ANALYSIS OF THE WESTERN SHORE CHESAPEAKE BAY BAY-BREEZE

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Abstract

Hourly surface data observations were employed to identify days on which bay-breezes occurred along the western shore of Chesapeake Bay. The data were from the warm season of a five-year period (March through September, 2001 through 2005). Bay-breezes tend to have limited inland penetration and bay-breeze days peak in June and August. Bay-breeze days are shown to usually occur under conditions of weaker zonal flow and stronger thermal contrast compared to non-bay-breeze days. A traditional index based on the above-mentioned variables was evaluated for prediction of bay-breeze days and was found to be only modestly reliable, with a tendency to over-predict the number of bay-breeze days. Discussion is devoted to the theoretical improvement of the index. Finally, recommendations for future work are provided, focusing on potential methods for testing the revised index on a bay-breeze dataset.

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1. Introduction

Chesapeake Bay (see Fig. 1) is the largest estuary in the United States. Approximately 330 kilometers in length, it stretches from the mouth of the Susquehanna River at Havre de Grace, MD to the Atlantic Ocean at Virginia Beach, VA. The western shore of Chesapeake Bay is in close proximity to the Baltimore, MD – Washington, DC urban corridor. Thus, its mesoscale meteorological phenomena often impact a large segment of society. One such phenomenon is the thermally-driven Chesapeake Bay bay-breeze, which is in the same family of phenomena as the more documented sea- and lake-breeze. Following Segal and Pielke (1985), these types of atmospheric circulations will be referred to as water body (WB)-breezes.

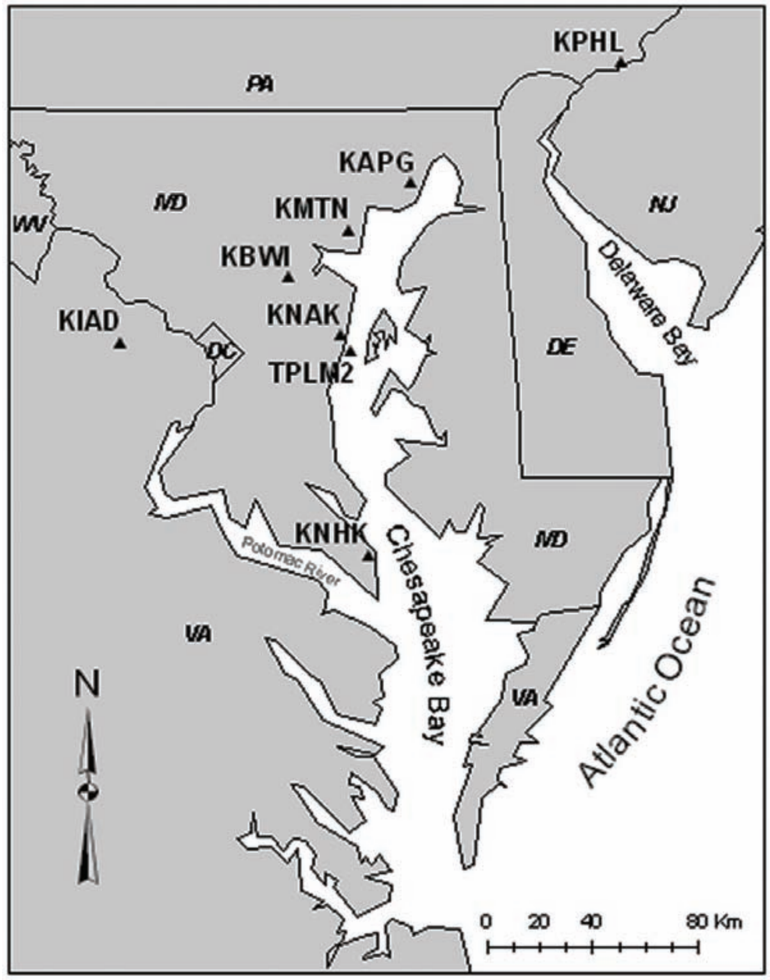


Fig. 1. Chesapeake Bay region. Darkened triangles indicate the locations of surface stations referenced in the text.

WB-breezes have been well-studied both empirically (e.g., Fisher 1960; Barbato 1978; Zhong and Takle 1992; Laird et al. 1995; Atkins and Wakimoto 1997; Laird et al. 2001; Miller and Keim 2003; Azorin-Molina et al. 2009) and numerically (e.g., Haurwitz 1947; Estoque 1962; Walsh 1974; Bechtold et al. 1991; Arritt 1993; Porsen et al. 2007). See Simpson (1994) and Miller et al. (2003) for thorough reviews of WB-breezes. Fundamentally, a WB-breeze develops in response to the hydrostatic pressure gradient resulting from the thermal contrast between air over a land mass and that over an adjacent body of water. During the daytime, air directly above the land warms more rapidly than air directly above the water. As a result, a low-level pressure gradient acceleration is directed perpendicular to the shoreline from water toward land, forcing the WB-breeze as a low-level onshore flow. Some studies have documented a subsequent veering of this flow owing to the Coriolis acceleration while others have not found a veering (see the discussion on this topic in Banta et al. (1993)). The strength of the corresponding WB-breeze frontal zone and its inland penetration are largely dependent on its strength relative to the shore-perpendicular component of the synoptic-scale wind (e.g., Segal and Pielke 1985; Atkins and Wakimoto 1997; Laird et al. 2001; Miller and Keim 2003, Porson et al. 2007). An offshore synoptic-scale wind tends to strengthen the frontal zone but at the same time retards the front's inland progress (Porson et al. 2007). Indeed, a sufficiently strong offshore synoptic-scale wind can overpower the sea breeze entirely (e.g. Fig. 3 and Eq. 5 of Porson et al. 2007). These dynamics suggest that the controlling parameters for WB-breeze existence and intensity include variables related to the horizontal difference in hydrostatic pressure (i.e., the overland - overwater temperature contrast, the depth of this contrast, gravitational acceleration, and average of the overland and overwater air temperatures) and the strength of the synoptic-scale wind component perpendicular to the coast.

