# A Climatology of Surface Baroclinic Zones

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ABSTRACT

Analyses of surface potential temperature for one year over the contiguous United States, southern Canada, northern Mexico, and adjacent oceanic areas showed three regions of relatively high frequency of strong gradients. These were the Atlantic and Gulf coasts, the Pacific coast, and the eastern slopes of the North American Cordillera. The Atlantic and Gulf zone was most pronounced at 1200 UTC during winter. The Pacific coast zone was most frequent at 0000 UTC in summer. The zone on the eastern slopes was present at both times and during all seasons but was most pronounced at 1200 UTC in winter. The locations of the maxima and the pronounced diurnal changes provide circumstantial evidence that the horizontal variation of diabatic heating and cooling due to surface heat flux is a more important physical mechanism than confluence is in the creation of surface baroclinic zones. An example is shown of the importance of the East Coast zone for weather forecasting during the cold season. In January of 2000 a low pressure center propagated along the surface baroclinic zone, separating cold air over land from warm air over the water, rather than following the deep baroclinic zone in the troposphere. During the warm season, the ''backdoor cold front,'' separating cool air over the water from hot air over land, presents a challenging forecast problem. Many analyzed fronts are not associated with significant horizontal density gradients. Because of these and other surface analysis issues, short-range forecasting might benefit from routine analysis of the surface temperature field.

## 1. Introduction

A concept that originated in the initial exposition by Bjerknes (1920) is that a front represents a discontinuity of density between two air masses that meet at a prominent wind shift line but do not mix. In fact, Bjerknes notes, ''Every moving cyclone has two lines of convergence, which are greater and more conspicuous than the others . . .'' (emphasizing the wind shift), but he adds in the same sentence, ''. . . and are distinguished by characteristic thermal properties.'' Later he refers to ''[t]he discontinuous character of the change of temperature . . .'' during the passage of these lines. Further, the warm air is said to ascend along the surface separating it from the colder air, but no reason is given for the ascent. We are left to infer that the warmer air rises because it is less dense. This view is confirmed later when it is stated that, ''as a combined effect of this turning motion [around the cyclone center because of the barometric depression and the deflecting force of the earth's rotation] and the different specific weights, the cold current is screwed underneath the warm one, and the warm current screwed up above the cold one.'' There can be little doubt that the original intent was that there should be a discontinuity of temperature at a front, or at least a strong gradient. We concur with this definition of a front.

We are aware that other definitions of a front are often used. Examples are change in the origin of the air, strong gradient of surface humidity, the leading edge of cold or warm advection, and (at sea) change from air that is warmer than the water to air that is colder. Only a definition that relies on a density discontinuity or strong gradient, however, can explain the ascent of warm air over cold or the abruptness of the change sometimes observed.

Hence it came as a surprise that fronts as shown on routine surface map analyses often are not accompanied by a strong contrast of temperature (e.g., Sanders and Doswell 1995) and that strong contrasts are often not denoted as fronts (Sanders 1999a). It is therefore of interest to establish a climatological description (''climatology'') of surface baroclinic zones. In particular we will be concerned in this paper with the occurrence of zones of strong gradient of potential temperature. Potential temperature is chosen in an attempt to adjust for

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the effects of variable station elevation. The significance of this choice is discussed by Sanders (1999a). A strong zone is identified by a difference of 8°C (14.4°F) over a distance of no more than 110 km and a moderate zone is identified by the same difference over no more than 220 km, as proposed by Sanders (1999a).

## 2. Analyses of surface potential temperature

The establishment of such a climatology was made feasible by the development of automated analyses of surface potential temperature prepared at the University at Albany, State University of New York (SUNYA), and made available there and on the Internet (http://www.atmos.albany.edu and subsequent choices). These maps cover the contiguous United States, southern Canada, northern Mexico, the Gulf of Mexico, and the extreme western portion of the Atlantic Ocean and eastern portion of the Pacific. All available surface observations are used, as well as observations from buoys, Coastal-Marine Automated Network (C-MAN) stations, and ships of opportunity. Analysis of these maps is obtained by use of a Barnes-type GEMPAK (General Meteorological Package) procedure (Barnes 1964; Koch et al. 1983) on a grid of 0.5° of latitude and longitude. Analyses are made at 3-h intervals, and regions of moderate and strong gradients are determined from gridpoint values.

