Numerical Model Simulation of Human Biometeorological Heat Load Conditions—Summer Day Case Study for the Chesapeake Bay Area

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ABSTRACT

During the warm season the spatial and temporal behavior of the biometeorological heat load in coastal regions is highly complex. Observational studies attempting to describe it usually are incomplete due to the scarcity of available meteorological data. The use of a mesoscale numerical model to overcome these disadvantages is suggested.

The mesoscale numerical model of the University of Virginia has been applied to the Chesapeake Bay region in order to study a typical severe heat load episode during a summer day. The possibilities offered by this method are illustrated and discussed. Results indicate a general agreement with the available observed data.

1. Introduction

The biometeorological interaction of people and the environment during the warm season has been expressed through various thermal indices, e.g., Fanger (1970), Munn (1970) and Givoni (1974). These indices attempt to evaluate comfort and thermal stress by relating them to meteorological conditions, with several also considering physiological conditions and type of clothing. Following ASHRAE (1966), the thermal comfort is defined as "that condition of mind which expresses satisfaction with thermal environment" while thermal stress can be evaluated (Givoni, 1974) "by the magnitude of deviation of any of the physiological systems or sensory manifestation of stress from their baseline at comfort state."

Regional development considerations and recreational plans should respond to this bioclimatological aspect of our environment, while the accurate prediction of daily heat conditions is of major importance to the public during the warm season, particularly as the use of air conditioning is reduced due to the limitations on available electric generating capacity.

The meteorological factors which are used in the common thermal indices are air temperature, air humidity, air velocity and sunshine. In the indoor environment air temperature and humidity are the most important factors, while for the outdoor environment it is essential to account for sunshine and wind speed. A very comprehensive and detailed discussion of the effects of these meteorological factors in the assessment of sultriness is given by Steadman (1979a,b). For coastal regions, the typical gradual onshore penetration of the sea breeze during warm and predominately clear days, in addition to the coupling with synoptic flow, results in pronounced spatial and temporal variability in the wind field. An increase of temperature accompanied by a decrease of absolute humidity inland from the coast is also typical. Because of these variations in space and time, a large amount of data is needed for the observational description of the bioclimatological heat load regimes in these areas. In practice, however, the available information is usually too poor for a detailed presentation of these patterns.

Up to now no emphasis has been placed on the use of numerical mesoscale models for biometeorological studies. However, in recent years, computer power has increased sufficiently to enable the use of accurate three-dimensional mesoscale models [e.g. Pielke (1974); Tapp and White (1976); Anthes and Warner (1978)]; hence offering a new possible method for these studies. The advantages in using such models lie in their ability to supply highly resolved spatial and temporal patterns of all meteorological parameters common in the various thermal indices.

The utilization of mesoscale numerical models for biometeorological purposes has the following goals.

- Simulation of regions which have persistent large-scale weather patterns conducive to excessive climatological heat loads. This is the primary goal.
- Obtaining estimates of the spatial and temporal variations of bioclimatological features to be used in establishing an observation network.
Supporting daily heat load forecasting by referring to results of simulations with various typical large-scale flows that are associated with severe heat conditions.

The greater Chesapeake Bay area (which includes the Delmarva peninsula, eastern Virginia and east-central Maryland) consists of a complicated series of embayments and estuaries. During the warm season, the large and persistent subtropical ridge known as the Bermuda High often produces weather conditions over this region associated with thermal discomfort conditions, particularly during the daylight hours. Observational studies by Scofield and Weiss (1977) have indicated that complex thermal patterns occur over this area during the summer season as a result of this juxtaposition of land and water, producing a spatially and temporally complex flow regime in response to the differential heating between land and water. A numerical simulation for this area (Warner et al., 1978) for a summer day without synoptic flow illustrated the expected complex patterns of sea breeze development under this particular synoptic condition.