The passage of a WB-breeze front is typically accompanied by a temperature leveling or decrease and a wind shift to an onshore component with some increase in wind speed (e.g., Laird et al. 2001). Aloft, some studies have documented a return flow, attributed to mass continuity (e.g., Atkins and Wakimoto 1997) while others have not found a return flow (e.g., Banta et al. 1993).

The Chesapeake Bay bay-breeze is particularly important because it has the ability to affect the sensible weather (e.g., air temperature, wind vector) experienced by the several million people who live within its reach. WB-breezes such as that associated with Chesapeake Bay can play a role in regional visibility, cloud cover (e.g., Segal et al. 1996), pollutant transport (e.g., Lyons and Olsson 1973) and thunderstorm development (e.g., Nicholls et al. 1991). Because of its effect on the wind vector, the bay-breeze impacts flight patterns at local civilian airports (e.g., Baltimore/Washington International Airport) and military bases (e.g., Naval Air Station Patuxent River), and can also influence the operations of recreational, commercial, and military small craft.

Because of the wide array of potential bay-breeze impacts, both civilian and military weather forecasters in the Chesapeake Bay region, including those focused on threat reduction, are sensitive to its occurrence. The goal of this modest study is to provide a climatological frequency of, and analysis of the near-surface meteorological conditions associated with, the western shore Chesapeake Bay bay-breeze. To the best of the authors' knowledge, there have been no refereed published reports detailing this phenomenon.

2. Methodology

The methodology used to identify bay-breezes loosely follows the related Lake Michigan lake-breeze research of Laird et al. (2001) and Atlantic Ocean sea-breeze research of Miller and Keim (2003). This study is focused on the morning and afternoon hours of March through September during the years 2001 through 2005. To identify bay-breeze events (i.e., the passage of a bay-breeze front), hourly surface data at three automated surface observing system (ASOS) stations within 18 km of the Maryland western shore of Chesapeake Bay (Baltimore-Washington International Airport [KBWI], Martin State Airport [KMTN], and Naval Air Station Patuxent River [KNHK]), were examined. The locations of these stations are identified in Fig. 1.

Other ASOS stations existed along the western shore of Chesapeake Bay during the research period. However, those stations were not employed for various reasons. For example, ASOS stations along the Virginia western shore of Chesapeake Bay are in close proximity to the Atlantic Ocean and were excluded because they are susceptible to Atlantic Ocean sea-breezes. Observations from the ASOS station at Phillips Army Air Field in Aberdeen, MD (KAPG, see Fig. 1) were too sporadic to be of value. And, data from the United States Naval Academy in Annapolis, MD (KNAK, see Fig. 1) were not archived prior to May 2004.

Corresponding marine meteorological data were garnered from a Coastal-Marine Automated Network (C-MAN) station at Thomas Point Light House (TPLM2, see Fig. 1). For ambient comparisons (i.e., synoptic; those not influenced by the bay-breeze), data from the ASOS at Washington Dulles International Airport (KIAD, see Fig. 1) in Virginia were examined. This site lies approximately 80 km inland from Chesapeake Bay, and is not susceptible to the meteorological impacts of the bay-breeze or similar mesoscale processes associated with the Potomac River. Table 1 contains geographic data for each station employed herein (site data taken from http://weather.noaa.gov/tg/site.shtml and http://www.ndbc.noaa.gov/).

A lack of eastern shore Chesapeake Bay data for the period of this study made it impossible to systematically distinguish western shore bay-breeze events from Atlantic Ocean sea-breeze events. There is little if any documented evidence that addresses the frequency of Atlantic Ocean sea-breezes reaching the western shore of the Maryland portion of Chesapeake Bay. However, to the north in Pennsylvania, the sea-breeze typically only travels as far inland as the Philadelphia International Airport (KPHL, see Fig. 1)—a smaller travel distance than that needed for the sea breeze to reach our ASOS stations of interest in Maryland—approximately 10 to 12 days per year (A. Cope, Science Operations Officer, National Weather Service Weather Forecast Office Philadelphia/Mount Holly, 2006, personal communication).

Compared to WB-breezes associated with a Great Lake, such as Lake Michigan (e.g., Laird et al. 2001), the Chesapeake Bay bay-breeze is a relatively shallow event. Simple comparison of the scale of Lake Michigan with Chesapeake Bay reveals why. Thus, the bay-breeze on many occasions is not associated with a discernable satellite signature (S. Zubrick, Science Operations Officer, National Weather Service Weather Forecast Office Baltimore/Washington, personal communication, 2007).

STATION	LATITUDE	LONGITUDE	SITE ELEVATION ABOVE MEAN SEA LEVEL (m)	APPROXIMATE DISTANCE INLAND (km)
NAVAL AIR STATION PATUXENT RIVER (KNHK)	38.30°N	76.40°W	12	2
BALTIMORE, MARTIN STATE AIRPORT (KMTN)	39.33°N	76.42°W	7	5
BALTIMORE-WASHINGTON INTERNATIONAL AIRPORT (KBWI)	39.17°N	76.68°W	45	23
DULLES INTERNATIONAL AIRPORT (KIAD)	38.93°N	77.45°W	95	82
THOMAS POINT LIGHT (TPLM2)	38.90°N	76.44°W	0	N/A

Table 1. Geographic data for each station used in analyses.

