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Spatial analysis of ice phenology trends across the Laurentian Great Lakes region during a recent warming period

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Abstract

We examined spatial patterns of trends in ice phenology and duration for 65 waterbodies across the Great Lakes region (Minnesota, Wisconsin, Michigan, Ontario, and New York) during a recent period of rapid climate warming (1975–2004). Average rates of change in freeze (3.3 d decade^{-1}) and breakup (-2.1 d decade^{-1}) dates were 5.8 and 3.3 times more rapid, respectively, than historical rates (1846–1995) for Northern Hemisphere waterbodies. Average ice duration decreased by 5.3 d decade^{-1}. Over the same time period, average fall through spring temperatures in this region increased by 0.7°C decade^{-1}, while the average number of days with snow decreased by 5.0 d decade^{-1}, and the average snow depth on those days decreased by 1.7 cm decade^{-1}. Breakup date and ice duration trends varied over the study area, with faster changes occurring in the southwest. Trends for each site were compared to static waterbody characteristics and meteorological variables and their trends. The trend toward later freeze date was stronger in large, low-elevation waterbodies; however, freeze date trends had no geographic patterns or relationships to meteorological variables. Variability in the strength of trends toward earlier breakup was partially explained by spatial differences in the rate of change in the number of days with snow cover, mean snow depth, air temperature (warmer locations showed stronger trends), and rate of change in air temperature. Differences in ice duration trends were explained best by a combination of elevation and the local rate of change in either temperature or the number of days with snow cover.

The formation and breakup of ice are important seasonal events in mid- to high-latitude lakes and rivers. The timing of these events, ice phenology, is sensitive to the characteristics of individual waterbodies (Assel and Herche 1998, 2000) and to broader-scale weather patterns and climate variability (Palecki and Barry 1986; Assel and Robertson 1995). For lakes and rivers throughout the Northern Hemisphere, during the period ranging from 1846 to 1995, ice formation has occurred 0.57 d later per decade, while breakup occurred 0.63 d earlier per decade (Magnuson et al. [2000] and errata, with a revised trend of 0.0 d decade^{-1} for Detroit Lake). Trends toward later freeze and earlier breakup have been observed over other time periods in many regions including Ontario (Schindler et al. 1990; Futter 2003), New England (Hodgkins et al. 2002, 2005), Lake Baikal (Livingstone 1999), and northern Europe (Livingstone 1997; Yoo and D'Odorico 2002; Korhonen 2006). While the overall pattern is clear, these waterbodies exhibit considerable variation in terms of their changes in ice phenology. As climate warming continues (IPCC 2001), understanding this variability will become

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increasingly important if we are to anticipate changes in specific waterbodies.

Changes in ice phenology may have important consequences for fish and zooplankton communities. Recent work by Winder and Schindler (2004*a*,*b*) in Lake Washington has demonstrated that earlier ice breakup and stratification has created a temporal mismatch between the peak spring phytoplankton bloom (the timing of which is closely tied to stratification) and the population dynamics of some species of zooplankton (influenced by water temperature and food availability as well as stratification). Schindler et al. (2005) found that earlier ice breakup was associated with higher summer *Daphnia* densities and higher growth rates of planktivorous juvenile sockeye salmon. The effect of climate warming and ice breakup date on zooplankton may depend on the life-history characteristics of individual species (Adrian et al. 2006). Reductions in the duration of ice cover are also likely to reduce or even eliminate winter-kill in shallow eutrophic lakes (Stefan et al. 2001).

Individual ice phenology records show considerable year-to-year variation. For this reason, records typically have been combined and analyzed as a group. Several recent analyses, however, consider spatial variability in climate change and its effects on the ice phenology of individual lakes in different locations. For example, comparisons of ice breakup dates for Canadian lakes from 1951 to 2000 revealed generally stronger trends toward earlier breakup in western Canada (Duguay et al. 2006). This result was consistent with patterns of change in the 0°C isotherm (Bonsal and Prowse 2003). A similar pattern of more rapid change in western Canada occurred for the timing of spring freshets and river ice breakup (Zhang et al. 2001). Korhonen (2006) identified north–south differences in ice phenology trends for Finnish lakes from 1885 to 2002. Freeze and breakup dates over this time period changed most rapidly in southern Finnish lakes; however, the spatial trend was not monotonic; ice phenology in northern lakes changed more rapidly than in central lakes.

Ice formation and breakup are fundamentally different processes, and we may expect variation in their trends to be driven by different factors. In stratified lakes, ice formation is generally preceded by heat loss and turnover. Therefore, lake-specific factors that influence the amount of heat in the lake (e.g., lake volume) and the rate at which heat is lost to cooler air (e.g., depth, surface area and exposure to wind; Hutchinson 1957) affect the timing of ice formation (Assel and Herche 1998, 2000). In large lakes, the depth of convective mixing can be much shallower than the maximum or mean depth, allowing ice formation to occur despite the existence of permanently stratified deep water (e.g., Lake Baikal; Wuest et al. 2005).

Breakup in lakes is dominated by external drivers (e.g., air temperature and solar radiation) (Vavrus et al. 1996) and, like many lake temperature variables (Benson et al. 2000*b*), is coherent at spatial scales of hundreds of kilometers (Magnuson et al. 2005). An empirical study of factors influencing freeze and breakup dates in 128 lakes in North America found that lake morphometry variables (surface area and mean depth) were consistently less important predictors than were air temperature, latitude, or elevation (Williams and Stefan 2006). Because snow changes the albedo of the surface and insulates ice from the surrounding air, snow also plays a role in determining breakup date (Vavrus et al. 1996). Breakup in rivers can proceed via two different mechanisms: thermal breakup, in which the ice melts in place (as in lakes), or dynamic breakup, in which large sheets of intact ice are displaced by hydraulic pressure (Prowse and Culp 2003). Both processes are influenced by regional climate and waterbody-specific characteristics, such as gradient and geographic direction of river flow.

Although much is known about the factors that influence ice phenology, relatively little work has succeeded in explaining spatial or among-lake variation in ice phenology trends (but *see* Duguay et al. [2006]). Our study focuses on variation in ice phenology and duration trends for a group of 65 waterbodies located across the Great Lakes region (Minnesota, Wisconsin, Michigan, Ontario, and New York) from 1975 through 2004. The time period was chosen because it is known to represent a period of rapid warming (IPCC 2001), making trends and potential differences among waterbodies more pronounced. Using statistical comparisons of ice phenology and climate time series, we address the following questions: (1) How does ice phenology respond during rapid warming?; (2) Are there spatial patterns in the response of ice phenology to climate warming?; and (3) How well can among-waterbody variability in trends in ice phenology be explained by a combination of individual waterbody characteristics and meteorological variables (temperature, days with snow cover, and average snow depth) and their trends?

Study areas and methods

Data—We analyzed ice phenology records from 65 waterbodies in Minnesota, Wisconsin, Michigan, Ontario, and New York (Table 1). These records include 62 lakes, 1 bay of Lake Superior, 1 river, and 1 record representing the average ice breakup dates for 8 lakes (Blue Chalk, Chub, Crosson, Dickie, Harp, Heney, Plastic, and Red Chalk Lakes) near Dorset, Ontario (Canada), for which only the mean data are available. Of the 65 waterbodies, 33 had freeze date records, 64 had breakup date records, and 32 had records of ice duration. All records included in the analysis had no more than three missing years during the 30-yr time period from 1975 to 2004 (years refer to the end year of a winter: e.g., 1975 is the winter of 1974–1975). Previous publications have examined the trends and coherence in ice phenology of some of these waterbodies (e.g., Benson et al. 2000*a*; Futter 2003; Magnuson et al. 2005) but have not analyzed their spatial patterns or relationships with meteorological variables.