The reasonable results obtained in verification analysis of the surface meteorology over south Florida (Pielke and Mahrer, 1978) and over Israel (Segal and Mahrer, 1979), using the University of Virginia mesoscale numerical model, suggested that this model can also be applied to the greater Chesapeake Bay area in order to evaluate mesoscale bioclimatological heat load conditions.

The presentation of the thermal indices used in this study and their properties are given in Section 2 followed by the climatological and synoptic description in Section 3. The model is described briefly in Section 4, while the numerical aspects are given in Section 5. The modifications in the model initialization needed to treat the inhomogeneous large-scale flow in this study are given in Section 6. The presentation of the results and the error analysis are given in Sections 7 and 8.

2. The thermal indices

Unfortunately, a thermal index is not available which is simple to apply, yet sufficiently comprehensive to include all the relevant meteorological information needed for both the indoor and outdoor environments. [The thermal stress index (Givoni, 1974) includes this information; however, it is limited to wind speeds \( \leq 1 \text{ m s}^{-1} \).] It is, therefore, necessary to utilize thermal indices which can be applied separately to each of these environments.

We have thus adopted the Temperature-Humidity Index (THI) as representing indoor conditions, while the skin temperature estimate is selected to represent the outdoor climate. The THI contains contributions from the wet- and dry-bulb temperatures, while the skin temperature estimate includes the dry-bulb temperature, wind speed, percentage of sunshine, and parameters which represent the activity level and clothing of an individual.

\[ T_x = C_1 T_d + C_2 T_w + C_3 T_g + C_4, \]

where \( T_d, T_w \) and \( T_g \) are the dry-bulb, wet-bulb and globe temperatures, respectively, while \( C_i \) are constants. For the outdoors the indices often take into account the globe temperature as equivalent to the equilibrium radiative temperature (net radiation). According to Steadman’s review, ASHRAE recommendations for comfortable conditions imply that the relative weighting \( C_1:C_2 \) given to dry- and wet-bulb temperatures varies over the range 84:16 to 88:12 while in Canada the Humidex corresponding to \( C_1:C_2 = 47:53 \) has been adopted. Following Sohar et al. (1963) \( C_1:C_2 = 50:50 \) is used in bioclimatological studies of the Mediterranean climate of Israel.

In this study we have chosen the THI which is common in the United States (Baldwin, 1973). This index has been defined originally by Thom (1959) and is based on the dry- and wet-bulb temperatures. Adopting \(^\circ\text{C} \text{ scaling of the THI we have defined} \)

\[ \text{THI} = 0.4(T_d + T_w) + 4.78 \text{ [}^\circ\text{C}] \]

Using (2) and following Baldwin, relatively few people will be uncomfortable when the index defined by this relation is \( 21.1^\circ\text{C} \) or lower, while about half of the population will be uncomfortable when the index reaches \( 23.9^\circ\text{C} \). Above \( 26.7^\circ\text{C} \) almost everyone will be uncomfortable.

b. Skin temperature

Petterssen (1969) suggests evaluating the sensation of heat by an individual using skin temperature. His estimation for the skin temperature is given by

\[ T_s = T_d + 0.034IM + \frac{0.24M - 15 + 120S(1 - A)}{2 + 9\sqrt{0.1 + V}}, \]

where

\( V \) wind speed (m s\(^{-1}\))
\( S \) sunshine, according to a scale from 0 to 1
\( A \) albedo of clothing; scale 0 to 1
\( M \) metabolic rate (W)
\( I \) thickness of clothing (cm).

For light clothing, for example, \( I \) is 0.5 cm while it is
1.0 cm for a heavy sweater. The albedo varies from
0 (black clothing) to 0.7 (white clothing). The meta-
bulb rate is about 100 W for a resting person, 165
during light exercise, 420 for walking at a moderate
pace and 830 or more for very strenuous exercise.
Comfort is felt when the skin temperature is be-
tween 31 and 35°C, with 33°C as optimal for most
people.