Moreover, Zubrick (2007) reports that National Weather Service Weather Surveillance Radar, 1988, Doppler (WSR-88D) in the vicinity of the study area (e.g., that at Sterling, VA [KLWX]) often does not capture the shallow features of the bay-breeze. Thus, to detect bay-breeze events, only daily ASOS time series of dry bulb temperature, wind speed, wind direction, cloud cover, and precipitation were examined. Future research of the bay-breeze could employ the newly available Terminal Doppler Weather Radar (TDWR) such as that at Baltimore-Washington International Airport. According to Zubrick (2007), data from TDWR are much better at detecting the bay-breeze than either the WSR-88D or satellite imagery. TDWR data were not available for the bulk of the time period of this study.

A bay-breeze day was documented for a particular ASOS station if the following conditions occurred at that station in combination and commenced between 1000 and 1600 local time: a) a leveling or drop of the dry-bulb temperature from its diurnal trend; b) a shift in the average wind direction from either having an offshore component, being light and variable, or being calm, to having an onshore component (defined below) and lasting more than 2 hours; and c) a short burst or steady increase in wind speed. Following Sikora and Halverson (2002), the positive cross-bay direction is taken to be towards 094°. To reduce the possibility of mistaking a synoptic-scale phenomenon for a bay-breeze event, days with any of the following conditions were excluded from being considered bay-breeze days: a) average sky condition of broken or overcast during daylight hours (sunrise-sunset) at a station; b) corresponding onshore component of wind at KIAD; and c) precipitation within six hours prior to the onset of a possible event at a station.

3. Bay-Breeze Climatology

Figure 2 provides the monthly (March through September) average bay-breeze day frequency for the 2001 through 2005 period at KBWI, KMTN, and KNHK. As expected from the WB-breeze dynamics discussed in section one, bay-breeze days are usually more frequent the closer an ASOS station is to the shoreline. All three stations exhibit more-or-less the same seasonal pattern in monthly bay-breeze day frequency. Those frequencies generally increase from March to a maximum in June, and possess a secondary maximum in August. As will be shown below, the seasonal pattern in monthly bay-breeze day frequency evident in Fig. 2 is not mirrored in the corresponding monthly average near-surface meteorology.

Monthly average near-surface meteorological conditions associated with both bay-breeze and non-bay-breeze days are now examined to determine if the seasonal pattern of daily occurrence can be explained in terms of the climatological intensity of thermodynamic forcing and synoptic-scale winds. Figures 3a-c show the monthly average of the daily maximum inland air temperature minus bay water-surface temperature ($\overline{\Delta T_{\text{max}}}$) for the 2001 through 2005 period on bay-breeze days and non-bay-breeze days for KBWI, KMTN, and KNHK. The daily maximum inland air temperatures are from hourly KIAD ASOS data from 1000 LST to 1600 LST. Bay water-surface temperature data are from TPLM2 at 1900 UTC. Figures 4a-c show the monthly average of the absolute value of daily average inland (KIAD) cross-bay wind speed ($\overline{|U|}$) for the 2001 through 2005 period on bay-breeze days and non-bay-breeze days for KBWI, KMTN, and KNHK. Hourly ASOS data from 1000 LST to 1600 LST were used to calculate the daily average inland cross-bay wind speed. The absolute value of daily average inland cross-bay wind speed was employed because, recall, both strong onshore and offshore synoptic-scale flow has been shown to suppress WB-breezes. Keep in mind that the dates of bay-breeze occurrence vary from station to station. Thus, the non-bay-breeze day data shown in Figs. 3 and 4 also vary from station to station. Moreover, because the number of non-bay-breeze days far outweighs that of bay-breeze days

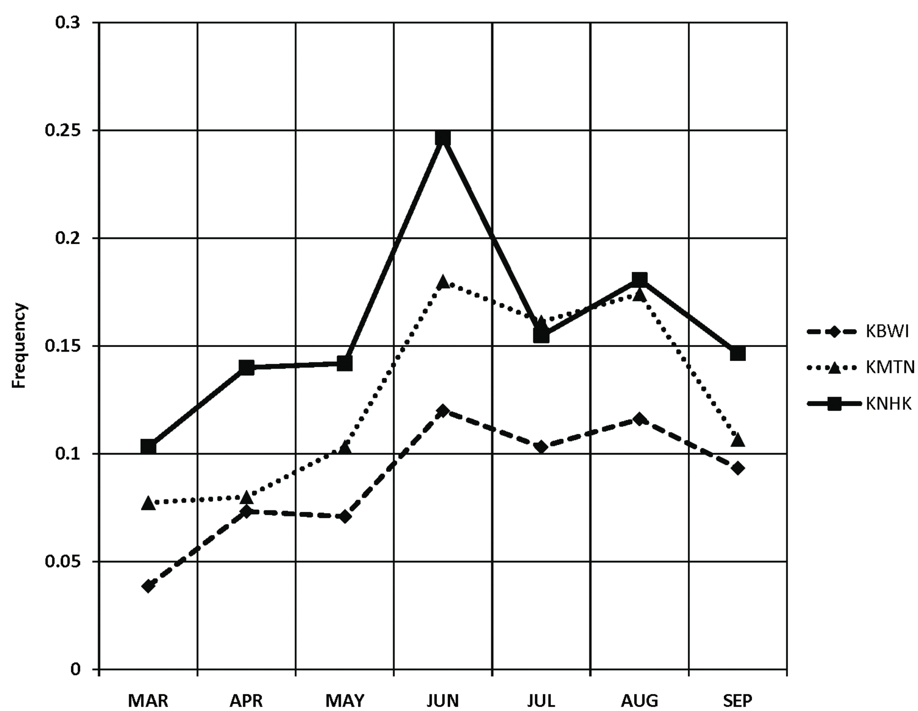


Fig. 2. Monthly average bay-breeze day frequency for the 2001 through 2005 period at KBWI, KMTN, and KNHK