Five lakes did not freeze in 2002, and one of these did not freeze in 1998. For ice duration, the quantification of these no-ice years is obvious, but for freeze and breakup dates, any approach is to some extent arbitrary. Removing these warmer years would clearly bias the results. Assel and Robertson (1995) inferred freeze and breakup dates for such situations by taking the average midpoint between the freeze and breakup dates of the five winters with the shortest ice durations. We followed their approach, but for comparison we also calculated trends using two other methods for handling the no-ice years: (1) we used the latest observed freeze date and the earliest observed breakup date, and (2) we dropped no-ice years from these records (an extremely conservative approach).

We used meteorological records from weather stations in the National Climatic Data Center (U.S.A.; http://www.ncdc.noaa.gov/oa/ncdc.html) and the National Climate Archive (Canada; http://climate.weatheroffice.ec.gc.ca). Daily records of average, minimum, and maximum temperatures were averaged for each year and station over the period ranging from 01 September through 31 May. This fall-to-spring time period covers the entire span of freeze and breakup date records. Two snow depth variables, the number of days with snow depth >0 (snow days) and the average snow depth for these days with snow, were calculated. We chose to summarize snow depth using these two variables rather than calculating a simple mean snow depth over a fixed time period for three reasons. First, using these two variables allows us to separate changes in the duration of the snow cover period from changes in snow depth during that period. These two aspects of snow cover are confounded in a simple mean. Second, using a simple mean requires defining a time interval a priori, as was done for temperature. This is not ideal, because a time interval long enough to cover the relevant period (preceding freeze through the end of breakup) in the north is longer than the relevant period in the south. Third, averaging snow depth over a time period with many snow-free days creates a zero-inflated distribution that violates the assumptions of many statistical procedures (Martin et al. 2005). All weather station records from states and provinces in the study area were used provided that they contained no more than 10 missing days during the fall-to-spring period or three missing years.

Analysis methods—Temporal trends were calculated as slope parameters from linear regressions (SAS v9.1, REG procedure) of freeze date, breakup date, ice duration, snow days, snow depth, and seasonal averages of daily minimum, maximum, and average temperatures on year. Trends in ice phenology and duration for each of the study waterbodies are described in Table 1. To test the significance of changes in these variables for all waterbodies or weather stations combined, an analysis of covariance (SAS v9.1, GLM procedure) was conducted using the ice or meteorological variable as the response, the year as a continuous explanatory variable, and the waterbody or weather station as a class explanatory variable. Interactions between these two explanatory variables were also tested. To examine spatial patterns in the temporal trends, the slope parameters from the linear regressions were used as the response variables in multiple linear regressions, with the X and Y coordinates of the waterbodies or weather stations and their interaction ($X \times Y$) as explanatory variables.

Weather stations were not at the exact locations of the waterbodies. Therefore, we interpolated the meteorological data using kriging (ArcView v8.3, Geostatistical Analyst Extension). Station and waterbody locations were first projected to North American Lambert Conformal Conic. First order spatial (X and Y) trends were removed prior to variogram fitting (spherical model) and kriging. Elevation was not used in the interpolation, as it accounted for only a small portion of the variability in meteorological variables (*see Results*). Ninety percent of waterbodies were within 2 km and 54 km of the nearest weather station. The maximum distance between a waterbody and the nearest weather station was 87 km, and the average distance was 22 km. Of the five meteorological variables, two (maximum daily temperature and mean snow depth) showed no significant spatial pattern in their temporal trends. Thus, trends in these two variables were not interpolated and were not used in the regression models of ice phenology or duration trends (described below).

Breakup date isophenes (contour lines connecting locations with the same breakup date) were created for 15 April. This date was chosen because isophenes for this date generally fell near the approximate center of the north–south distribution of the waterbodies studied. For each of the 30 yr, maps of ice breakup date were created by kriging (as above). The resulting 5-km resolution raster maps were averaged by 5-yr intervals (1975–1979, 1980–1984, etc.), and isophenes were created (ArcView v8.3, Spatial Analyst Extension). The difference in the average Y (north–south) coordinate between the 1975–1979 and 2000–2004 breakup isophenes was calculated for each column in the raster maps to estimate the average movement of the breakup date isophenes over the study period. Freeze isophenes and maps of freeze date and ice duration trends were not created because there were too few records (33 and 32, respectively) to permit accurate mapping across the entire study region.

Trends in ice phenology and duration were compared to waterbody characteristics (elevation, surface area, and maximum and average depth) and meteorological variables using single- and multiple-variable regression. While mixing depth should be a better measure than mean or maximum depth for this analysis, mixing depths are not available for all of these lakes, and many do mix completely in the fall. Potential confounding of waterbody characteristics and weather station elevation with geographic location (e.g., waterbody elevation is not randomly distributed across the study area) was addressed by including geographic coordinates in the multiple regression whenever significant relationships between location and these characteristics were found. In such cases, we report the partial R^2 for the waterbody characteristic, controlling for the effect of geographic location.

Because some of the scatter plots indicated potential nonlinear relationships, we also applied generalized additive models (GAMs; Hastie and Tibshirani 1990) based on regression splines (mgcv package v1.3 for R v2.4; Wood and Augustin 2002). GAMs are capable of incorporating both linear and nonlinear response curves from multiple individual predictor variables. Degrees of freedom (i.e., flexibility) for individual response curves were modified manually based on adjusted R^2. That is, the flexibility of

Table 1. Locations and ice phenology and duration trends for all waterbodies. Asterisks indicate statistically significant ($p < 0.05$) trends.