Under evaporative cooling conditions induced by
sweating (when \( T_s \) is above 35°C), the tempera-
ture of the air in contact with the skin approaches
that of a wet-bulb thermometer; hence the tempera-
ture of the wet-bulb rather than of the dry-bulb
thermometer is approximately the relevant one.
However, we have not considered this modification
in this paper; therefore, \( T_s \) values greater than
35°C, although indicative of the need for sweating,
are not representative of actual skin temperature
obtained under such conditions. Skin temperatures
greater than 35°C can, therefore, be considered
indicative of the potential thermal stress.

It is worth mentioning that Fanger (1970) evaluates
the thermal comfort for the indoor climate by intro-
ducing constraints on the skin temperature and
sweat secretion; however, the solution of a set of
equations is required. This added complexity is
not desires in this first experiment on mesoscale
variations in heat load. In simulations presented
in this study the following values of variables are
assumed: \( M = 165 \) W, person doing light exercise;
\( I = 0.5 \) cm, thin clothing; \( A = 0.45 \), normal clothing.

3. Climatological and synoptic description

During the summer season the eastern area of the
United States is affected by the semi-permanent
Bermuda High. Consequently, mostly sunny days
characterized by a weak southwesterly flow are
frequent in the coastal area. This overland synoptic
flow transports a large amount of moisture north-
eastward from the Gulf of Mexico. Because the
large-scale flow is offshore, the onshore penetra-
tion of the relatively cool sea breeze air is limited
to a narrow strip along the coast. With the coupling
of the synoptic flow and the sea breeze, convergence
zones are created inland near the coast which
have poor ventilation and, therefore, a local in-
crease of temperature and humidity results. This
leads to the creation of severe heat load episodes

![Fig. 1. Synoptic surface pressure pattern for 0700 LST 21 July 1978 (based on National Weather Service analysis).](image-url)
suggested by Deardorff (1974). The roughness parameter over the land is prescribed, while over water it is calculated according to Clarke (1970).

b. Surface heat balance

The temperature at the soil-air interface is calculated using a heat balance equation for solar radiation, incoming atmospheric longwave radiation, latent, sensible and soil heat fluxes, and the outgoing surface longwave radiation.

c. Radiation

The changes of air temperature due to shortwave and longwave radiative flux divergence are parameterized following the methods of Atwater and Brown (1974). Heating of the atmosphere by shortwave radiation is confined to water vapor, while carbon dioxide and water vapor are considered in the longwave radiation heating/cooling algorithm.

5. Numerical aspects

The governing hydrostatic equations of motion, heat, moisture and continuity are approximated in a finite difference form using a 60 s time step, with a horizontal grid interval of 10 km. Details of the computational scheme used in the model are discussed most recently in Mahrer and Pielke (1978). In the vertical the atmosphere, up to the model top at 7000 m, is divided into 13 layers. The bases of the layers are at the following heights: 0, 5, 15, 100, 300, 500, 700, 900, 1200, 1500, 2000, 3000 and 5000 m. Starting at 0600 LST the model is integrated for 12 h.

6. Initial conditions and input parameters

Because of the significance of the interaction between seabreeze and synoptic-scale flows in the determination of surface layer thermal properties, emphasis should be placed on the accurate initialization of the large-scale flow pattern. In previous studies using the model over flat terrain, a horizontally constant geostrophic flow has been introduced initially above the PBL. In the PBL the Ekman profile of velocity, \( \mathbf{v} \), was determined by the solution of the balance equation:

\[
\frac{\partial}{\partial z} \left( K_z \frac{\partial \mathbf{v}}{\partial z} \right) - f \mathbf{k} \times (\mathbf{v} - \mathbf{v}_0) = 0, \tag{4}
\]

where \( K_z \) is the vertical exchange coefficient of momentum, \( \mathbf{v}_0 \) the geostrophic wind at the top of the PBL, \( f \) the Coriolis parameter and \( \mathbf{k} \) the vertical unit vector.

