JP2.10 **ANALYSIS OF THE WESTERN SHORE CHESAPEAKE BAY BAY-BREEZE**

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1. INTRODUCTION

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.

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. 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). These dynamics suggest that the controlling parameters for WB-breeze existence and intensity include variables related to the horizontal difference in hydrostatic pressure 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). 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.

The goal of the present research is to provide a climatological frequency of, and analysis of the near-surface meteorological conditions associated with, the western shore Chesapeake Bay bay-breeze. A more detailed report on the present research can be found in Sikora et al. (2010).

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2. METHODS

The present research 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.

Corresponding marine meteorological data were garnered from a Coastal-Marine Automated Network (C-MAN) station at Thomas Point Light House (TPLM2). For ambient comparisons (i.e., synoptic; those not influenced by the bay-breeze), data from the ASOS at Washington Dulles International Airport (KIAD) 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.

To detect bay-breeze events, only daily ASOS time series of dry bulb temperature, wind speed, wind direction, cloud cover, and precipitation were examined. 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. To reduce the possibility of mistaking a synoptic-scale phenomenon for a bay-breeze event, days with any of the following conditions were not 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. CLIMATOLOGY

Figure 1 provides the monthly (March through

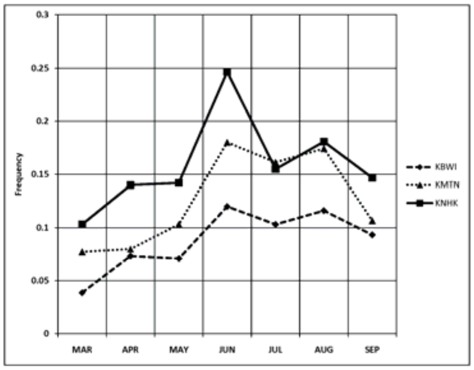

Fig. 1. Monthly average bay-breeze day frequency for the 2001 through 2005 period at KBWI, KMTN, and KNHK.

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 1, the frequency of bay-breeze days typically increases with proximity of the ASOS station 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. 1 is not mirrored in the corresponding monthly average near-surface meteorology.

Figures 2 shows the monthly average of the daily maximum inland air temperature minus bay water-surface temperature ($\overline{\Delta T_{max}}$) 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 3 shows 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 both strong onshore and offshore synoptic-scale flow has been shown to suppress WB-breezes. Darkened symbols within Figs. 2 and 3 indicate monthly average pairs whose difference is statistically significant at the 95% confidence level via a two-sample T-test assuming unequal variances.

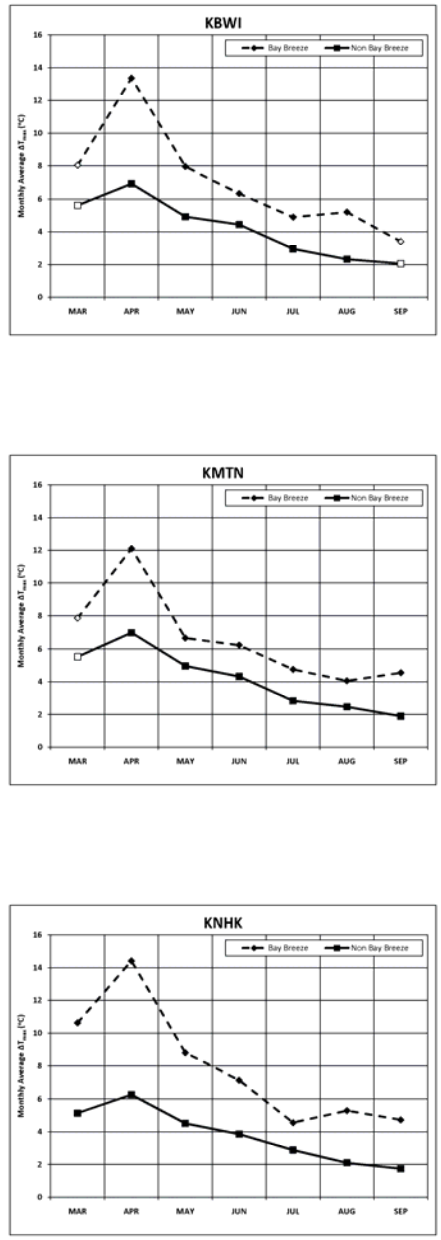

Fig. 2. Monthly average of the daily maximum inland (KIAD) air temperature minus bay water-surface temperature for the 2001 through 2005 period on bay-breeze days and non-bay-breeze days for KBWI, KMTN, and KNHK. Darkened monthly pairs are different with at least 95% confidence.