		Freeze (d decade^{-1})		Breakup (d decade^{-1})		Duration (d decade^{-1})				
State/province	Lake/river/bay name	Slope	SE	Slope	SE	Slope	SE	Latitude (°N)	Longitude (°W)	Elevation (m)
Michigan	Duck	–	–	−7.07*	2.98	–	–	42.39	84.74	283
	Fair	2.18	2.14	−1.41	3.16	−3.59	4.39	42.49	85.33	280
	Gull†	4.42	3.09	−4.71	3.16	−9.41	4.86	42.4	85.41	268
	Houghton	–	–	−4.67	2.39	–	–	44.35	84.73	347
Minnesota	Bemidji	–	–	−1.12	1.52	–	–	47.5	94.83	408
	Big Stone	–	–	−2.16	2.01	–	–	45.5	96.5	–
	Clear	–	–	−2.19	2.74	–	–	44.07	93.5	311
	Detroit	–	–	−2.20	1.91	–	–	46.78	95.93	407
	Galpin	–	–	−3.58	2.58	–	–	44.9	93.56	287
	Green	2.87	1.87	−2.42	2.05	−4.92	3.14	45.25	94.9	352
	Kabetogama	–	–	−0.70	1.65	–	–	48.53	93.08	341
	Leech	–	–	−1.50	1.65	–	–	47.12	94.12	395
	McKinney	–	–	−2.72	1.99	–	–	47.25	93.53	392
	Mille Lacs	–	–	−3.12	1.88	–	–	46.42	93.37	381
	Minnetonka	–	–	−2.53	2.03	–	–	44.87	93.57	283
	Minnewaska	–	–	−2.32	1.99	–	–	45.6	95.47	347
	Osakis	–	–	−4.08*	1.85	–	–	45.87	95.13	403
	Rainy	–	–	−0.57	1.62	–	–	48.6	93.36	338
	Vermillion	–	–	−0.42	1.53	–	–	47.17	93.87	390
	White Bear	–	–	−4.37*	2.01	–	–	45.07	92.99	282
New York	Bonaparte	3.48	2.09	0.60	2.05	−1.75	3.46	44.16	75.4	240
	Brant	1.67	2.13	−1.37	1.59	−3.04	2.93	43.68	73.74	243
	Cassadaga (Lower)	2.62	2.50	−0.46	2.43	−3.09	3.61	42.34	79.32	398
	Cazenovia	3.20	2.08	−0.70	2.43	−3.90	3.94	42.93	75.86	363
	Chateaguay (Lower)	4.29	2.57	−1.03	2.06	−4.95	3.76	44.84	74.04	399
	Chautauqua South	2.43	2.08	−2.87	2.80	−5.29	4.10	42.11	79.1	399
	Cranberry	4.44*	1.72	−0.97	1.92	−5.40	2.76	44.22	74.83	457
	Deep	3.80	2.83	−0.65	2.07	−4.70	3.78	43.03	77.57	–
	Genegantslet	0.81	2.07	−0.69	1.94	−1.50	3.22	42.51	75.77	454
	George‡	5.38	3.22	−1.48	2.40	−6.85	4.85	43.83	73.43	97
	Glen	0.70	2.41	0.49	2.27	−0.21	3.95	41.9	75.02	398
	Loon	−1.34	2.55	−2.42	2.37	−1.08	3.99	42.48	77.56	518
	Mirror	1.36	1.76	−0.95	1.87	−2.32	2.77	44.29	73.99	565
	Mohansic	4.00	2.57	0.76	2.64	−4.09	4.01	41.28	73.81	448
	Mohonk	3.42	2.14	−0.26	2.21	−3.68	3.48	41.76	74.16	380
	Oneida	6.93*	2.64	−2.59	3.45	−10.06	5.31	43.24	76.14	112
	Otsego§	4.27	3.18	−2.20	2.75	−6.47	4.83	42.69	74.93	363
	Placid	1.13	1.99	−0.74	1.65	−1.87	2.77	44.3	73.99	566
	Saranac (Lower)	1.36	1.69	−0.38	1.94	−1.74	2.78	44.29	74.19	507
	Schroon	5.71*	2.21	−1.42	1.48	−7.13*	2.77	43.73	73.81	246
	St. Regis (Lower)	–	–	−0.03	1.89	–	–	44.43	74.29	493
	Star	1.86	2.12	−0.77	1.91	−2.63	3.05	44.15	75.04	442
	Sylvia	2.74	3.26	−0.32	2.30	−2.95	5.05	44.26	75.41	199
	Titus	–	–	0.00	1.98	–	–	44.74	74.29	426
Ontario	Ashby	–	–	−1.31	2.24	–	–	45.13	77.35	–
	Bass	–	–	−2.24	1.95	–	–	45.13	79.69	–
	Crowe River	–	–	−0.84	1.83	–	–	44.84	77.93	–
	Dorset Lakes (Mean)	–	–	−1.31	2.24	–	–	45	78	–
	Opeongo	–	–	−0.93	1.60	–	–	45.7	78.37	403
	Rice Lake at Indian River	–	–	−0.01	3.45	–	–	44.23	78.15	–
	Simcoe	–	–	−4.17	2.74	–	–	44.38	79.68	256
Wisconsin	Anderson	–	–	−1.26	1.78	–	–	46.17	89.35	520
	Big Green‖	2.28	2.99	−8.42*	3.99	−10.70	5.95	43.8	89	243
	Black Oak	–	–	−1.05	1.90	–	–	46.16	89.31	–
	Chequamegon Bay	–	–	−3.34	2.16	–	–	46.67	90.88	183
	Devils	6.98*	2.72	−3.95	2.42	−10.93*	3.60	43.42	89.73	294
	Geneva¶	7.81*	3.14	−10.54*	3.70	−18.35*	5.81	42.58	88.51	263

Table 1. Continued.

State/province	Lake/river/bay name	Freeze (d decade^{-1}) Slope	SE	Breakup (d decade^{-1}) Slope	SE	Duration (d decade^{-1}) Slope	SE	Latitude (°N)	Longitude (°W)	Elevation (m)
	Mallalieu	–	–	−1.21	1.82	–	–	44.98	92.77	211
	Maple	2.27	2.21	−1.55	1.68	−3.82	3.14	46.13	89.73	498
	Mendota	4.65*	2.26	−4.70	2.48	−9.34*	3.92	43.1	89.4	259
	Monona	3.40	2.46	−4.52	2.36	−7.92*	3.83	43.05	89.37	258
	North Twin	–	–	−3.22	2.36	–	–	46.05	89.13	513
	Shell	2.77	1.88	−0.81	1.66	−3.43	2.78	45.73	91.9	371
	Shishebogama	5.83*	2.03	–	–	–	–	45.9	89.82	480
	Wingra	2.47	2.54	−5.31*	2.58	−8.74*	3.96	43.05	89.42	258

† Record contained one no-freeze year (2002). Dropping this year gives a freeze date trend of 2.65 ± 3.00 d decade^{-1} and a breakup date trend of −2.84 ± 3.10. Substituting the latest observed freeze date and earliest observed breakup date gives a freeze date trend of 3.93 ± 2.97 d decade^{-1} and a breakup date trend of −4.82 ± 3.19.

‡ Record contained one no-freeze year (2002). Dropping this year gives a freeze date trend of 3.26 ± 3.03 d decade^{-1} and a breakup date trend of 0.69 ± 2.00. Substituting the latest observed freeze date and earliest observed breakup date gives a freeze date trend of 4.47 ± 2.99 d decade^{-1} and a breakup date trend of −0.51 ± 2.06.

§ Record contained one no-freeze year (2002). Dropping this year gives a freeze date trend of 1.88 ± 2.89 d decade^{-1} and a breakup date trend of −0.13 ± 2.50. Substituting the latest observed freeze date and earliest observed breakup date gives a freeze date trend of 3.56 ± 2.96 d decade^{-1} and a breakup date trend of −2.24 ± 2.77.

|| Record contained one no-freeze year (2002). Dropping this year gives a freeze date trend of 0.54 ± 2.90 d decade^{-1} and a breakup date trend of −6.2 ± 3.90. Substituting the latest observed freeze date and earliest observed breakup date gives a freeze date trend of 1.93 ± 2.90 d decade^{-1} and a breakup date trend of −9.55 ± 4.33.

¶ Record contained two no-freeze years (1998 and 2002). Dropping these years gives a freeze date trend of 5.00 ± 3.00 d decade^{-1} and a breakup date trend of −6.76 ± 3.34. Substituting the latest observed freeze date and earliest observed breakup date gives a freeze date trend of 7.10 ± 2.97 d decade^{-1} and a breakup date trend of −10.69 ± 3.74.

the spline for each significant predictor variable was changed until the adjusted R^2 of the model was maximized. A term with a single degree of freedom corresponds to a straight-line relationship between the response (ice phenology or duration trends) and a predictor (waterbody characteristics or meteorological variables). Higher degrees of freedom allow the relationship between the response and the predictor variable to take a more flexible form including, for example, dome-shaped, threshold, or multimodal curves.

Results

Ice phenology—Over the 30-yr study period, the breakup date trended earlier at an average rate of 2.1 d decade^{-1} (standard error [SE] ± 0.3, $p < 0.001$), the freeze date trended later at an average rate of 3.3 d decade^{-1} (SE ± 0.4, $p < 0.001$), and ice duration became shorter at an average rate of 5.3 d decade^{-1} (SE ± 0.7, $p < 0.001$). While these patterns were highly significant for the waterbodies as a group, most individual waterbodies did not show statistically significant trends (Table 1) owing to the high interannual variability and relatively short time series. Alternative methods of handling no-ice years resulted in substantially different trends for the five individual lakes that did not freeze in one or more winters (*see footnotes to Table 1*) but altered the average trends for all waterbodies combined by no more than 10%. All subsequent analyses of trends were conducted with freeze and breakup dates for no-ice years inferred using the method of Assel and Robertson (1995).