The observed spatial variation of the large-scale flow in this study, however, necessitated a modification in the model initialization procedure. Available model initialization methods include those of Sasaki (1970) who suggested a variational analysis procedure which minimizes errors between the observed values and analyzed ones, and of Hoke and Anthes (1976) who describe a dynamic initialization procedure where the model equations include terms which nudge the prediction toward the observation as the model develops the quasi-balanced fields which are used to commence a model simulation. These procedures, however, are desirable only when sufficient observed data are available. With only a few observations, non-representative values in the measured fields can cause the generation of erroneous initial fields for the model unless some assumption such as geostrophy or other large-scale balanced wind field is assumed. Hence, in the absence of sufficient data, a different form of dynamical initialization has been adopted. In this approach, adopted here, the spatial variations in the synoptic flow have been incorporated into the model through an objective interpolation based on the wind velocity available from the 0600 LST radiosonde data for Wallops Island and Dulles International Airport, which are the only radiosonde stations within the domain (their locations are fortunately far enough apart to give a general representation of the spatial changes in the synoptic flow pattern over the domain). For both of these
sites the Ekman wind profile has been derived using (4). The initial height of the PBL was set at 250 m as estimated from the radiosonde soundings at both sites. Above the PBL the observed wind was assumed to be in geostrophic balance. The winds in the remainder of the model domain are then evaluated by an objective analysis using an inverse-squared distance interpolation formula

\[ u_p = \frac{u_p r_{pd}^{-2} + u_w r_{pw}^{-2}}{r_{pd}^{-2} + r_{pw}^{-2}}, \]  

(5)

where \( u_p \) refers to either horizontal component of velocity (the east-west and north-south components are evaluated separately) at the grid point \( p \), \( u_p \) and \( u_w \) are the corresponding observed velocities at Dulles Airport and Wallops Island, and \( r_{pd} \) and \( r_{pw} \) are the distances of the interpolated point from Dulles Airport and Wallops Island.

Temperature and moisture observations could be treated in a similar fashion. However, as shown by Hoke and Anthes, the proper initialization of winds is much more important on the mesoscale than that of temperature. Our previous numerical experimentation has shown that differences between locations within the boundary layer of temperature and moisture are quickly removed after sunrise as enhanced vertical mixing between the surface layer and free atmosphere commences. For these reasons, the temperature and dew-point profiles at Wallops, as given in Fig. 3, were used in the initialization.

Using the thermodynamic sounding given in Fig. 3 along with the objectively analyzed wind field, the model equations were integrated for 3 h of simulated time in order to obtain the adjusted fields of the dependent variables. Diurnal heating was introduced after this initialization procedure was completed. Vertical velocities resulting from this initialization in an east-west cross section through Wallops, for example, were less than 0.4 cm s\(^{-1}\). The surface wind field obtained by this procedure is shown in Fig. 4.

The specific humidity near the ground \( q_g \) is weighted according to

\[ q_g = F_w q_g |_{\text{saturated}} + (1 - F_w) q(1), \]  

(6)

where \( q(1) \) is the specific humidity at the first level (10 m), and the moisture parameter \( F_w \) is set equal to 0.1. The roughness parameter is prescribed as 10 cm over the land surface with a surface albedo of 0.2. The water surface temperature is assumed to be 22°C. Other parameters needed for the model calibration are given in Pielke and Mahrer (1978).