In these automated analyses, strong gradients are defined as at least 7°C (100 km)$^{-1}$ and moderate gradients are taken as one-half of this value. During the summer, these thresholds are reduced to 5° and 2.5°C (100 km)$^{-1}$, respectively. The cold-season values are close to the criteria employed in this paper, and the analyzed regions are accepted at face value for the most part, subject to some checking of differences between adjacent stations. During the warm season, the regions of strong gradient are determined manually, using the cold-season criteria. The analyses are modified along the Atlantic coast, where the automated procedure sometimes yields values much colder than those reported at the C-MAN stations along the coast and the buoys and ships near the shore. This discrepancy is attributable to the numerous observations inland in the colder air and the few in the warmer air offshore, because the analyzed gridpoint value is the weighted average of the observations within a specified distance from the point. No account is taken of the intercorrelation of observations in the algorithm. A similar problem did not arise on the West Coast, presumably because the number of buoys and C-MAN and routine stations is sufficient to yield a reliable analysis. The automated analysis is also completed or altered over Mexico, where the station density is often marginal or insufficient for the GEMPAK algorithm.

This climatology is based on maps for 0000 and 1200 UTC from August of 1999 through July of 2000. Of the possible total of 732 maps, 657 were collected, the shortfall being due to the author not having access to the Internet source or to technical problems at SUNYA. On each map,

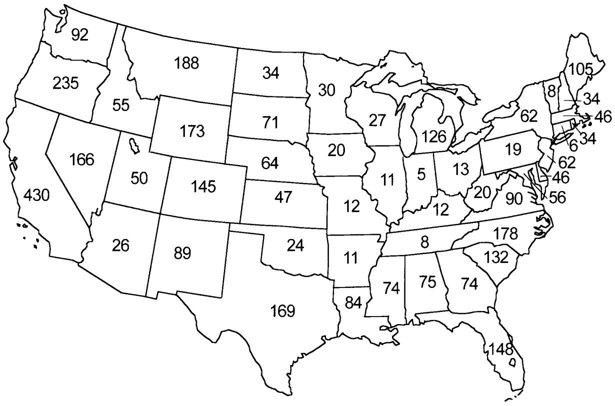


FIG. 1. For each state and adjacent water area, the number of maps containing a strong baroclinic zone for 0000 and 1200 UTC, Aug 1999–Jul 2000. Total number of maps is 657.

strong zones were identified and located according to the state or states that each zone covered. Zones lying immediately offshore were assigned to the adjacent state. Results for the entire year appear in Fig. 1.

## 3. Overall results

The two most prominent features in Fig. 1 are maxima along both the East and West Coasts. The former extends from North Carolina to Florida with an extension westward along the coast of the Gulf of Mexico. In the West, the maximum is in California with an extension northward to Washington. These major features strongly suggest the influence of difference in surface heat flux over the oceans, where the large thermal inertia keeps surface air temperatures from varying strongly, and over the adjacent land, where the temperatures can vary widely depending on season, time of day, and the meteorological circumstances.

A third region of maximum frequency of strong baroclinic zones, less well defined than the others, runs along the eastern slopes of the North American Cordillera from Montana through Wyoming and Colorado to New Mexico. The frequency in each of these states is larger than in the adjacent states to the east and west. These instances occur when cold air from Canada sweeps down the plains but does not penetrate to the high elevations to the west.

There is a general correlation between elevation and surface potential temperature at a given time, reflecting the stable stratification of the atmosphere on average. Sanders (1999a) has pointed out, however, that when and where the surface boundary layer is well mixed, the isentropic surfaces are vertical and thus the horizontal temperature gradient is identical to the gradient along the earth's surface.

Because the size of the states is far from uniform, an attempt at normalization was undertaken. This was done by dividing the raw value for each state by the ratio of that state's area to the area of the smallest state, Rhode

TABLE 1. Number of statedays by season and time.

| Season | 0000 UTC | 1200 UTC | Both times |
|---|---|---|---|
| Summer | 473 | 235 | 708 |
| Autumn | 391 | 580 | 971 |
| Winter | 440 | 794 | 1234 |
| Spring | 458 | 344 | 802 |
| All seasons | 1762 | 1953 | 3715 |

Island. The major features identified above survive the process, but the smaller states showed distinctly higher counts, reflecting state size rather than something in the atmosphere. No further attempt at normalization was tried.

It might be argued that the strong gradients in the surface boundary layer are shallow and thus of little dynamical significance. We note, however, that all three zones of strong contrast are associated with cyclonic developments. East Coast cyclogenesis is well known, frequent, and much studied. The lee slopes on the High Plains are likewise a region of frequent cyclogenesis, and perturbations on the West Coast baroclinic zone, although relatively rare, have received considerable recent attention (e.g., Mass and Steenburgh 2000; Skamarock et al. 1999; Jackson et al. 1999). These West Coast perturbations do not produce the important precipitation and gale winds of the East Coast cyclones, probably because of the summer lack of a mobile upper-level trough to interact with the surface disturbance, as found in winter in the East, and because of dryness of the warm air in the high western deserts rather than the moisture and low static stability of warm air in the Atlantic cyclogenetic situation. In all three cases, the cyclone tends to follow the isotherms of the surface boundary layer with warmer air to the right of the path, at least in the early stages of its life history.