(see Fig. 2), there is more inter-station variability to the bay-breeze day data seen within Figs. 3 and 4 than for the non-bay-breeze day data. Darkened symbols within Figs. 3 and 4 indicate monthly average pairs whose difference is statistically significant at the 95% confidence level via a two-sample T-test assuming unequal variances.

The data presented in Fig. 3 reveals that $\overline{\Delta T_{max}}$ is consistently greater on bay-breeze days than on non-bay-breeze days, as expected given the discussion within the Introduction. The difference between the two is usually statistically significant, with exceptions being March and September for KBWI and March for KMTN. This finding for the two most inland stations may be due to the exclusion of any bay-breeze days from the climatology due to the signatures of synoptic-scale weather systems listed in sections one and two, or to the impact of those systems' winds on inland penetration of the bay-breeze. The frequency of synoptic-scale weather systems is expected to be greater during the months of March and September, compared to the intervening months.

In general, the seasonal patterns for bay-breeze and non-bay-breeze days mimic each other at each station, with $\overline{\Delta T_{max}}$ peaking in April and decreasing to July. Thereafter, $\overline{\Delta T_{max}}$ continues to decrease for non-bay-breeze days at all stations while that for bay-breeze days is more variable between stations. For KBWI and KNHK, $\overline{\Delta T_{max}}$ increases slightly from July to August, before decreasing again in September, while for KMTN, $\overline{\Delta T_{max}}$ reaches a minimum in August. The monthly averages of ΔT_{max} for bay-breeze days fall within the range of that reported by other researchers (Miller et al. 2003). Taken together these results show the seasonal maximum in the climatological intensity of the thermal forcing occurs two months earlier than the seasonal peak in bay-breeze day frequency. Thus, additional variables related to the overland - overwater difference in hydrostatic pressure, the seasonal patterns of wind, or correlations between those variables, must also play a role in determining the seasonal pattern of bay-breeze frequency.

Figure 4 shows that $|\overline{U}|$ is generally smaller on bay-breeze days than on non-bay-breeze days (again, as expected from the discussion within section one), with exceptions being July and August for KMTN and August for KNHK. There are a larger number

Fig. 3(a)

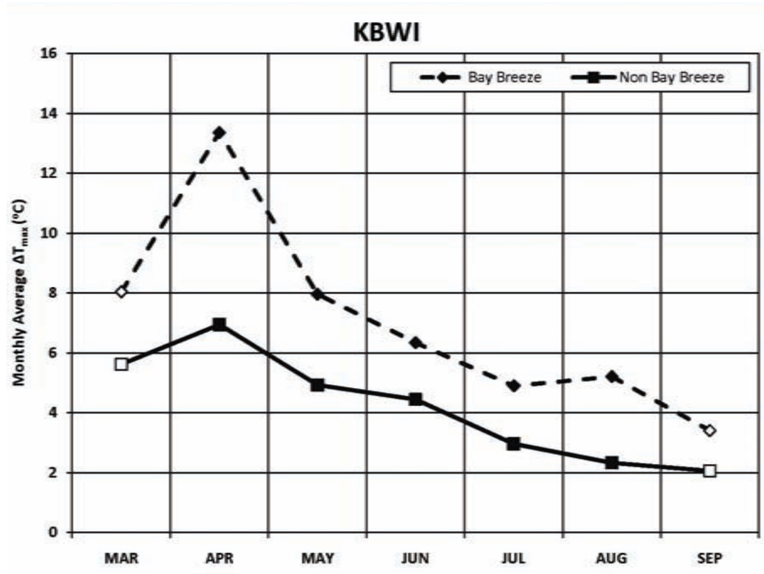


Fig. 3(b)

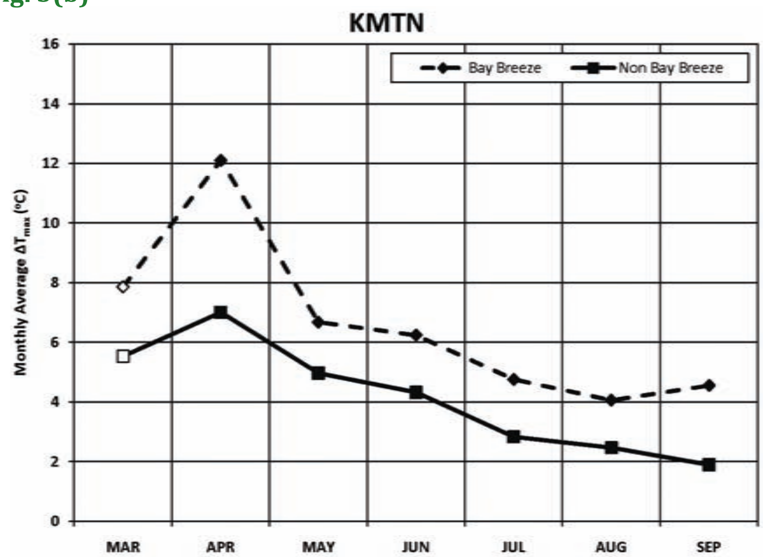


Fig. 3(c)