7. Results

In this section the model predictions for the fields of surface wind velocity, temperature, Temperature-Humidity Index and skin temperature are given for 0900, 1200, 1500 and 1800 LST. These hours are considered to be representative of the thermal discomfort period during the day. In addition, comparisons
of the calculated and observed thermal indices are depicted and an analysis of model error performed.

a. Surface wind field

Because of the important role of the sea breeze in the determination of thermal heat load conditions, it is valuable to describe its variations during the day as shown in Fig. 5. At 0900 LST the sea breeze does not yet dominate the prevailing synoptic flow along the coast. However, the sea breeze interaction with the offshore synoptic flow along the northeast Delmarva coastline and along the southern Virginia coast is already evidenced by a wind speed reduction to values below 1 m s\(^{-1}\). By 1200 LST a sea breeze flow with wind speeds up to 5 m s\(^{-1}\) prevails along a coastal strip with a typical width of about one grid interval (10 km) along the coasts of the northeastern Delmarva peninsula, southern Virginia, and at several locations along the western shore of the Chesapeake Bay. Inland from this strip poor ventilation occurs in the sea breeze convergence zones. These convergence zones are located ~20 km inland by 1500 LST [in a simulation for the Chesapeake Bay area without synoptic flow performed by Warner et al. (1978) for 21 June climatological conditions, the inland penetration of the sea breeze front was ~30–40 km by 1530 LST]. In contrast, over the western portions of the Delmarva peninsula the sea breeze convergence zone moves rapidly inland as the mesoscale and synoptic winds are in the same direction.

By 1800 LST the southeasterly sea breeze along the eastern Virginia coast and the southerly sea breeze along the coast of the western Chesapeake Bay interact with the southwesterly synoptic flow to create a relatively strong southerly flow which eliminates the poor ventilation area formerly located there. Elimination of calm areas to a certain extent is observed also over the northeast Delmarva peninsula.

From these wind fields the important role of the daily veering and intensification of the sea breeze in the creation and elimination of comparatively calm areas is evident. On the other hand, only small changes in the initial flow pattern, due to vertical mixing, are observed in the western areas, remote from the water.

A prominent feature of this stimulation is the intensified sea breeze flow over the southern Chesapeake Bay as compared with its northern narrower part. Also, local breezes develop along the wide sections of the Potomac and the James Rivers causing a reduction in wind speed along their southern banks. In the afternoon, these local minima of wind speed are eliminated due to the penetration and movement inland of the pressure trough associated with the “large-scale” sea breezes generated by the Bay and the ocean.

b. Temperature field

Fig. 6 presents the air temperature fields at a 2 m level (the standard meteorological shelter height). The large differences between the land and the water surface temperatures (calculated to be over 15°C around noon) results in sharp gradients in the air temperatures along the coasts of the Atlantic Ocean and the southern part of the Chesapeake Bay, with isotherms parallel to the shorelines. Along the central and northern Chesapeake Bay, where the water area is less, temperature gradients become weaker; however, the isotherms still follow the shoreline, a feature which can be seen also along the estuaries of the Potomac and the James Rivers.

At 0900 LST, temperatures are above 28°C over the inland regions as compared with typical values of 26°C along the shore. At 1200 LST the calm zones over the land are characterized by a relatively high temperature due to the reduction in wind speed and an increase in the ground surface temperature as vertical turbulent mixing of heat is reduced. The highest values are located in the Norfolk vicinity, where a maximum temperature of 35°C is predicted, which is higher than that calculated for the inland areas. This region of higher temperature (a type of “heat island”) has been dissipated by 1500 LST,
however, due to the sea breeze penetration and the resultant onshore advection of marine air, while an increase in temperature to values above 34°C is predicted over most of the inland domain. [Average July daily maximum temperatures are ~32°C for Richmond and 30.6°C for Washington (Climates of the States, 1974) so that the simulated day can be considered as somewhat hotter than normal.] By 1800 LST, as the sun's elevation angle decreases, temperature reductions up to 2°C are observed over the whole domain.

c. Temperature-Humidity Index (THI) fields

Fig. 7 shows the daylight patterns of the THI fields. Increases of the wet bulb temperature (which
Fig. 6. The predicted temperature field at 2 m height for 0900, 1200, 1500 and 1800 LST.

is considered in the THI calculations) are caused either by the warming or moistening of the air. Therefore, the differences in wet-bulb temperatures between the relatively cool but humid air along the shore and of relatively warmer but less moist air over the inland areas are smaller than predicted for temperatures. Consequently, THI gradients normal to sea shores are moderate compared to those calculated for temperature; therefore, according to the THI index, the difference in perceived dis-
comfort between the inland and coastal areas is not as sharp as one would conclude from the temperature fields only.