## 4. Distributions by season and by time

To elucidate the effect of season and time on the frequency of strong zones, the year was divided into summer (August 1999 and June–July 2000), autumn (September–November 1999), winter (December 1999–February 2000), and spring (March–June 2000). The total number of statedays (defined as the occurrence of a strong zone within a particular state on a particular day, as in Fig. 1) for each season and for each of the two times studied is given in Table 1. It is seen that the number is somewhat larger during the cold season than during the warm season, reflecting the stronger meridional temperature gradient and synoptic activity from September through March. It is further seen that the number is larger at 0000 UTC than at 1200 UTC during spring and summer. The contrast between warm air over land and cold air over water evidently dominates the warm season, because 0000 UTC is not long after the time of diurnal maximum temperature, whereas 1200 UTC is near the time of daily minimum. During autumn and winter, the number is larger at 1200 UTC, reflecting the dominant effect of cold air over land relative to warmth over the water.

The detailed patterns for each time and season are shown in Figs. 2–5. We note the following features of interest.

1) The California maximum peaked at 0000 UTC during summer (Fig. 2a), when it was present almost every day. Even at 1200 UTC (Fig. 2b), it persisted on most days. In winter, the strong zone occurred in California at 0000 UTC (Fig. 4a) on more than one-half of the days in the sample. The northward extension into Oregon and Washington occurred at both times in all seasons, with decreasing number of days. There was evidently no separate maximum of strong zones in either of these states.
2) The maximum along the eastern slopes of the North American Cordillera was present at both times and in all seasons. Its intensity was measured by comparing the average number of days in Montana, Wyoming, Colorado, and New Mexico (on the eastern slopes) with the average in Idaho, Utah, and Arizona (to the west) and in North Dakota, South Dakota,

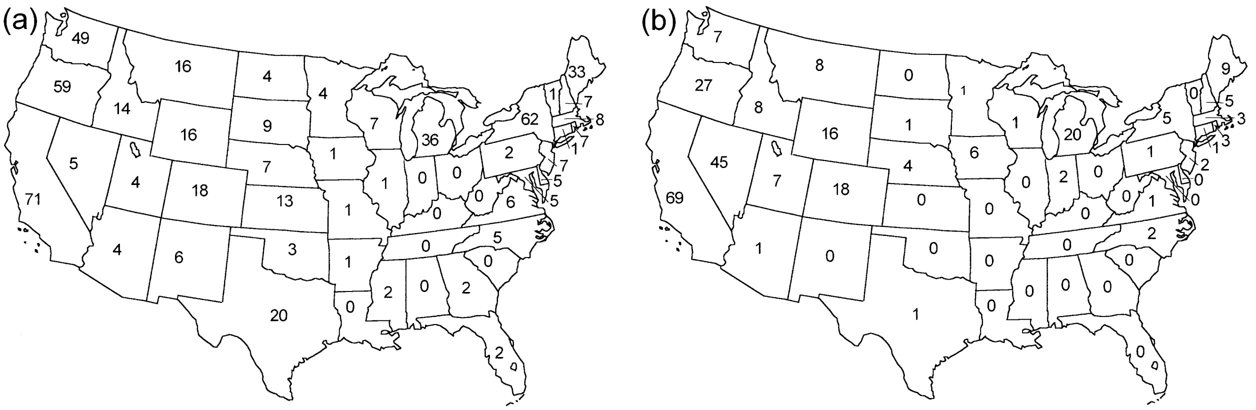


FIG. 2. Same as Fig. 1, but for (a) 0000 UTC (73 maps) and (b) 1200 UTC (72 maps) in summer.