Trends in breakup date caused breakup isophenes to move northward (Fig. 1A). The 15 April isophene, which connects waterbodies that break up on 15 April, moved northward by an average of 95 (standard deviation [SD], ± 29) km between the first (1975–1979) and last (2000–2004) pentads.

Three waterbody characteristics (elevation, mean depth, and surface area) were significant predictors of trends in ice phenology and/or duration, even after controlling for a spatial trend toward larger, higher-elevation lakes in the northern part of the study area (Fig. 2; Table 2). Higher-elevation lakes had weaker trends in freeze, breakup, and ice duration; elevation explained 29% of the variation in freeze date trends ($n = 32$, $p = 0.002$), 7% of the variation in breakup date trends ($n = 55$, $p = 0.03$), and 23% of the variation in ice duration trends ($n = 31$, $p = 0.002$). Greater mean depth was weakly associated with stronger trends in breakup date ($R^2 = 0.15$, $n = 27$, $p = 0.049$). Greater surface area was associated with stronger trends in freeze date ($R^2 = 0.24$, $n = 29$, $p = 0.008$). No other comparisons of trend versus waterbody characteristics were statistically significant at $\alpha = 0.05$.

There were pronounced geographic patterns in the breakup date and ice duration trends. Trends toward earlier breakup ($n = 64$) and shorter ice duration ($n = 33$) were strongest in the southwest part of the study area. The north–south, east–west, and interaction terms of the spatial patterns were all highly significant ($p < 0.01$). The geographic pattern explained 50% of the variation in breakup date trends and 51% of the variation in ice duration trends. Breakup date trends for individual lakes

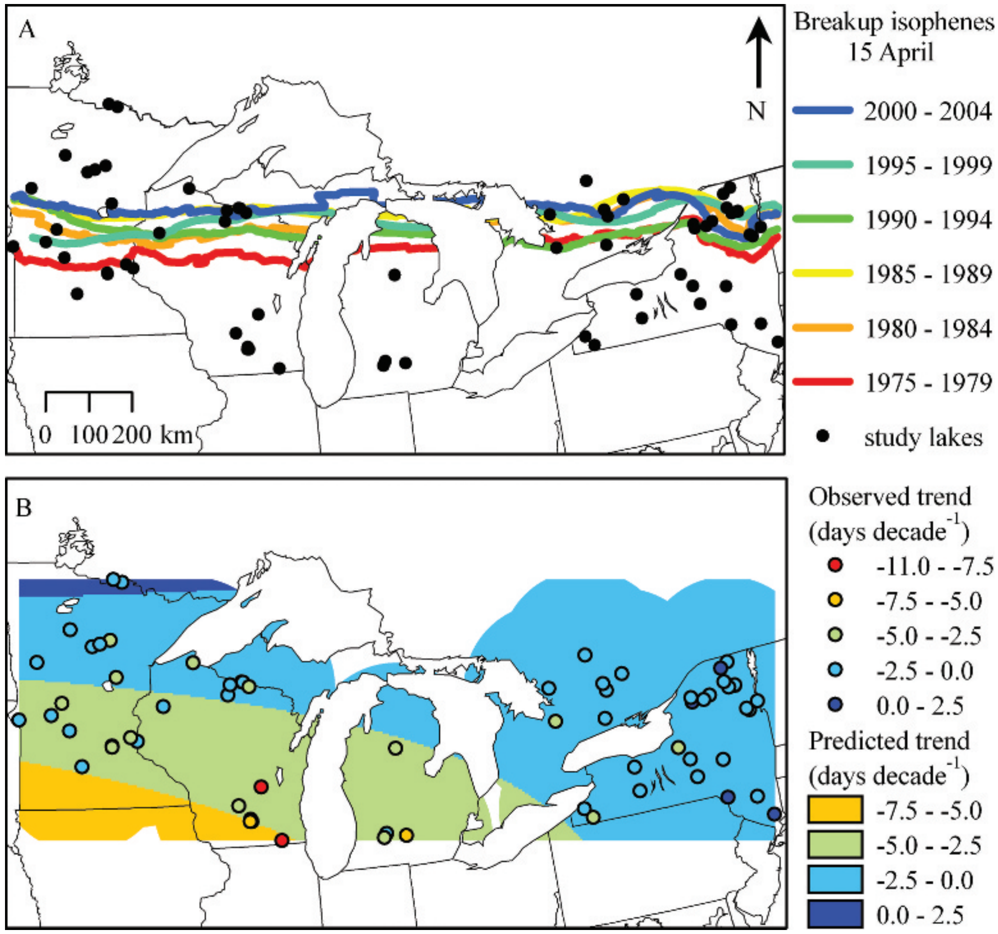


Fig. 1. (A) The 15 April ice breakup isophene averaged by 5-yr interval and (B) regression surface predictions (trend = X + Y + (X × Y); where X and Y are spatial coordinates) and observed temporal trends in ice breakup date across the Laurentian Great Lakes region. Areas in panel B falling beyond 200 km from the nearest waterbody are not shown.

and the spatial regression predictions for the entire study area are shown in Fig. 1B. No significant spatial pattern in freeze date trend was found ($n = 34$, $p > 0.05$ for all terms).

Meteorological variables—Between 1975 and 2004, the mean fall-to-spring temperature increased at an average rate of 0.69°C decade^{-1} (SE ± 0.03, $p < 0.001$), the average daily minimum temperature increased at a rate of 0.79°C decade^{-1} (SE ± 0.03, $p < 0.001$), and the average daily maximum temperature increased at a rate of 0.56°C decade^{-1} (SE ± 0.02, $p < 0.001$). The number of days with snow on the ground decreased at an average rate of 5.0 d decade^{-1} (SE ± 0.49, $p < 0.001$), and the average snow depth on days with snow on the ground decreased at an average rate of 1.7 cm decade^{-1} (SE ± 0.20, $p < 0.001$). As with ice phenology, these meteorological trends were highly significant for the weather stations as a group, but many individual stations did not show statistically significant trends. For air temperature, 53 out of 144 stations had significant ($p < 0.05$, with no correction for multiple comparisons) positive trends in average air temperature, 94 out of 235 stations had significant positive trends in average daily minimum air temperature, and 72 out of 238 stations had significant positive trends in average daily maximum air temperature. Fewer than 4% of stations had negative trends in any of the air temperature values, and none of these trends were statistically significant. Similarly, for snow days, 98 out of 105 stations had negative trends, 13 of which were statistically significant. For mean snow depth, 89 out of 105 stations had negative trends, and 11 of these trends were statistically significant. None of the positive trends in snow variables were statistically significant.