At 0900 LST the whole domain is already characterized by THI values regarded as uncomfortable, with maximum values above 25°C over the interior of the Delmarva peninsula and the inland regions of Virginia and Maryland. A general increase of more than 1°C which is attributed mainly to warming, is observed throughout the whole domain by 1200 LST. Over the poor ventilation area local maxima for THI are predicted.
Further increases of the THI values are observed by 1500 over the interior area and over those coastal areas where the sea breeze onshore penetration is slight (e.g., in the western shores of the central Chesapeake Bay). However, over the southern coasts of Virginia where the sea breeze is dominant, the decrease in temperature is apparently accompanied by an approximately equal increase in wet-bulb temperature, thus closely maintaining the earlier THI values. By 1800 THI values have fallen slightly, but severe heat load conditions still prevail over the entire land region.

It should be noted that available shelter-height temperature observations for the simulated domain are relatively dense. Unfortunately, such is not the case with the wet-bulb temperature. The latter meteorological parameter is only available from regular weather observing stations; thus, it is impossible without a model such as presented here to obtain the THI patterns over the area.

d. Skin temperature $T_s$

Following the discussion of the skin temperature $T_s$, given in Section 2b, it is assumed that values of skin temperature above $35^\circ C$ are associated with potential thermal stress. Therefore, a qualitative scaling of the potential thermal stress can be expressed by the measure of deviation from this value. On 21 July 1978 the skies were generally sunny so that $S$ in Eq. (3) was assigned its maximum value of 1 [the values of $M$, $I$ and $A$ required in (3) are given in Section 2b].

$T_s$ values higher than $35^\circ C$ prevailed over most of the domain as early as 0900 LST (Fig. 8). The role of low wind speed in accentuating discomfort was well pronounced by 1200 around Norfolk and over the northeast Delmarva peninsula where $T_s$ values of 48 and 46$^\circ C$, respectively, were calculated.

Under these conditions it is reasonable to assume that the body is covered with perspiration; hence replacing $T_d$ in (3) by $T_w$ will give the real skin temperature of the body. Then by using the relation $T_w - T_d = 2.5(THI - 4.78) - 2T_d$ and substituting $T_d = 33.2^\circ C$, THI = $27.5^\circ C$ the real skin temperature at the Delmarva maximum site is 36.4$^\circ C$, which indicates that sweating cannot maintain comfort.

By 1500 LST the high values near Norfolk were reduced and the location of the maximum shifted slightly to the south while an increase in discomfort is forecast over the northeast portion of the Delmarva peninsula, with the maximum shifting slightly to the northwest of its 1200 position. By 1800 a reduction of the $T_s$ values is observed throughout most of the domain, with the largest falls at locations of previous maxima. The sharp decrease of $T_s$ values over the northeastern Delmarva peninsula is attributed to the strengthening of the low-level winds as the sea breeze and the synoptic flow become less opposed in direction.

It must be emphasized in interpreting these predictions that the locations of the highest values of $T_s$ (and also THI) are temporary and cover small areas so that their detection via routinely available observational data is a very difficult task.

8. Verification analysis

A verification analysis has been performed by comparing the calculated against the available observed measurements of the thermal indices. An error analysis of the predicted fields is also undertaken.

