulated mid-summer (16 July – 15 August) volume-averaged epilimnion and
4 hypolimnion temperatures, and (b) epilimnion-hypolimnion temperature difference.
5 Stationary means for three selected periods are denoted by solid lines

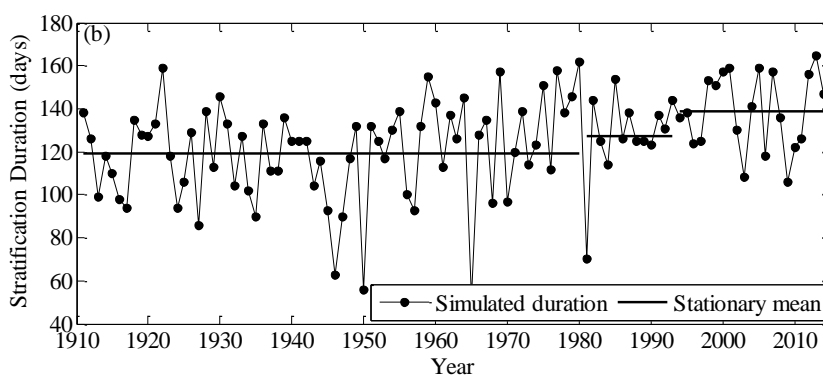
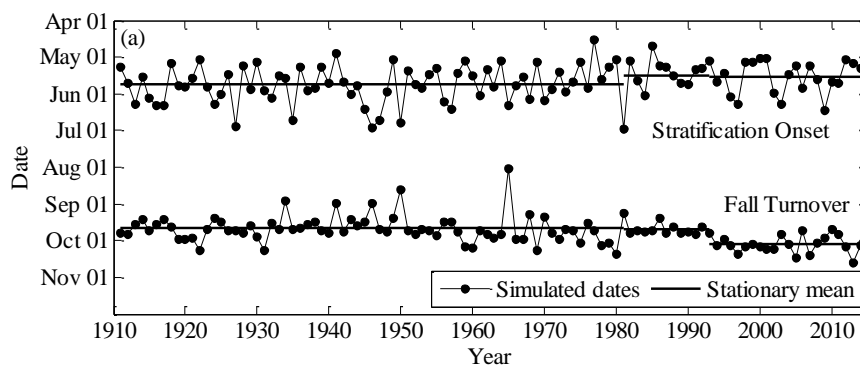
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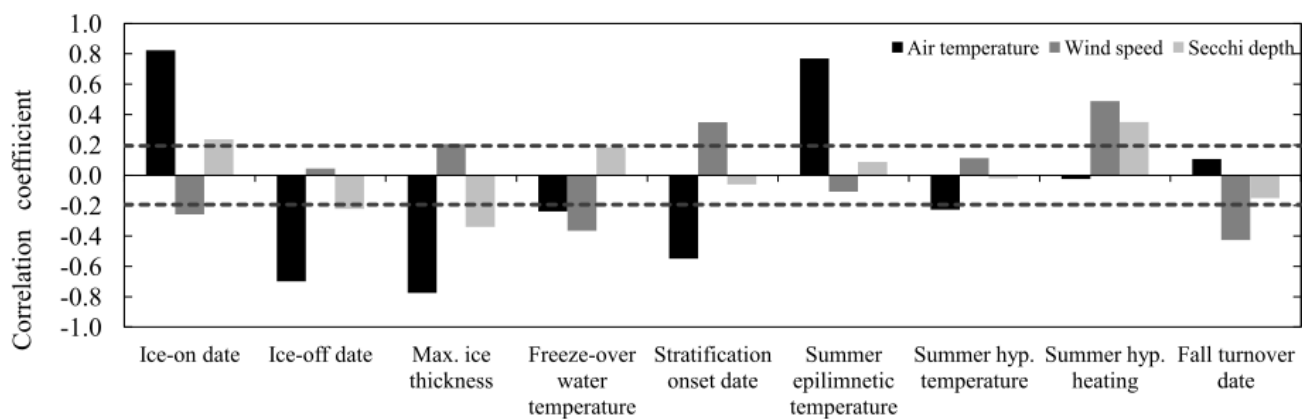
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2 Figure 10: Simulated mean under-ice water temperatures (volume-weighted averages of all
3 depths) during the ice-covered period between 1911 and 2014. The stationary means for three
4 selected periods of under-ice water temperatures and freeze-over water temperatures are also
5 plotted.

6



3 Figure 11: (a) Simulated date of stratification onset and fall turnover and (b) stratification
4 duration. Stationary means for three selected periods are denoted by solid lines.



1

2 Figure 12: Correlation coefficients between lake variables and drivers. The critical value (dashed lines) for significant correlation ($p < 0.05$)

3 0.193 ($n = 104$).