Even when controlled for spatial trends in elevation, higher-elevation weather stations tend to be colder ($p < 0.001$ for all temperature variables, partial $R^2 = 0.04$ for average and maximum temperatures and 0.06 for minimum temperature) and have more days with snow ($p < 0.001$, partial $R^2 = 0.10$) and a higher average snow depth on those days ($p = 0.001$, partial $R^2 = 0.06$). Of the temporal trends in meteorological variables, only the trend in days with snow ($p = 0.006$, partial $R^2 = 0.05$) was significantly related to elevation, with higher-elevation stations tending

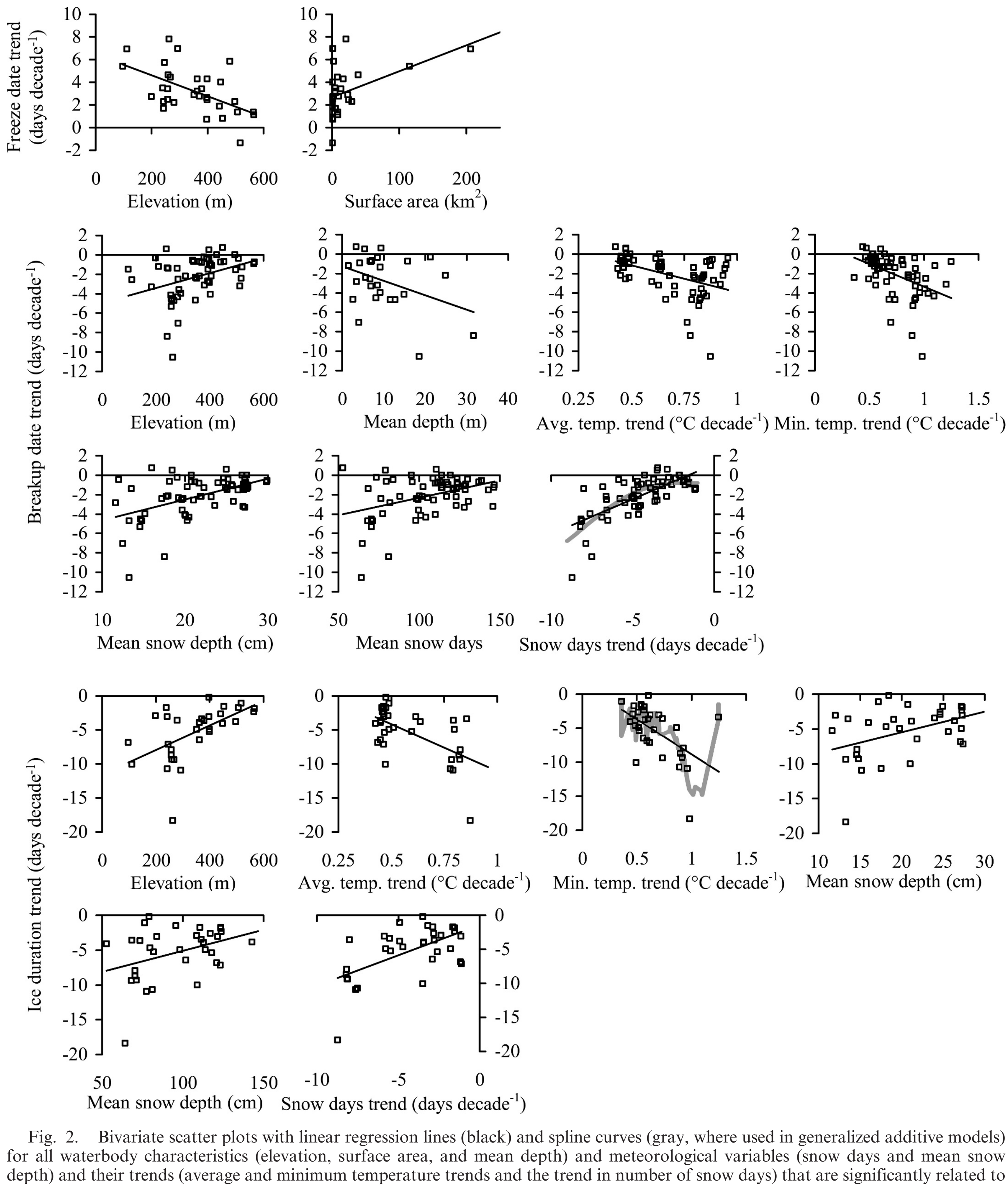


Fig. 2. Bivariate scatter plots with linear regression lines (black) and spline curves (gray, where used in generalized additive models) for all waterbody characteristics (elevation, surface area, and mean depth) and meteorological variables (snow days and mean snow depth) and their trends (average and minimum temperature trends and the trend in number of snow days) that are significantly related to ice phenology (freeze and breakup) and duration trends.

Table 2. Values of R^2 for single- and multiple-variable regressions of ice phenology (freeze and breakup dates) and duration trends against waterbody characteristics (elevation, mean depth, and surface area); fall-to-spring means of daily average, minimum, and maximum temperatures; temporal trends in average and minimum temperatures; mean snow depth and number of days with snow cover; and the temporal trend in the number of days with snow cover. Signs for single-variable regression models represent the direction of the relationship (e.g., a positive sign indicates that waterbodies with higher [more positive] values of that variable have stronger trends [positive for freeze date and ice duration; negative for breakup date] in ice phenology or duration). Values of R^2 for single regression models on waterbody characteristics are partial R^2, controlled for geographic trends in elevation, mean depth, or surface area. Values of R^2 for multiple regression models represent partial R^2 and are not additive. Model R^2 is given under "Total."*

		Lake characteristics			Temperature			Temp. trends		Snow		Snow trends		
Trend	Model	Elev.	Depth	Area	Avg.	Min.	Max.	Avg.	Min.	Depth	Days	Days	Interactions	Total
Freeze	Single reg.	−0.29	–	+0.24	–	–	–	–	–	–	–	–	–	
	Mult. reg.	0.29	–	–	–	–	–	–	–	–	–	–	–	0.29
Breakup	Single reg.	−0.07	+0.15	–	–	–	–	+0.21	+0.21	−0.26	−0.16	−0.50	–	
	Mult. reg.	–	–	–	–	–	–	–	–	–	–	0.50	–	0.50
	Mult. reg.	–	–	–	0.13	–	–	0.48	–	–	–	–	0.25	0.56
	Mult. reg.	–	–	–	–	0.14	–	0.40	–	–	–	–	0.26	0.55
	Mult. reg.	–	–	–	–	–	0.13	0.50	–	–	–	–	0.25	0.55
Duration	Single reg.	−0.23	–	–	–	–	–	+0.35	+0.29	−0.18	−0.14	−0.35	–	
	Mult. reg.	0.29	–	–	–	–	–	–	–	–	–	0.33	–	0.54
	Mult. reg.	0.26	–	–	–	–	–	0.31	–	–	–	–	–	0.53
	Mult. reg.	0.26	–	–	–	–	–	–	0.24	–	–	–	–	0.48

* Temp., temperature; elev., elevation; avg., average; min., minimum; max., maximum; mult., multiple; reg., regression.

to have less-rapid declines, or even small increases, in days with snow.

The predominant geographic trends in the means of all three temperature variables were in the north–south direction, although some coastal or urban influence is apparent in southern New York (Fig. 3A,B,D). Spatial trends (north–south, east–west, and their interaction) accounted for 82–91% of the variability among weather stations in the means of the temperature variables. Of the three temperature variables, two, average air temperature ($R^2 = 0.25$, $p < 0.001$) and average daily minimum air temperature ($R^2 = 0.18$, $p < 0.001$), had significant geographic patterns in their temporal trends. Both air temperature variables increased more rapidly in the western part of the study area (Fig. 3C,E).

North–south trends in snow depth (Fig. 4A) and snow days are apparent (Fig. 4B), and there were statistically significant north–south, east–west, and interaction terms for both variables. These geographic patterns explained 50% of the variation in snow depth and 75% of the variation in snow days. The geographic pattern in the trend in snow days (Fig. 4C) is similar to that of the trend in ice breakup date, with the most rapid changes occurring in the southwest part of the study area. There were significant north–south, east–west, and interaction terms in the spatial regression model that explained 33% of the variation in snow day trends.