Discrepancies between reality and the model simulations may be attributed to a wide range of causes, some of them due to the methods of collecting data. Therefore, not all disagreements between observed and calculated values should be considered as inadequacies in the model physics. The principal causes of the discrepancies are the following:

- It is impossible to obtain a resolution of the initial velocity, temperature and moisture fields which is entirely consistent with the grid spacing used in the model. In the model domain, radiosonde observations were only available at two points (Dulles and Wallops).
- Small-scale irregularities in the coastline (small bays, for example) may create local sea breeze circulations (Yoshino, 1974; McPherson, 1970), which are not adequately resolved with the 10 km grid. Furthermore, regions along the estuaries of the rivers may also develop a local river breeze. Scofield and Weiss present schematic flow patterns for the Chesapeake Bay area, based on cloud observations which illustrate this phenomenon. These local sea and river breezes can be dominant over a very limited area prior to the penetration of the "larger scale" sea breeze.
- Microclimate and urban effects which are not considered by the model.
- Occurrence of meteorological disturbances such as cloudiness which are not treated by the model.
- Since wind observations reflect short period averages of speed and direction, such nearly instantaneous measurements can be non-representative of average conditions over a longer time interval.

Only 12 sites in the model domain had sufficient meteorological information with enough spatial separation for verification of the thermal indices (observed data was at 0900, 1200, 1500 and 1800 LST and included temperature, humidity and wind speed). The available solar radiation records indi-
cated that 21 July 1978 was mostly sunny over the domain so $S$ was set to unity in Eq. (3) (during the day some cloudiness was observed over the area but the day was predominantly clear). The locations of stations used for the model verification are given in Fig. 2.

\textit{a. Temperature}

Similar qualitative patterns of the observed and calculated temperatures are obtained for most of the stations. Generally the discrepancies are within a range of 1°C and not larger than 2.7°C (Fig. 9).
Most stations are located sufficiently far from the large urban centers that possible urban heat island effects on the observed data, as pointed out by Scofield and Weiss for Baltimore during summer days, are largely avoided. Cloud effects of the incident solar radiation at the ground, which are not considered by the model, appear to be an important cause of model overpredictions at Patuxent River and Davison during the noon and afternoon hours. The amount of cloud cover (in terms of clear, scattered, or in tenths of coverage) is indicated every 3 h at the top of the figure.

b. Temperature-Humidity Index

As seen in Fig. 10, the observed and calculated THI values were closely correlated with each other, attaining maximum values after noon and then decreasing as the afternoon progresses. Discrepancies which exist between the simulated and measured indices were caused by both the dry-bulb and wet-bulb temperatures. The maximum disagreement between the calculated and observed THI values occurs at Patuxent River and Langley during the morning hours due to the underprediction of the wet- and dry-bulb temperatures. Generally the observed and calculated patterns are similar with most differences < 1°C.

c. Skin temperature

The $T_s$ values are very sensitive to changes in wind speed, mainly when speed is low. For example, a change of the wind speed from 2 to 1 m s$^{-1}$ causes a 1.9°C increase in $T_s$. It is worth noting that according to Fanger's comfort diagrams (pp. 45–47), which are derived as mentioned in Section 2b by introducing constraints on skin temperature

![Graph showing temperature variations over time at different locations.](image)

**Fig. 9.** Observed and predicted temperature during the day at 12 selected sites (cloudiness is indicated either by: (C-clear, S-scattered, B-broken, or in tenths of cloud coverage).
and sweat secretion, there is also a high dependence of comfort sensation on wind speed (as he illustrates for wind speeds < 1.5 m s⁻¹).