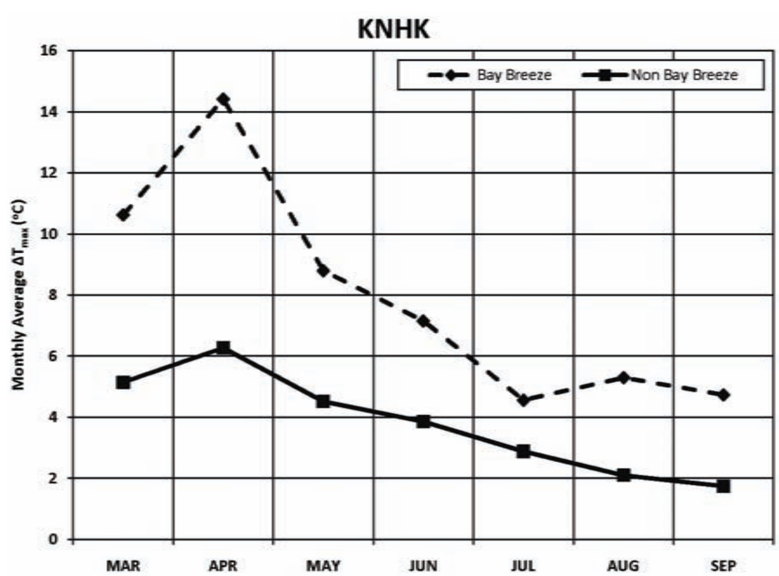


Fig. 3(a-c). Monthly average of the daily maximum inland (KIAD) air temperature minus bay water-surface temperature for the 2001 through 2005 period on bay-breeze days and non-bay-breeze days for KBWI, KMTN, and KNHK. Darkened monthly pairs are different with at least 95% confidence.

Fig. 4(a)

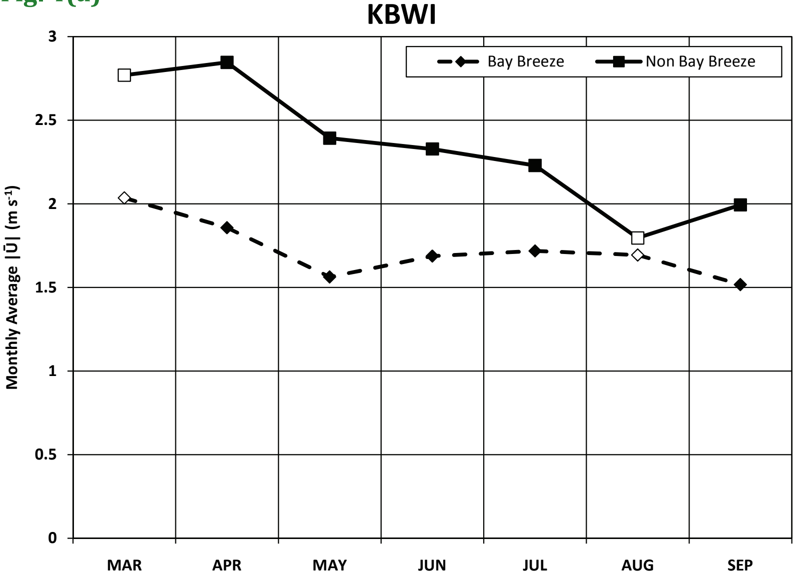


Fig. 4(b)

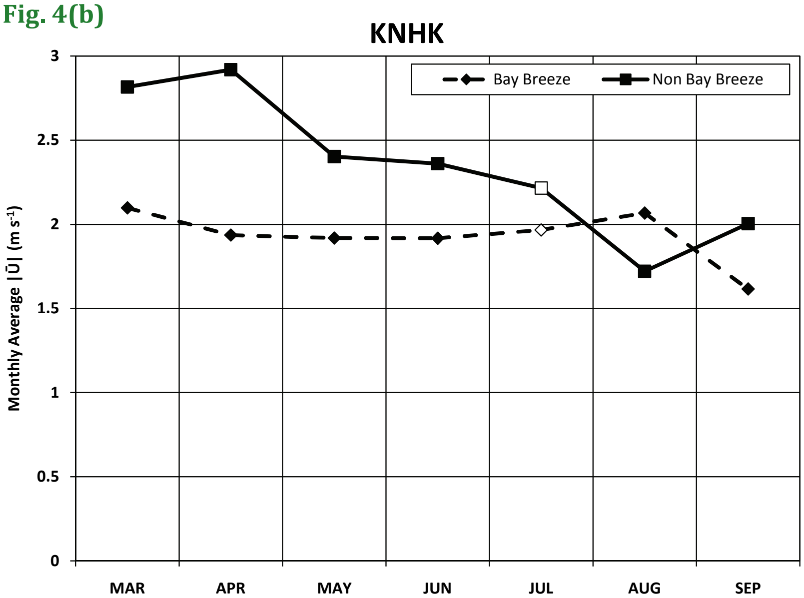
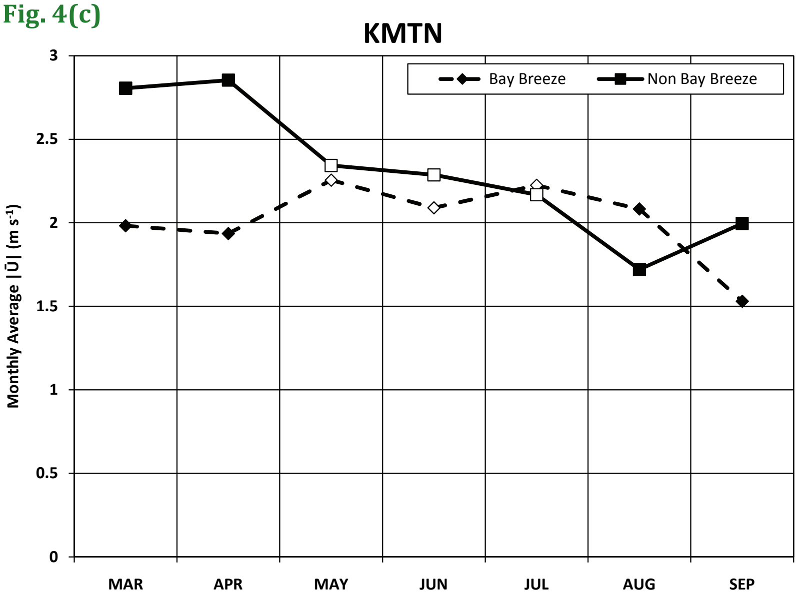