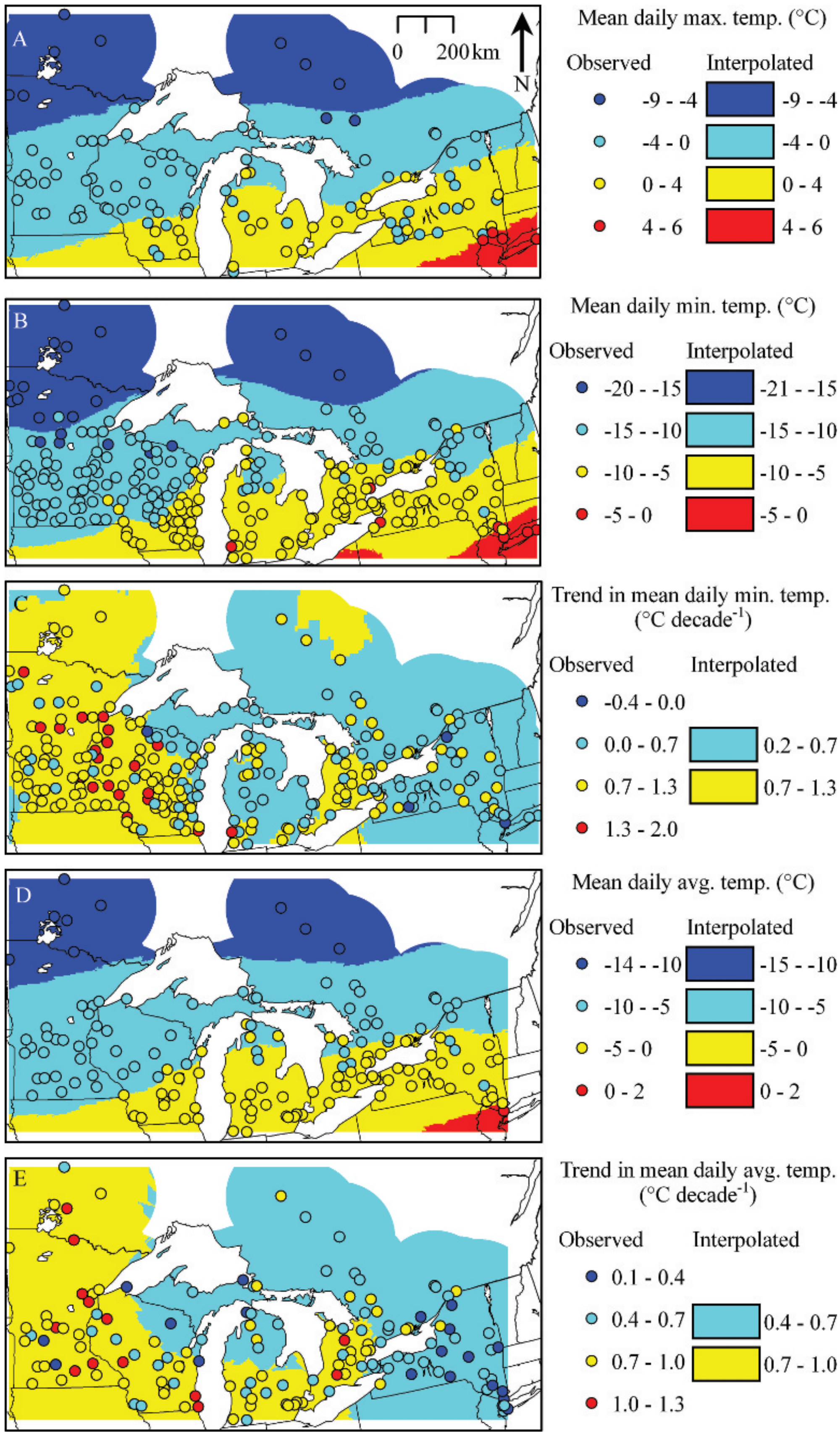
Modeling ice phenology and duration trends using waterbody characteristics and meteorological variables—Significant linear relationships were found between the temporal trends in ice breakup date and ice duration and the meteorological variables (Fig. 2; Table 2). The strongest relationship was between the trend in ice breakup date and the trend in snow days ($R^2 = 0.50$, $p < 0.001$). In addition, mean snow depth, snow days, the trend in average temperature, and the trend in minimum daily temperature were significantly related to trends in breakup date, but each of these meteorological variables individually explained only 16–26% of the variability of the trend in breakup date. The three remaining meteorological variables (average, maximum, and minimum daily temperatures) were not significantly correlated with trends in ice breakup date. The trend in ice duration was linearly related to the trend in average temperature ($R^2 = 0.35$, $p < 0.001$), the trend in snow days ($R^2 = 0.35$, $p < 0.001$), the trend in minimum temperature ($R^2 = 0.29$, $p = 0.001$), the mean snow depth ($R^2 = 0.18$, $p = 0.014$), and the average number of snow days ($R^2 = 0.14$, $p = 0.030$). Trends in freeze date were not significantly correlated with any of the meteorological variables.

Trends in ice breakup could be predicted most parsimoniously by a simple one-variable linear regression against trends in snow days ($R^2 = 0.50$, $p < 0.001$; model: trend in breakup date = 1.1 + 0.7 × trend in number of snow days; trends are in d decade^{-1}). Waterbodies that showed more rapid decreases in breakup date were located in areas with more rapid declines in snow days. Because both of these trends are likely driven by temperature variables, we also considered multivariable models using temperature and trends in temperature. Owing to strong correlations among the temperature variables, several similar two-variable models were nearly equal at predicting ice breakup trends. For example, models based on trend in average temperature plus any one of the following temperature variables and the interaction term: average temperature ($R^2 = 0.56$), minimum temperature ($R^2 = 0.55$), or maximum temperature ($R^2 = 0.55$), all performed almost equally well. The model with average temperature is explained as follows: the trend in breakup date = 8.8 + 1.0 × average temperature − 19.7 × trend in average temperature − 2.0 × (average temperature × trend in average temperature) (trends are in d decade^{-1} and temperature is in °C). Waterbodies with greater increases in air temperature or higher average, minimum, or maximum temperatures experienced faster declines in ice breakup date. Although significantly correlated with trends in ice breakup date, elevation did not appear in the top 10 two-variable models. Models containing three or more variables offered no improvement over the two-variable models.

Trends in freeze date were best predicted by elevation ($R^2 = 0.29$, $p = 0.001$; model: the trend in freeze date = 6.4 − 0.01 × elevation (m); trend is in d decade^{-1}). Higher-elevation lakes tended to have weaker (less-positive) trends in freeze date. No other model in which all terms were significant had a higher R^2.

Trends in ice duration were explained best by a combination of elevation and the trend in snow days ($R^2 = 0.54$, $p < 0.001$; model: the trend in ice duration = −7.0 + 0.01 × elevation (m) + 0.7 × trend in number of snow days; trends are in d decade^{-1}) or the trend in average air temperature ($R^2 = 0.53$). Three-variable models offered minimal ($R^2 = 0.55$ for the best model) increases in R^2 relative to the best two-variable models.