Differences between the model-calculated wind speeds and the observed ones, for example, can be due to the short averaging interval of the wind measurements, the accuracy of the reported wind speed (usually ±0.5 m s⁻¹), as well as the fact that the observation represents a point measurement, while the model is simulating grid-box-averaged wind speed (and therefore also average \( T_s \) values). Hence, considering the high sensitivity of \( T_s \) values to the wind speed, in addition to the reasons mentioned in the beginning of Section 8, one would expect the predicted and observed \( T_s \) values to often be somewhat different. Fig. 11 illustrates the observed and calculated \( T_s \) patterns. The large deviation at Salisbury at 1200 LST is due to a calm wind report at that time which illustrates the major dependence of \( T_s \) on wind speed. (An observed report of calm is usually due to wind speeds below the anemometer threshold, so we have used a minimum value of 0.5 m s⁻¹ in the observed \( T_s \) calculation.)

d. Error analysis

In order to quantify the skill of the model in the prediction of the thermal indices, a quantitative error analysis has been performed. The root-mean-square error (RMSE), standard deviation of observed values (\( \sigma_{\text{obs}} \)) and standard deviation of predicted values (\( \sigma_{\text{cal}} \)) are given in Table 1 for \( T \), \( \text{THI} \) and \( T_s \), along with the ratio RMSE/\( \sigma_{\text{obs}} \).

This analysis was performed using the four individual input values for each of the 12 stations yielding a total of 48 data points. As pointed out by Keyser and Anthes (1977), \( \sigma_{\text{obs}} \approx \sigma_{\text{cal}} \) and RMSE/\( \sigma_{\text{obs}} < 1 \) indicates that a model has demonstrated skill in simulating the individual variables. In the model calculations performed for 21 July 1978 each of the biometeorological parameters, \( T \), \( \text{THI} \) and \( T_s \), are simulated realistically with the temperature providing the greatest skill.

9. Discussion

The model results provide a quantitative estimate of the biometeorological heat load over the greater Chesapeake Bay area for a typical summer day. This detailed spatial description cannot be obtained by existing observational means. A verification of the simulation using available data indicated that the model has skill in predicting these discomfort indices. The results of this study include the following:

1) The temperature (\( T \)), Temperature-Humidity Index (THI) and skin temperature (\( T_s \)) fields show similar qualitative spatial and temporal distribu-
Fig. 11. Skin temperature ($T_s$) based on observed and predicted meteorological data, during the day at 12 selected sites.

tions; relatively low values at the coastal strips and to some extent in the estuaries of the Potomac and James Rivers. Relatively homogeneous fields were predicted in the interior areas.

2) Calm areas in the northeast Delmarva peninsula were forecast to persist for long hours of the day accompanied by severe heat load conditions. (This pattern was most pronounced in the $T_s$ fields.)

3) Transient small areas with severe heat conditions were calculated in the Norfolk vicinity. This situation, as well as that listed under 2), develops when the sea breeze and synoptic flow components are approximately equal in speed but opposite in direction. The result of this pattern is an area of exceptionally light winds.

4) The inland penetration of the sea breeze is also very limited when the synoptic wind is offshore, so that the moderating effect of the onshore marine air on the biometeorological properties is reduced to a narrow strip along the coasts. This amelioration of hot weather immediately along the Delmarva and southeastern Virginia coast is well known to tourists and helps explain the popularity.

<table>
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<th>RMSE</th>
<th>$\sigma_{obs}$</th>
<th>$\sigma_{cal}$</th>
<th>RMSE/$\sigma_{obs}$</th>
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<td>THI (°C)</td>
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of Ocean City, Maryland and Virginia Beach as coastal recreation areas during the summer.

The model, however, cannot resolve wavelengths with scales smaller than twice its grid interval. Therefore, in regions with a highly convoluted coastline, such as the Chesapeake Bay region, it will be necessary to increase the model resolution by decreasing the grid size in order to simulate the wind circulations along the individual bays and estuaries as well as over urban areas. At the same time, however, the entire Bay region must be represented in order to include the large-scale Bay and sea breezes, perhaps suggesting that a nested-grid approach is needed where only the immediate area of interest has the fine-scale resolution.

Finally, clouds and vegetation are being incorporated into the model (McCumber, 1980) and this may lead to an improvement in further simulations. Models such as presented here offer the best hope of obtaining this information at reasonable cost because a detailed network over this region would be exorbitantly expensive.

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REFERENCES