Fig. 4(c)



of differences that are not statistically significant compared to the results for $\overline{\Delta T}_{max}$. This implies that $|\overline{U}|$, alone, is at times a less-robust predictor of bay-breeze days than is $\overline{\Delta T}_{max}$. This is especially true within the weaker wind regime during the heart of the warm season, at stations closer to the shoreline. Unlike what was seen for $\overline{\Delta T}_{max}$, the seasonal patterns of $|\overline{U}|$ for bay-breeze days and non-bay-breeze days do not mimic each other at any station. For non-bay breeze days, for all stations, $|\overline{U}|$ peaks in April, then steadily decreases through the warm season to minimum value in August, before increasing in September. For bay-breeze days, $|\overline{U}|$ exhibits less of a discernable pattern at each station, oscillating between 1.5 m s^{-1} and 2.5 m s^{-1}. The implication is that values of $|\overline{U}|$ outside this range are sufficient to either eliminate, or prevent inland penetration of, the bay-breeze, given the hydrostatically-induced horizontal pressure gradient acceleration typical of the Chesapeake Bay region.

4. Bay-Breeze Prediction

The prediction of WB-breezes has long received considerable attention, with several researchers stressing the importance of a number of atmospheric variables in its development and demise, including the geostrophic wind speed (e.g., Lyons 1972), nonlinear advection processes (e.g., Walsh 1974), friction (e.g., Haurwitz 1947; Fisher 1960), and the topography and concavity of the coastline (e.g., McPherson 1970).

Perhaps the most frequently cited WB-breeze forecasting tool is that developed by Biggs and Graves (1962) who applied simple physics and dimensional analysis to determine the forces that contribute to lake-breeze events along portions of the Great Lakes. Their "lake-breeze index" (hereafter, L-B index) employs the relationship between the inertial acceleration, dominated by wind speed, and the hydrostatic pressure gradient acceleration that results from the temperature contrast between the air over land and the air over lake. Specifically, the following relationship was used to forecast lake-

Fig. 4(a-c). Monthly average of the absolute value of daily average inland (KIAD) cross-bay wind speed for the 2001 through 2005 period on bay-breeze days and non-bay-breeze days for KBWI, KMTN, and KNHK. Darkened monthly pairs are different with at least 95% confidence.

breeze events on the western shores of Lakes Erie and Michigan:

$$\varepsilon_{BG} = \frac{V^2}{C_p \Delta T_{max}}. \quad (1)$$

Here, V is the daily average surface wind speed (m s^{-1}), C_p is the specific heat of dry air at constant pressure (1.003 J K^{-1} gm^{-1}), and ε_{BG} is the L-B index. Oddly, ε_{BG} does not include the depth of the thermal contrast, gravitational acceleration, or the average of the overland and overwater air temperatures, elements necessary in addition to ΔT_{max} to compute the overland - overwater hydrostatic pressure difference. Realize, then, that the application of ε_{BG} characterizes the variability in the mean temperature averaged across the thermal front and depth of the thermal contrast from day to day as unvarying, with that unrealistic characterization resulting in forecast error. We next discuss the results of testing a slightly modified version of ε_{BG}. Then, we present a physically complete version of the index.

Following Laird et al. (2001), our slightly modified version of ε_{BG} (hereafter, ε) is calculated in the same manner as ε_{BG}, except the numerator is $|\bar{U}|^2$ instead of V^2 (for reasons already discussed). After calculating ε values for each day of the study period for which data were available, we grouped them by month, for bay-breeze and non-bay-breeze days, for KBWI, KMTN, and KNHK. An example of this analysis can be found in Fig. 5. If ε yielded an accurate prediction of bay-breeze days, all calculated values of ε for bay-breeze days would be located to the left of a critical curve, while that for non-bay-breeze days would lie to the right of that critical curve. At each of the three western shore ASOS stations, the vast majority of bay-breeze days correspond to an ε below 2.00, a threshold that could be used to alert operational meteorologists to the possibility of bay-breeze formation. However, as can be deduced from Fig. 5, ε tends to over-predict the number of bay-breeze days, since non-bay-breeze days are liberally distributed on both sides of this threshold curve.