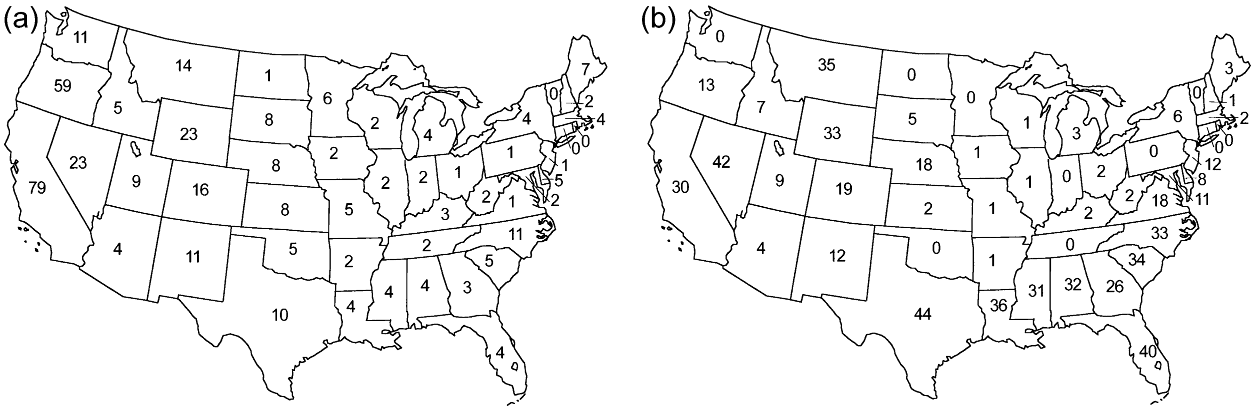


FIG. 3. Same as Fig. 1, but for (a) 0000 UTC (87 maps) and (b) 1200 UTC (88 maps) in autumn.

Nebraska, Kansas, and Oklahoma (to the east). The ratio of the two averages was greater than 1 at both times and during all seasons. The ratio was greater at 1200 UTC than at 0000 UTC, indicating that nocturnal cooling exacerbated the contrast. The maximum ratio was 5.6 at 1200 UTC in spring (Fig. 5b), and the minimum of 1.9 was at 0000 UTC in summer (Fig. 2a).

3) The maximum along the southeast Atlantic and Gulf coasts showed a pronounced seasonal variation, the total number of days in the coastal zone from North Carolina to Louisiana (summed over both times) varying from 13 in summer to 386 in winter. Except in summer, the count was larger at 1200 UTC than at 0000 UTC, indicating the influence of nocturnal cooling over land.
4) An isolated maximum in Michigan was seen at 0000 UTC in spring and summer (Figs. 2a and 5a) and is attributable to the persisting cold water in Lake Superior relative to hot days over land, especially in the Upper Peninsula.
5) A maximum in Maine at 0000 UTC in summer (Fig. 2a) and at 1200 UTC in winter (Fig. 4b) reflects the slow variation of water temperature in the Gulf of Maine relative to cold air over land in winter and warm air over land in summer. This is a singular example of a seasonal reversal of the sense of a frequent strong baroclinic zone.

Although the sample size is small, physical considerations indicate that the major features would likely be replicated in a sample over a number of years or in a method in which frequencies were counted in areas of equal size rather than in states.

## 5. A winter example

The importance of the East Coast baroclinic zone is illustrated by the behavior of a cyclone that was responsible for a major storm on 24–26 January 2000, producing hail and damaging winds from thunderstorms along the Gulf coast, strong winds and flooding along the Atlantic coast, and 1 ft or more of snow at inland locations from South Carolina to Vermont. The storm followed a track well to the west of model forecasts even up to the last moment. Some surface analyses for this storm, prepared by the National Weather Service (NWS), appear in Fig. 6. On 24 January at 0000 UTC

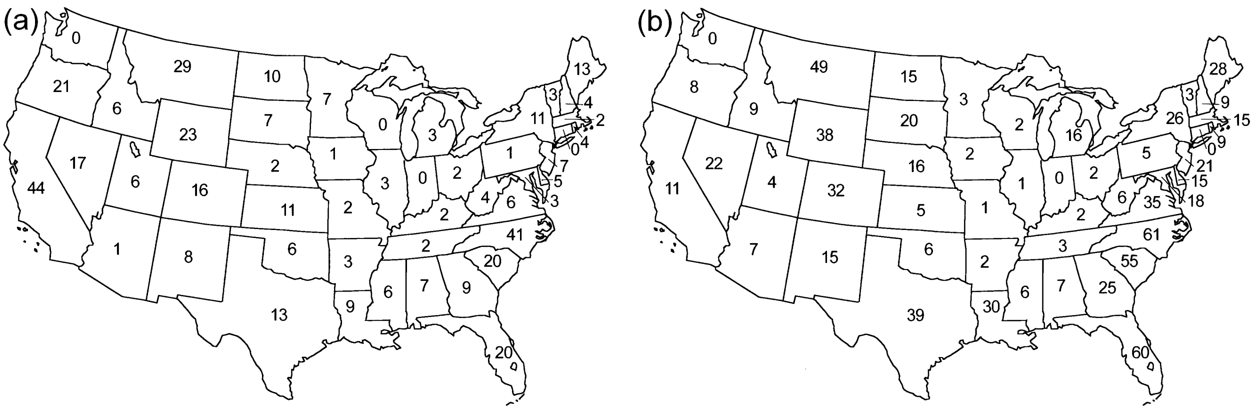


FIG. 4. Same as Fig. 1, but for (a) 0000 UTC (84 maps) and (b) 1200 UTC (88 maps) in winter.