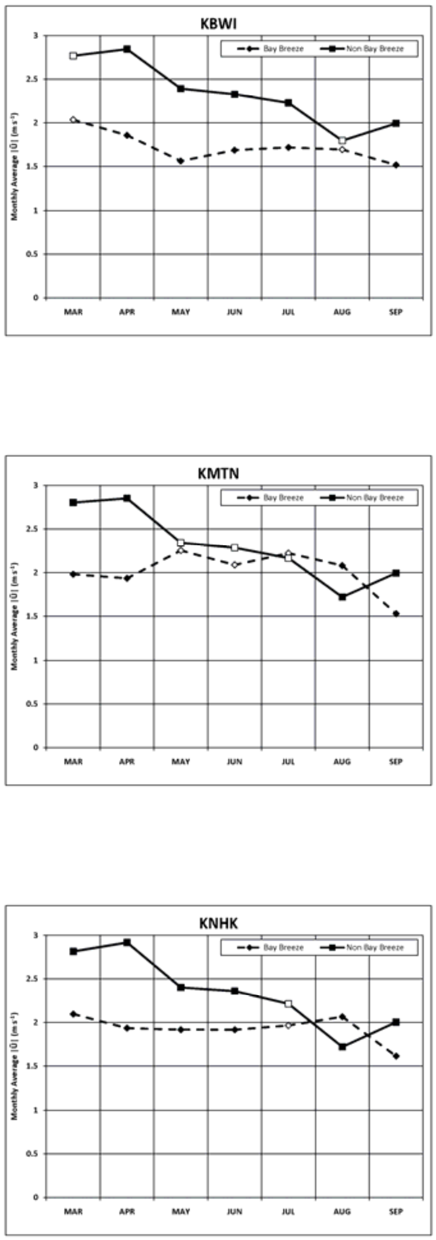

Fig. 3. Monthly average of the absolute value of daily average inland (KIAD) cross-bay wind speed for the 2001 through 2005 period on bay-breeze days and non-bay-breeze days for KBWI, KMTN, and KNHK. Darkened monthly pairs are different with at least 95% confidence.

The data presented in Fig. 2 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 Section 1. 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, or due 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 March and September than it is during the intervening months.

The monthly averages of $\overline{\Delta T_{\max}}$ 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 3 shows that $\left|\overline{U}\right|$ is generally smaller on bay-breeze days than on non-bay-breeze days (again, as expected from the discussion within Section 1), with exceptions being July and August for KMTN and August for KNHK. There are a larger number of differences that are not statistically significant compared to the results for $\overline{\Delta T_{\max}}$. This implies that $\left|\overline{U}\right|$, 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. For bay-breeze days, $\left|\overline{U}\right|$ oscillates between 1.5 m s^{-1} and 2.5 m s^{-1}. The implication is that values of $\left|\overline{U}\right|$ outside this range are sufficient to either eliminate, or to prevent inland penetration of, the bay-breeze given the hydrostatically-induced horizontal acceleration typical of the Chesapeake Bay region.

4. PREDICTION

Biggs and Graves (1962) 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 inertial forces, dominated by wind speed, and the hydrostatic pressure gradient force that result from the temperature contrast between the air over land and the air over lake:

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

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.

Realize 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.

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. Application of the Mann-Whitney test (Mann and Whitney 1947) of difference in medians revealed that the median values of ε do not differ significantly between bay-breeze days and non-bay-breeze days.

5. THEORY

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. Given that the WB-breeze results from the hydrostatic pressure gradient caused by the thermal contrast, the corresponding Buckingham Pi theorem (Buckingham 1914) variables are $\Delta\langle T\rangle_{\max}$, h, $g/\overline{\langle T\rangle}$, 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, $\overline{\langle T\rangle}$ 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. The resulting non-dimensional index (i.e., group in the parlance of fluid dynamics) is

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

with WB-breezes occurring when γ is less than some critical value.