We examined potential nonlinearities in the response of ice phenology and duration trends to meteorological and lake characteristic variables using scatter plots (Fig. 2) and GAMs based on regression splines. In some cases, GAMs yielded slightly higher explanatory power, even when R^2 was adjusted for model degrees of freedom. A GAM of trend in breakup date versus trend in snow days had an adjusted R^2 of 0.57 ($p < 0.001$, model df = 2). This model indicated that the trend in snow days has little relationship with the trend in breakup date where the loss of snow days was less than 4 d decade^{-1}; however, at locations in which the loss of snow days was more rapid, there was a strong relationship between these two trends (*see the spline in Fig. 2*). Using temperature variables as predictors of trend in breakup date, GAMs offered no improvement over the multiple regression model (average temperature and trend in average temperature), which had an adjusted R^2 of 0.54. For ice duration, the best two-variable GAM in terms of adjusted R^2 (0.66) included a linear term for elevation ($p = 0.004$) and a highly flexible nonlinear term for trend in minimum temperature ($p = 0.002$, model df = 5.9; *see the spline in Fig. 2*). The best multiple regression model for ice duration had an adjusted R^2 of 0.51. GAMs provided no improvement over regression models for the prediction of trends in freeze date.



A
0
200 km
N
Mean daily max. temp. (°C)
Observed
Interpolated
-9 - -4
-4 - 0
0 - 4
4 - 6
B
Mean daily min. temp. (°C)
Observed
Interpolated
-20 - -15
-21 - -15
-15 - -10
-10 - -5
-5 - 0
C
Trend in mean daily min. temp. (°C decade-1)
Observed
Interpolated
-0.4 - 0.0
0.0 - 0.7
0.2 - 0.7
0.7 - 1.3
1.3 - 2.0
D
Mean daily avg. temp. (°C)
Observed
Interpolated
-14 - -10
-15 - -10
-10 - -5
-5 - 0
0 - 2
E
Trend in mean daily avg. temp. (°C decade-1)
Observed
Interpolated
0.1 - 0.4
0.4 - 0.7
0.7 - 1.0
1.0 - 1.3

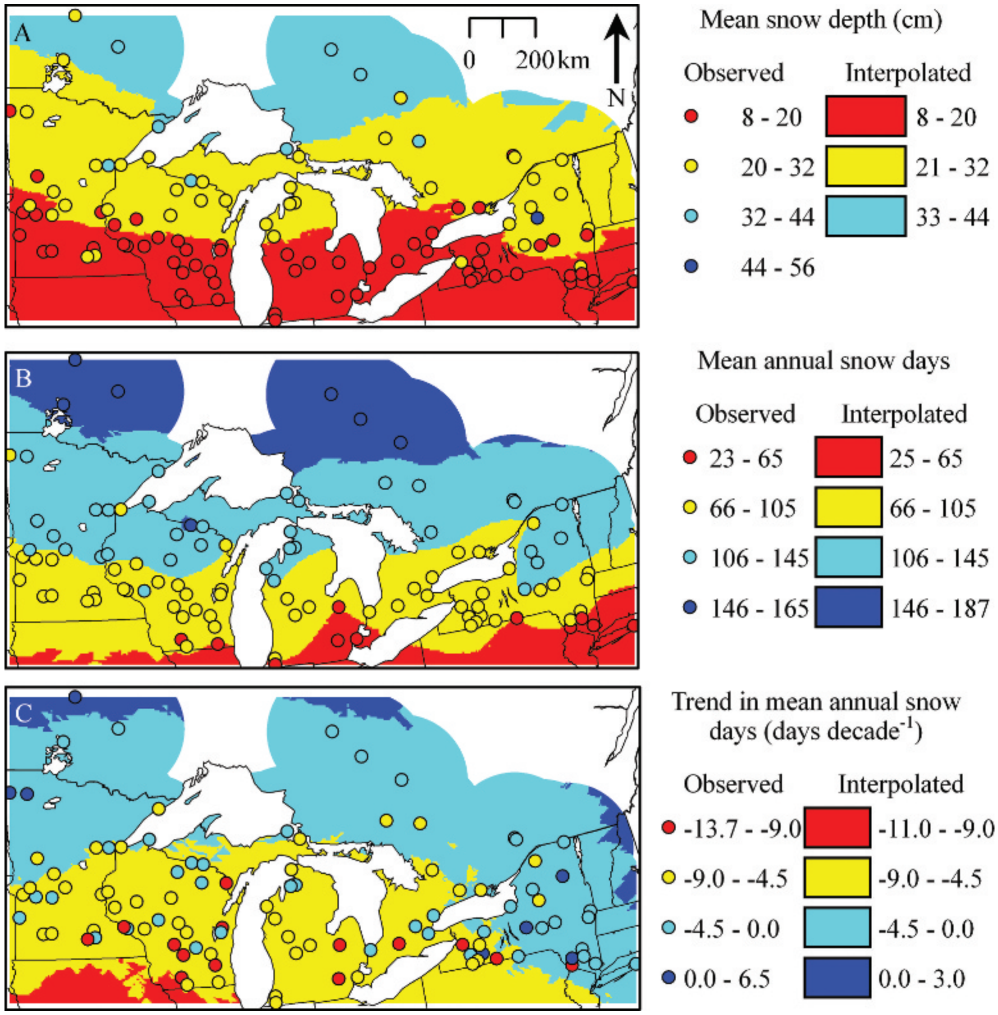


Fig. 4. Observed and geostatistical interpolations of (A) mean snow depth on days with snow on the ground (cm), (B) mean number of days with snow on the ground, and (C) trend in mean number of days with snow on the ground (d decade^{-1}). Areas falling beyond 200 km from the nearest weather station are not shown.

Discussion

Trends in ice phenology observed in the study waterbodies during this recent warming period are substantially more rapid than the historical trends reported by Magnuson et al. (2000) for the period ranging from 1846 to 1995. Over the study period, 1975–2004, a 2.1°C (0.7°C decade^{-1}) increase in average fall-to-spring temperature was accompanied by a 10-d (3.3 d decade^{-1}) increase in the average freeze date, a 6-d (2.1 d decade^{-1}) decrease in the average breakup date, and a 16-d (5.3 d decade^{-1}) decrease in ice duration. Average snow depth for the region decreased by 5 cm (1.7 cm decade^{-1}), and the average number of days with snow decreased by 15 d (5.0 d decade^{-1}). Rates of change were 5.8 (freeze date) and 3.3 (breakup date) times more rapid than the rates reported by Magnuson et al. (2000) for Northern Hemisphere waterbodies (0.57 d decade^{-1} later freeze and 0.63 d decade^{-1} earlier breakup, substituting an updated value of 0.0 d decade^{-1} for Detroit Lake).

←

Fig. 3. Observed and geostatistical interpolations of temperatures for the period 01 September–31 May, including (A) mean daily maximum temperature (°C), (B) mean daily minimum temperature (°C), (C) trend in mean daily minimum temperature (°C decade^{-1}), (D) mean daily average temperature (°C), and (E) trend in mean daily average temperature (°C decade^{-1}). Areas falling beyond 200 km from the nearest weather station are not shown.

The trend toward earlier ice breakup resulted in a northward movement of breakup isophenes. The average rate of movement of the 15 April breakup isophene was 3.8 km yr^{-1}. In comparison, isotherms in the Northern Hemisphere are estimated to have moved northward at a similar average rate of 4 km yr^{-1} from 1975 to 2000 (Hansen et al. 2006).