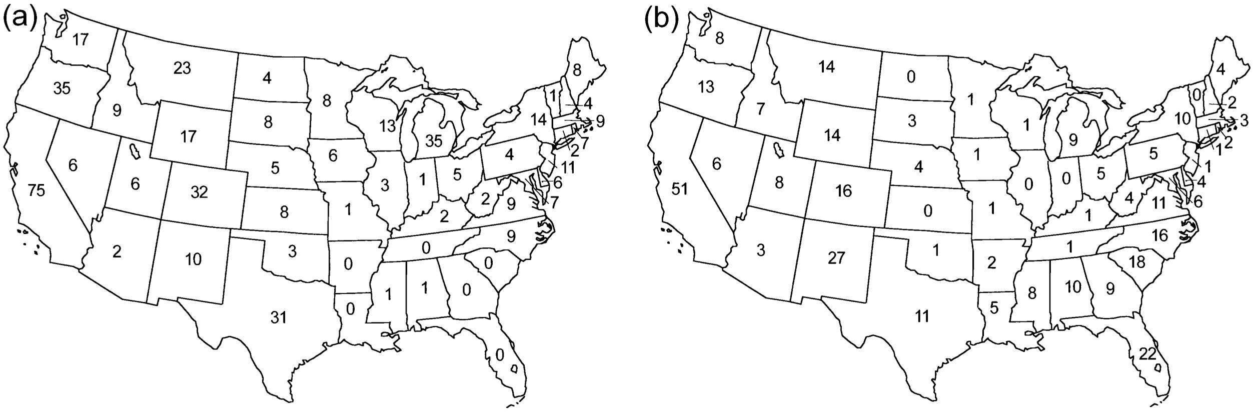


FIG. 5. Same as Fig. 1, but for (a) 0000 UTC (85 maps) and (b) 1200 UTC (83 maps) in spring.

(Fig. 6a), a moderate-to-strong baroclinic zone extended from the Gulf and southeast Atlantic coasts northeastward south of New England. The analysis showed a frontal system near the warm edge of this zone, in good accord with the Norwegian model. A weakly defined low pressure center was seen along this front east of the Mid-Atlantic states, and another was found along the coast of Mississippi. The baroclinic zone was especially intense adjacent to the first low and in advance of the second. The 500-mb flow over the second low suggested a path well offshore, and, indeed, forecasts at this time indicated precipitation mainly over the water (NCDC 2000).

Twelve hours later (Fig. 6b), the first low had moved to the northeast and the baroclinic zone had relaxed behind it, presumably because of mixing in the absence of the convergence that had moved with the low, as in the case discussed by Sanders (1999b). A separate center of intensity along the coast of Maine moved slightly southward as the coastal wind shifted to northerly. It was not directly associated with the first low. The second low had propagated along the isotherms of the baroclinic zone to a location near the eastern edge of the Florida Panhandle. The baroclinic zone itself remained largely stationary and had perhaps strengthened slightly in advance of the low while beginning to move southward in its wake.

By 0000 UTC 25 January (Fig. 6c), the first low had moved northeast beyond the region of interest and the intense part of the baroclinic zone was associated entirely with the developing second low, now off the coast of South Carolina. In its wake, the baroclinic zone was beginning to weaken as it moved east of the coast of Florida. Again, the convergence was limited to the region in advance of this low. Last, (Fig. 6d), the zone was intense over and ahead of the now-powerful low east of the North Carolina–Virginia border. To its south, the analyzed cold front had moved well out into the Atlantic Ocean in a region of weak temperature gradient while the main contrast was observed close to shore and along the edge of the Gulf Stream. The analyzed front still displayed a significant wind shift and is probably best regarded as a nonfrontal baroclinic trough (Sanders 1999a).

Two aspects of the behavior of this situation warrant particular attention. First, the path of the low center tended to follow the initial orientation of the coastal baroclinic zone in the surface boundary layer rather than the initial orientation of the 500-mb flow and of the thermal wind over a deep layer of the atmosphere. Although the flow over the deep layer evolved rapidly to a meridional orientation over the low, as shown in Figs. 6a,d, the field in the boundary layer moved little and appeared to determine the path of the storm. The front ahead of the low behaved as a ''steering line,'' to use the term introduced by Bjerknes (1920).