Another way to look at the ability of ε to distinguish bay-breeze from non-bay-breeze days is via a hypothesis test along the lines of those summarized in Figs. 3 and 4. A non-parameteric test must be used, however, because ε is not normally distributed (a consequence of it being a ratio whose denominator frequently approaches zero). Thus, the Mann-Whitney test (Mann and Whitney 1947) of difference in medians is used instead of the T-test of difference in means. Test results by station and month are given in Table 2. Of the 21 combinations of stations and months, only four reach the 95% confidence level. Thus, for most of the analyzed data, the median values of ε do not differ significantly between bay-breeze days and non-bay-breeze days.

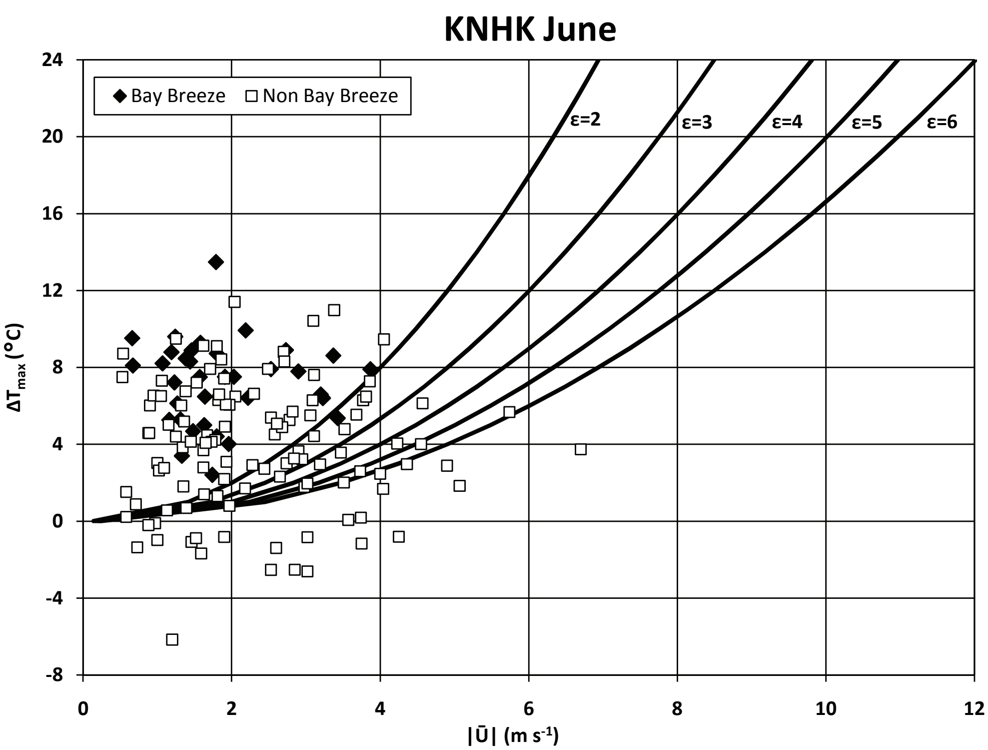


Fig. 5. Example of ε values for bay-breeze days and non-bay-breeze days. These data are for June at KNHK for the 2001 through 2005 period.

	KBWI	KMTN	KNHK
March	0.92	0.80	0.55
April	0.04	0.14	0.01
May	0.11	0.49	0.46
June	0.20	0.73	0.04
July	0.04	0.85	0.23
August	0.33	0.35	0.56
September	0.33	0.43	0.51

Table 2. Mann-Whitney test results, displayed as two-sided P values. Those reaching the 95% confidence level are in burgundy colored cells.

5. A Theoretical Improvement over ε

Our results for ε are in keeping with previous studies that attempted to employ such an index (see Miller et al. 2003). We next examine the poor performance of ε reported here and elsewhere in the context of its simplification of the dynamics it is meant to capture. Combining variables into a single non-dimensional index via the Buckingham Pi theorem (Buckingham 1914) is a common approach in atmospheric fluid mechanics (e.g., Stull 1988). In fact, that is how Biggs and Graves (1962) proceeded. Given that the WB-breeze results from the hydrostatic pressure gradient caused by the thermal contrast, the variables going into this non-dimensional group are $\Delta\langle T\rangle_{max}$, h, $g/\langle \overline{T}\rangle$, and $\langle|U|\rangle$, where $\langle\ \rangle$ indicates the layer-average through h, the depth of the bay-induced thermal contrast. $\Delta\langle T\rangle_{max}$ is the maximum difference between the layer-averaged overland air temperature and layer-averaged overwater air temperature, g is the gravitational acceleration, $\langle \overline{T}\rangle$ is the spatial average of the layer-averaged overland and overwater air temperatures, and $\langle|U|\rangle$ is the layer-average of the absolute value of the cross-shore component of the synoptic-scale wind. Thus, these variables fully account for the WB-breeze dynamics discussed in section one. The resulting non-dimensional index (i.e. group in the parlance of fluid mechanics) is

$$\gamma = \frac{\langle|U|\rangle}{\sqrt{\dfrac{g\Delta\langle T\rangle_{max}\, h}{\langle \overline{T}\rangle}}} \qquad (2)$$

with WB-breezes occurring when γ is less than some critical value. Because of surface drag and turbulent mixing, the critical value of γ is expected to be less than 1.

Comparing ε (and ε_{BG}) with γ, we see that while ε is dimensionally consistent, it amounts to a simplification of the γ index, a simplification based largely on the unstated assumption that all WB-breezes exhibit the same depth. Only to the extent that this extreme assumption holds is ε expected to provide skillful forecasts of WB-breeze formation.