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.

6. 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). These sensitivity 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 $\Delta\langle T\rangle_{\max}$, $\overline{\langle T\rangle}$, and $\langle|U|\rangle$. An option for testing γ on such a data set is to apply 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.

Application of encroachment theory using the KIAD 1200 UTC sounding to the calculation of γ for a limited number of our cases yielded results that were less statistically significant than those found for ε. In the future, we plan to advance that 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. Monthly average bay-breeze frequency was highest 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 ($\overline{|\bar{U}|}$) 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. $\overline{|\bar{U}|}$ 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 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 $|\bar{U}|$ and $\Delta T_{\max}$. ε was found to have only modest success in identifying days on which a bay-breeze occurred.

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. 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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References

Atkins, N.T., and R.M. Wakimoto, 1997: Influence of the synoptic-scale flow on sea breezes observed during CaPE. *Mon. Wea. Rev.*, 125, 2112–2130.

Biggs, W.G., and M.E. Graves, 1962: A lake breeze index. *J. Appl. Meteor.*, 1, 474–480.

Bryan, G.H. and J.M. Fritsch, 2002: A benchmark simulation for moist nonhydrostatic numerical models. *Mon. Wea. Rev.*, 130, 2917-2928.

Buckingham, E., 1914: On physically similar systems; illustrations of the use of dimensional equations. *Phys. Rev.,* 4, 345–376. doi:10.1103/PhysRev.4.345.

Laird, N.F., D.A.R. Kristovich, X.-Z. Liang, R.W. Arritt, and K. Labas, 2001: Lake Michigan lake breezes: climatology, local forcing, and synoptic environment. *J. Appl. Meteor.*, 40, 409–424.

Mann, H.B., and D.R. Whitney, 1947: On a test of whether one of two random variables is stochastically larger than the other. *Ann. Math. Statist.*, 18, 50–60.

Miller, S.T.K., and B.D. Keim, 2003: Synoptic-scale controls on the sea breeze of the central New England Coast. *Wea. Forecasting*, 18, 236-248.

__________, __________, R.W. Talbot, and H. Mao, 2003: Sea breeze: Structure, forecasting, and impacts. *Rev. Geophys.*, 41, 1011, doi: 10.1029/2003RG000124.

Porsen, A., D. G. Steyn, and G. Schayes, 2007: Formulation of an index for sea breezes in opposing winds. *J. Appl. Meteor. Clim.*, 46, 1257-1263.

Reen, B.P., D. Tyndall, G.S. Young, and D.R. Stauffer, 2009: Idealized simulations of circulations forced by land surface heterogeneity. Preprints, *23rd Conf. on Weather Analysis and Forecasting / 19th Conf. on Numerical Weather Prediction*, Omaha, NE, Amer. Meteor. Soc., CD-ROM, 10A.2.

Segal, M., and R.A. Pielke, 1985: The effect of water temperature and synoptic winds on the development of surface flows over narrow, elongated water bodies. *J. Geophys. Res.*, 90, 4907-4910.

Shannon, H.D., G.S. Young, M.A. Yates, M.R. Fuller, and W.S. Seegar, 2002: Measurements of thermal updraft intensity over complex terrain using American White Pelicans and a simple boundary-layer forecast model. *Bound.-Layer Meteor.*, 104, 167-199.

Sikora, T.D., G.S. Young, and M.J. Bettwy, 2010: Analysis of the western shore Chesapeake Bay bay-breeze. *Natl. Wea. Dig.*, in press.

Skamarock, W.C., J.B. Klemp, J. Dudhia, D.O. Gill, D.M. Barker, M.G. Duda, X.-Y. Huang, W. Wang, and J.G. Powers, 2008: A description of the advanced research WRF Version 3. NCAR Tech. Note NCAR/TN-475+STR, 113 pp. [Available from National Center for Atmospheric Research, P.O. Box 3000, Boulder, CO 80307-3000.]

Stull, R.B., 1988: *An Introduction to Boundary Layer Meteorology*. Kluwer Academic, 666 pp.