Second, the surface zone strengthened ahead of the low pressure center and weakened in its wake, illustrating the short life of a frontal system and the rapid changes within it. This weakening is associated with the propagation of the low-level convergent wind shift ahead of the baroclinic zone. The reason for this tendency of the wind shift to ''outrun'' the temperature gradient is that convergence ahead of and divergence behind the cyclonic trough contribute to its motion but not to the motion of the temperature gradient whenever there is a component of thermal wind across the trough from cold to warm. In this case, horizontal variation of surface heat flux from over land to over the Gulf Stream is an important additional factor.

## 6. Baroclinic zones and analyzed fronts

The above example raises an important question that must be addressed, namely, what the degree of correspondence between baroclinic zones, as presented here, and operational frontal analyses is. First, we refer to a climatology of fronts presented by Morgan et al. (1975), who examined the Daily Weather Maps produced by the NWS for the period from 1 January 1961 to 31 December 1970, obtaining frequencies for areas of 3600 (n mi)$^2$ (4900 mi$^2$). The maps are for a time of approxi-

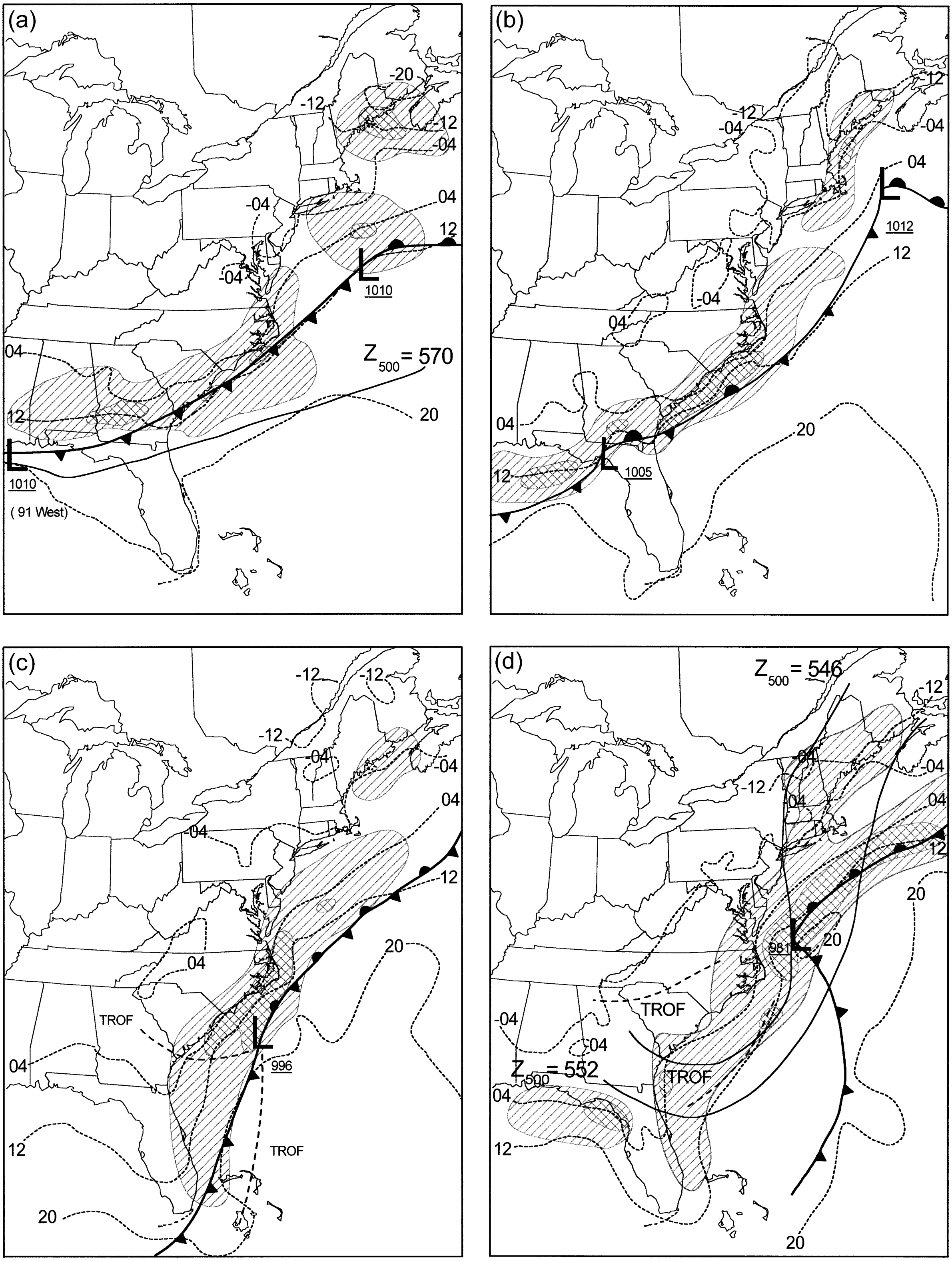


FIG. 6. Isotherms of surface potential temperature, at intervals of 8°C (dashed lines) at (a) 0000 and (b) 1200 UTC 24 Jan and at (c) 0000 and (d) 1200 UTC 25 Jan 2000. The heavy solid line is a selected 500-mb geopotential contour, labeled in dekameters. The NWS analysis of fronts and low pressure centers is shown in conventional notation. Areas of moderate and strong gradient of potential temperature gradient are shown by single and double hatching, respectively.