6. Continuing and Future Work

Idealized numerical model experiments testing the applicability of γ have begun (Reen et al. 2009). Those experiments employed both the large eddy simulation (LES) described in Bryan and Fritsch (2002) and the Advanced Research Weather Research and Forecasting model (WRF-ARW) version 3.0.1.1 (Skamarock et al. 2008). Those experiments focused on the difference in surface buoyancy flux between adjacent surface areas, and the effect of varying background wind. The typical critical value found from those experiments was in the vicinity of 0.5, with the LES predicting slightly larger critical values than the WRF-ARW. Reen et al. (2009) speculate that the differences between the results of the LES and the WRF-ARW are, in part, a function of the vertical momentum flux. This idealized numerical modeling research is ongoing.

The testing of γ on a WB-breeze dataset, such as that described in section 3, requires knowledge of upper air conditions for the determination of h as well as $\Delta\langle T\rangle_{max}$, $\langle \overline{T}\rangle$ and $\langle|U|\rangle$. The operational radiosonde network is too spatially and / or temporally sparse to directly provide that information in most regions, including Chesapeake Bay. Therefore surface values of those variables could serve as substitutes, and a constant value could be assumed for h. However, these simplifications would introduce error along the lines of that previously discussed for ε.

The next simple option in the testing of γ on real data is the application of encroachment theory (Stull 1988). This theory assumes that the overland boundary layer at the time of maximum heating exhibits the dry adiabatic lapse rate of a convective mixed layer and that the vertical temperature profile above that convective mixed layer is that of the early morning sounding. Thus, to employ this theory in the calculation of γ, the requirements are an early morning sounding in the vicinity of the water body of interest and an estimate of the maximum surface air temperature at the location of the sounding. Because KIAD serves as the inland station for the dataset described in section 3, both of those requirements are met for the present research.

To use encroachment theory to predict h, one simply extrapolates the maximum surface air temperature upward along a dry adiabat until it intersects the 1200 UTC operational sounding. Similarly, the vertical profile of ΔT_{max} can be obtained by subtracting the 1200 UTC sounding (assumed to remain unchanged over the water) from that provided by encroachment theory at the time of the maximum surface air temperature. $\Delta\langle T\rangle_{max}$ is thus the average of this difference from the surface up to height h. The calculation of $\langle \overline{T}\rangle$ and $\langle|U|\rangle$ would proceed with similar ease. This approach offers the benefit of not requiring the radiosonde network to resolve the thermal contrast and the WB-breeze circulation, thus making it particularly useful for smaller bodies of water such as Chesapeake Bay.

Application of encroachment theory to the calculation of γ for a limited number of cases found in Table 1 yielded results that were less statistically significant than those found for ε. In the future, we plan to advance the testing of γ by following the approach of Shannon et al. (2002).

Shannon et al. (2002) employed a modified encroachment theory, by substituting multiple operational numerical weather prediction model soundings for the lone morning sounding called for in traditional encroachment theory. Thus, their approach has the advantage of capturing changes to the boundary layer vertical structure due to advection.

7. Summary

Hourly ASOS data were employed to identify days on which bay-breezes occurred at three stations west of Chesapeake Bay during March through September, 2001 through 2005. The three stations were Baltimore-Washington International Airport (KBWI), Martin State Airport (KMTN), and Naval Air Station Patuxent River (KNHK). Stations nearest Chesapeake Bay were generally found to experience a greater monthly average of bay-breeze day frequency than stations farther inland. That monthly average was maximized in June and possessed a secondary maximum in August at each station.

Hourly ASOS data from Washington Dulles International Airport (KIAD) and hourly data from the C-MAN station at Thomas Point Light House (TPLM2) were used to characterize the monthly average of the absolute value of daily average inland (KIAD) cross-bay wind speed ($\overline{|\overline{U}|}$) and the monthly average of the daily maximum inland air temperature minus bay water-surface temperature ($\overline{\Delta T_{max}}$) on bay-breeze days and non-bay-breeze days. $\overline{\Delta T_{max}}$ values were greater on bay-breeze days than on non-bay-breeze days. $\overline{|\overline{U}|}$ values were generally smaller on bay-breeze days than on non-bay-breeze days. While these results are in keeping with previously reported research on water-body breezes, the patterns of those monthly averages do not mimic that of the bay-breeze day frequency. Thus, it is reasonable to conclude that other variables play a role in determining the occurrence of a Chesapeake Bay bay-breeze.

After identifying bay-breeze days, a slightly modified predictive index (ε), based on that presented by Biggs and Graves (1962), was tested to investigate its potential use in forecasting bay-breeze days. ε is dependent on $|\overline{U}|$ and ΔT_{max}. ε was found to have only modest success in identifying days on which a bay-breeze occurred. Although most bay-breeze days correspond to an ε below 2.00, ε over-predicts the number of bay-breeze days, a problem also noted in earlier studies (see Miller et al. 2003).

It is argued that much of the lack of success of ε results from its failure to capture the full dynamics of water body-breezes. It is suggested that the input parameters for such an index should be the depth of the thermal contrast, the maximum difference between the layer-averaged overland air temperature and layer-averaged overwater air temperature, the gravitational acceleration, the spatial average of the layer-averaged overland and overwater air temperatures, and the layer-average of the absolute value of the cross-shore component of the synoptic-scale wind. A corresponding revised, non-dimensional index is proposed. A summary of ongoing idealized numerical model testing of that index is provided as are recommendations for its future testing on a WB-breeze dataset.

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