The changes in ice phenology and duration we observed are comparable to changes observed in other locations during the later part of the 20th century and are substantially more rapid than trends exhibited over longer time periods. Schindler et al. (1990) reported a decline in ice duration of 20 d over the 20-yr period ranging from 1969 to 1988 (equivalent to 10 d decade^{-1}) for lake 239 in the Experimental Lakes Region of northwestern Ontario. In the northeastern U.S.A., trends in lake ice breakup dates for the period ranging from 1850 to 2000 were 0.6 and 1.0 d decade^{-1} earlier in the northern/mountainous and southern regions of New England, respectively (Hodgkins et al. 2002). From 1840 to 1994, the breakup date of Lej da San Murezzan in the Swiss Alps became earlier at a rate of 0.76 d decade^{-1} (Livingstone 1997). From 1885 to 2002, the average rate of change in freeze dates for Finnish lakes ranged from 0.36 to 0.79 d earlier decade^{-1} in different regions of the country, while breakup dates moved later at average rates of 0.66 to 0.86 d decade^{-1} (Korhonen 2006). Johnson and Stefan (2006) found that ice breakup on 73 lakes in Minnesota came earlier, at an average rate of 1.3 d decade^{-1}, over the period from 1965 to 2002, while freeze on 34 Minnesota lakes has come later, at a rate of 7.5 d decade^{-1} from 1979 to 2002. The rates of change in ice phenology on these lakes were approximately twice as rapid during the period extending from 1990 through 2002.

Climate warming in the Northern Hemisphere during the 30-yr study period occurred more rapidly than in the past (IPCC 2001), and steeper trends in many other climate-related phenomena have been reported. For example, the decrease in the total number of days with ice-affected flow in northern New England rivers and a trend toward earlier last dates of ice-affected flow were more marked during the 1960s to 2000 than they were earlier (Hodgkins et al. 2005). A variety of other phenological variables, including the blooming of lilacs and honeysuckle and spring snowmelt, have changed more rapidly in the past 30 yr (Cayan et al. 2001). In Minnesota, peak spring runoff came earlier at a rate of 2.3 d decade^{-1} from 1964 to 2002 (Johnson and Stefan 2006), and average stream-water temperatures increased by 1.1°C decade^{-1}.

Spatial patterns of trends in annual number of snow days were well correlated with patterns of trends in breakup date. The relationship between trends in breakup date and mean snow depth was weaker, but snow depth was still a better predictor of breakup date trends than were any of the temperature variables. There are at least two possible (and not mutually exclusive) explanations for the stronger predictive power of snow variables compared to temperature variables. The first is that snow has a direct effect on ice breakup trends through albedo effects (white or snow-covered ice absorbs less heat) and/or insulation (heat transfer between air and ice is reduced when the ice is covered by snow). Both of these direct mechanisms are represented in the ice-cover model developed by Vavrus et al. (1996). In that study, halving the snowfall resulted in the ice breakup date occurring 4 d earlier, while doubling the snowfall delayed the breakup date by 12 d. Comparisons of our results with the predictions of that model are hampered by the fact that the model used snowfall rather than snow depth. Snowfall records for the study period were not available at most of the weather stations used for this study.

The direct effect of snow on ice cover may be more complicated than could be captured by the two snow variables we used. For example, the insulating effect of snow cover during the early part of the winter may reduce heat flux from the ice to the air, thus impeding ice growth. In the spring, snow cover may impede melting, because snow reflects solar radiation and reduces heat flux from the air to the ice.

A second interpretation of the relationship between snow and ice phenology trends is that snow day trends are simply driven by the same temperature variables that drive ice cover. Snow day trends reflect temperature and temperature trends during the same time of year that ice cover is present. This critical time of year varies over the study region. Snow day trends reflect this variability, but temperature summarized over a single time period does not. We chose not to summarize temperature for different seasonal periods at different locations because this would have confounded the spatial comparisons. Like ice phenology, snow depth shows long-term effects of warming, with stronger trends since the mid-1970s (Brown and Braaten 1998).

Within the broad trends in ice phenology and meteorological variables, substantial variability exists among waterbodies and weather stations. Much of this variability showed clear spatial patterns. The trend toward earlier ice breakup was strongest for waterbodies in the southwestern part of the study area. The maps of change in average and average minimum temperature show stronger trends in the western part of the study area. This east–west pattern in the temperature trends could result from changes in the predominant large-scale climate oscillation affecting different parts of the region or from potential differences in coastal versus continental climate trends. The timing of ice breakup in southern Wisconsin lakes is associated with changes in the El Nino Southern Oscillation (Anderson et al. 1996). In the temperate and south boreal regions of central and eastern North America ice breakup is associated with the winter PNA, the Pacific/North American Pattern (Benson et al. 2000*a*). Ice phenology in the study area may also be influenced by Atlantic climate oscillations, such as the North Atlantic Oscillation (NAO). The NAO has been associated with effects on Western European lakes (George et al. 2004) and on Lake Baikal (Livingstone 1999). The NAO signal in the ice phenology records for Lake Mendota (one of the lakes included in this analysis) was strong in the latter half of the 19th century through the first half of the 20th century, but has since weakened (Livingstone 2000). The positive air temperature trends across the study area are part of a global pattern of long-term rising temperature (IPCC 2001) combined with periodic climate oscillations, such as the warm phase of the Pacific Decadal Oscillation that began in the mid-1970s (Benson et al. 2000*a*; Schindler et al. 2005). Not surprisingly, waterbodies in locations with stronger temperature trends also showed stronger trends in ice breakup date and ice duration.

In contrast to the general observation that climate changes are occurring more rapidly at higher latitudes (IPCC 2001, p. 13), the greatest rate of change in ice breakup dates in this region is occurring at the lower latitudes (Fig. 1B), near the southern boundary of the area in which lakes are routinely ice covered during winter. This greater response is consistent with the nonlinear relationship observed between ice breakup date and air temperature in Swedish lakes (Weyhenmeyer et al. 2004) and indicates that the disappearance of ice in southern lakes may occur rapidly with only small changes in average air temperature. This nonlinear relationship may result from the simple fact that a small increase in temperature is more likely to switch conditions from freezing to melting when it occurs near 0°C. In addition, thicker ice in colder regions is more able to withstand brief melting periods. Other factors may play a role as well: for example, the influence of direct radiation would be greater in the south owing to the greater height of the sun and the longer day length.

The dampening effect of elevation on trends in freeze date and ice duration may be related in part to the generally colder temperatures experienced by higher-elevation lakes as a result of adiabatic cooling. This effect of elevation on temperature was evident from comparisons among weather stations at different elevations. The absence of a relationship between any of the temperature or snow variables and freeze date trends casts doubt on this simple explanation of the elevation effect. Faster temperature increases are predicted for higher elevations by climate models (Giorgi et al. 1997). However, within the relatively narrow range of elevations in our study region (waterbodies ranged from 97 to 566 m, weather stations from 3.4 to 807 m), we found no significant effect of elevation on temperature trends once the relationship was corrected for geographic location. The ice-cover model developed by Thompson et al. (2005) predicts greater changes in ice duration at higher elevation, up to a maximum at ~1,500–2,000 m. That model, however, predicts absolute change, not rate of change, and because warmer, lower-elevation lakes start out with shorter ice duration, the maximum possible change, i.e., from current duration to zero, is lower.

During the late 1980s and early 1990s, ice breakup occasionally occurred in midwinter for two southern lakes in Michigan (Fair and Gull Lakes) that previously had broken up in the spring. By the end of the study period, in the years 1998 and 2002, several lakes did not freeze over. These lakes were in the southern portion of the study area (approximately the southernmost one third of the waterbodies) and tended to be larger, deeper, and lower elevation than average lakes. This result is not surprising, because deeper lakes require a more prolonged period of below-freezing air temperatures before they freeze.

Taken together, our results indicate that (1) in general, recent changes in ice phenology and duration have been much more rapid than the long-term average trends; (2) waterbodies in the southwestern region of the study area are experiencing the most rapid decreases in ice duration and breakup date; (3) large, deep, low-elevation waterbodies are losing ice cover more rapidly; and (4) spatial patterns in breakup date trends are correlated with spatial patterns in temperature trends, snow depth, and snow day trends.

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