TABLE 2. Comparison of seasonal frequencies of strong baroclinic zones (SBZ) at 1200 UTC and analyzed fronts for selected states.

| | Maine | | North Carolina | | California | | Montana | | Indiana | |
|---|---|---|---|---|---|---|---|---|---|---|
| | SBZ | Fronts | SBZ | Fronts | SBZ | Fronts | SBZ | Fronts | SBZ | Fronts |
| Winter | 28 | 18 | 61 | 36 | 11 | 33 | 49 | 56 | 0 | 35 |
| Spring | 4 | 24 | 16 | 40 | 51 | 25 | 14 | 50 | 0 | 40 |
| Summer | 9 | 42 | 2 | 44 | 69 | 18 | 8 | 37 | 2 | 48 |
| Autumn | 3 | 34 | 33 | 33 | 30 | 29 | 35 | 47 | 0 | 35 |
| Annual | 44 | 118 | 112 | 153 | 161 | 105 | 106 | 190 | 2 | 158 |
| Area | 6.3 | | 9.9 | | 31.8 | | 29.8 | | 7.3 | |

mately 1200 UTC. Because of differences in the measurement areas, because of the single time represented, and because of stratification by frontal type in the frontal climatology, a direct comparison with our results is difficult.

A comparison was made, however, for selected states. The frequencies of intense baroclinic zones were taken directly from Figs. 2b, 3b, 4b, and 5b. Morgan et al. (1975) present 10-yr frequency maps for each month, separately for cold fronts, warm fronts, stationary fronts, and occluded fronts. For each selected state, the mean frequency was estimated for each 3-month season, was summed over the four frontal types, and was divided by 10 to give an estimated seasonal frequency. Results are shown in Table 2. Although some larger states comprise many areas of the size used by Morgan et al. (1975), it is difficult to know how to adjust for state size, because a single front may extend over several adjacent areas. Some baroclinic zones, moreover, may extend over a number of states. These difficulties aside, a few points stand out.

1) There were considerably more analyzed fronts than intense baroclinic zones. Many analyzed fronts evidently are associated with only a moderate baroclinic zone or with no significant contrast at all. An extreme example is the state of Indiana, for which only two intense zones were found at 1200 UTC in the year studied but for which the estimated area-mean frequency of fronts was 158 per year per 3600 square nautical miles.
2) The annual cycles of baroclinic zones and fronts are different. In North Carolina, for example, a maximum of baroclinic zones occurs in winter, when the contrast between the warm water offshore and cold air inland is a maximum, but the frequency of analyzed fronts is nearly a minimum. In summer, the opposite situation is found: few baroclinic zones but many fronts. In California, where an intense zone is observed on most days, even at 1200 UTC, when surface temperatures over land are near their coolest, there are almost no analyzed fronts in the southern part of the state. This zone evidently is not regarded as frontal. Indeed, at this time the intense baroclinic zone may not represent a horizontal density contrast, owing to strong stratification. In winter there conversely is a minimum of baroclinic zones but a maximum of analyzed fronts. The Norwegian model evidently is followed, and the lack of surface contrast is attributed to the homogenizing effect of the ocean in airflow with a long trajectory across the Pacific Ocean.
3) In Maine, the seasonal cycle of baroclinic zones and analyzed fronts is reversed, with a maximum of zones in winter and of fronts in summer. In Montana, the seasonal cycles coincide, but the amplitude of the cycle for zones is much greater than that for fronts.

A more direct comparison of fronts and baroclinic zones was carried out during the winter period from December of 1999 through February of 2000. On the NWS analyses for 0000 and 1200 UTC, frontal segments and troughs were identified. The edge of a segment was taken to be either where the analysis indicated a change of frontal type or where a frontal system passed through the center of a cyclone. Segments were stratified as cold fronts, warm fronts, or stationary fronts, following the NWS identification. Occlusions were ignored. The analysis was then compared with the unmodified surface potential temperature analysis for the same time. Each segment was identified as being associated with a baroclinic zone, of either moderate or strong intensity, or not. The association was regarded as existing if such a zone was present within 200 km of the segment, along a substantial portion of its length, on either the warm or cold side of the front. The orientations of the frontal segment and the baroclinic zone differed by no more than 45°. The same association was tested for troughs, and note was also taken of such baroclinic zones that were not identified with an analyzed frontal segment. Results are shown in Table 3.

Overall there were 677 frontal segments, of which 63% were associated with baroclinic zones. Of 302 troughs, 39% were associated with baroclinic zones. Further, there were 668 baroclinic zones not associated with frontal segments. Thus, there is a degree of association of fronts with baroclinic zones, as prescribed by Bjerknes. There are some differences when the data for 0000 UTC are compared with those for 1200 UTC. At the former time, 66% of 334 frontal segments were associated with zones; at the latter time, 60% of 343 segments were so associated.

There were important differences among the three

TABLE 3. Association of analyzed fronts and troughs with moderate or strong surface baroclinic zones, Dec 1999–Feb 2000. Numbers in parentheses are percentages of the number of fronts or troughs.

| | 0000 UTC | | | 1200 UTC | | |
|---|---|---|---|---|---|---|
| Type | No. | Associated | Not associated | No. | Associated | Not associated |
| All fronts | 334 | 221 (66) | 113 (34) | 343 | 206 (60) | 137 (40) |
| Warm fronts | 62 | 49 (79) | 13 (21) | 70 | 48 (69) | 22 (31) |
| Stationary fronts | 114 | 98 (81) | 22 (19) | 121 | 89 (74) | 32 (26) |
| Cold fronts | 158 | 80 (51) | 78 (49) | 152 | 69 (45) | 83 (55) |
| Troughs | 165 | 57 (35) | 108 (65) | 137 | 62 (45) | 75 (55) |
| | | Frontal | Nonfrontal | | Frontal | Nonfrontal |
| All baroclinic zones | 889 | 221 (25) | 668 (75) | 603 | 206 (34) | 397 (66) |

frontal types. In support of the Norwegian cyclone model, 73% of warm fronts and 77% of stationary fronts were associated with baroclinic zones. A prominent synoptic example was recently presented by Sanders (2000). These results support the original steering-line concept proposed by Bjerknes (1920), because cyclones tend to propagate along the wind shift line extending eastward from the center. This feature is usually denoted as a warm front or a stationary front.

Cold fronts were a different matter. Only about one-half of the 310 cold fronts were associated with a baroclinic zone: 51% at 0000 UTC and 45% at 1200 UTC. It appears that cold fronts are identified in the analysis more on the basis of wind shift than of temperature contrast. Examples of alleged cold fronts accompanied by an abrupt temperature rise at night were given by Sanders and Kessler (1999), and Sanders (1999b) gave a detailed analysis of a case in which the wind shift propagated eastward away from the temperature contrast. Such a separation appears from quasigeostrophic theory to be necessary when there is a component of thermal wind normal to the front from colder air toward warmer air. Such a structure is the usual case. Note that, if a more stringent criterion were applied (such as requiring that the baroclinic zone lie on the cold side of the front), the percentage of analyzed frontal segments accompanied by baroclinic zones would be reduced.

These comparisons do not imply an indictment of map analysis as practiced by NWS. Similar characteristics are seen in limited experience with analyses made by other national services and by research and other private organizations. Implied criticisms refer to the commonly accepted methods.

## 7. Concluding summary

A preliminary climatology of strong surface baroclinic zones was obtained from one year of twice-daily automated surface analyses, at 0000 and 1200 UTC, over the contiguous United States, northern Mexico, southern Canada, and adjacent portions of the western Atlantic and eastern Pacific Oceans, as well as the Gulf of Mexico. A strong zone was defined as one in which an 8°C contrast of surface potential temperature occurred over a distance of no more than 110 km. A zone was assigned to a state, or states, in which a zone lay or in which it occurred over the immediately adjacent coastal waters.

Over all seasons and both times there were three zones of maximum frequency of occurrence: one along each coast and a third on the eastern slopes of the North American Cordillera. The highest frequency was in the zone along the West Coast, in California in summer at 0000 UTC (late afternoon local time). On the southeast Atlantic and Gulf coasts, there was a high frequency in winter at 1200 UTC (near dawn local time), but there were few cases during the warm season at 0000 UTC. The maximum on the eastern slopes of the western mountains occurred at both times and in all seasons but was somewhat more pronounced at 1200 UTC and in winter. Circumstantial evidence points to the dominance of the horizontal variation of surface heat flux in creation of these zones. Examples are shown of the importance of the East Coast baroclinic zone for forecasting in both the cold and warm seasons. A comparison with NWS analyses indicates that many fronts are not associated with a moderate or intense baroclinic zone. The discrepancy is particularly great for analyzed cold fronts.

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