EXTRATROPICAL SYNOPTIC-SCALE PROCESSES AND SEVERE CONVECTION

Chapter 2 in

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I. Introduction:

Within this chapter, we intend to give a broad perspective of the *interaction* between severe convection and extratropical synoptic-scale processes. A traditional view of this interaction is that the synoptic-scale processes simply provide a setting in which severe convection develops (see, e.g., Newton 1963; Barnes and Newton 1983; Johns and Doswell 1992). This view could be interpreted as implying that convection has little or no direct impact on synoptic scales. However, there have been many recent developments in *mesoscale* meteorology as it relates to severe convection (as described in Ch. 3 of this volume), wherein upscale effects of convection are seen most clearly. Mesoscale processes often act as a sort of "intermediary" between convective and synoptic scales. We take the view that, in spite of the intermediation by mesoscale processes, it still is possible to take a synoptic-scale view of the impacts of deep, moist convection, especially in its most severe manifestations.

The subject of scale separation is always a thorny one. Orlanski's (1975) scale divisions are essentially arbitrary, based on powers of ten in space and time. There are also dynamically motivated ways to divide scales [Emanuel 1986; Doswell 1987], but there is no universally accepted way to separate scales of motion. For the purposes of this review, we are concerned with the processes associated with extratropical weather systems in midlatitudes; tropical synoptic-scale processes are considered in Ch. 7 of this volume.

Quasigeostrophic (QG) theory is arguably the simplest statement of what it means to be "synoptic-scale" (Doswell 1987), at least outside of the Tropics. In section II, we provide brief overviews of QG principles, potential vorticity thinking and basic jet streak-related processes. Section III presents a discussion of planetary boundary layer processes, focusing on how these relate both to synoptic scales and to severe convection. Section IV provides some basic climatological distributions of convection, both in space and in time. These observed climatological distributions provide important clues as to the interaction between synoptic-scale processes and convection. The climatology of *severe* forms of deep, moist convection is the topic of Section V. In section VI, brief overviews of a number of cases are presented, in part to illustrate the principles developed, but also to show the variety of synoptic-scale structures in which severe convection can develop. Section VII presents some perspectives on the synoptic contributions to severe convection, and section VIII provides a discussion of the reverse feedbacks of convection to the synoptic scale. Finally, section IX provides some discussion and conclusions.

II. Brief overviews

Deep, moist convection (DMC)[1] is associated with a triad of necessary and sufficient ingredients: moisture, low static stability, and ascent of parcels to their level of free convection (LFC) by some lifting mechanism (Doswell 1987).[2] Synoptic-scale processes, notably the extratropical cyclone (ETC), play a reasonably well-understood role in moistening and destabilization, but the relatively weak vertical motions (on the order of a few cm s^{-1}) of synoptic-scale systems usually are too slow to lift potentially buoyant parcels to their LFCs in less than about a day. On the other hand, ETCs provide an environment that favors processes operating on smaller scales, such as drylines (see Schaefer 1986; Ziegler and Rasmussen 1998) and fronts, and vertical motions substantially larger than those of the synoptic scale can be created by those subsynoptic processes. Thus, there is a strong association between the development of DMC and ETCs, even though DMC is not confined exclusively to the ETC environment. Severe convection also exhibits this association, perhaps even more strongly than ordinary convection. To understand this connection, we begin with consideration of the simplest model of synoptic-scale processes in midlatitudes, quasigeostropic theory. Then, we move to consider potential vorticity, a modern perspective on synoptic-scale processes. Jet streaks are reviewed briefly in this same context.

1. QG principles

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[1] As in Ch. 1 of this volume, we use the term "deep moist convection" rather than "thunderstorm" since not all severe forms of deep, moist convection produce lightning and, hence, thunder.

[2] We assume that the presence of the LFC implies that the convection will, indeed, be “deep” although rare exceptions to this might be found.

Quasigeostrophic (QG) theory is not a concept that springs from a single, brilliant exposition; instead, it is a child of many parents. This is manifest in the excellent historical treatments of QG theory's development by Phillips (1990) or Bosart (1999).[3] Many respected scientists in the history of modern meteorology have made contributions to QG theory. The quasi-geostrophic system is contained in the two equations (following Holton 1992, p. 158 ff.)

$$\left[\nabla^2 + \frac{\partial}{\partial p}\left(\frac{f_o^2}{\sigma}\frac{\partial}{\partial p}\right)\right]\chi = -f_o\mathbf{V}_g \cdot \nabla\left(\frac{1}{f_o}\nabla^2\Phi + f\right) - \frac{\partial}{\partial p}\left[-\frac{f_o^2}{\sigma}\mathbf{V}_g \cdot \nabla\left(-\frac{\partial\Phi}{\partial p}\right)\right], \qquad (1)$$

and

$$\left[\nabla^2 + \frac{f_o^2}{\sigma}\frac{\partial^2}{\partial p^2}\right]\omega = \frac{f_o}{\sigma}\frac{\partial}{\partial p}\left[\mathbf{V}_g \cdot \nabla\left(\frac{1}{f_o}\nabla^2\Phi + f\right)\right] + \frac{1}{\sigma}\nabla^2\left[\mathbf{V}_g \cdot \nabla\left(-\frac{\partial\Phi}{\partial p}\right)\right], \qquad (2)$$

where ∇ is the horizontal gradient operator on a p-surface, Φ is geopotential height, the height tendency (χ) is defined by $\chi \equiv \partial\Phi / \partial t$, $\omega \equiv dp/dt$, the static stability (σ) is defined by $\sigma \equiv -(\alpha_o/\theta_o)(d\theta_o/dp)$. Static stability and the basic-state variables α_o (specific volume, or inverse density) and θ_o (potential temperature) are assumed to be functions of p alone, and f_o denotes a constant reference value of the Coriolis parameter. Equations (1) and (2) exhibit a near-symmetry that is apparent in the physical discussions presented by Holton (1992; p. 177 ff.) and Bluestein (1992; Ch. 5). Together, (1) and (2) constitute the QG forecasting system. It is possible to use this system, along with a mass continuity equation, to make forecasts; this system was used operationally at the National Meteorological Center (now the National Centers for Environmental Prediction, or NCEP) in the 1960s.

However, the real value in the QG system today is *not* in prediction, but in qualitative understanding of midlatitude, synoptic-scale processes (Durran and Snellman 1987). Equations (1) and (2) are derived from the primitive equations by making a number of assumptions about the flow they describe: namely, the flow is adiabatic and hydrostatic and the ageostrophic part of the flow

[3] Phillips attributes the first use of the term "quasi-geostrophic" to Sutcliffe (1939). However, Sutcliffe (1938) includes the following passage: "It is suggested that the term 'quasi-geostrophic' would be a better description in that the motion is geostrophic only to a first approximation."

makes no contribution to advection, etc. Textbook discussions (e.g., Holton 1992, p 166 ff.; Bluestein 1992, Ch. 5) point out that vertical motion in the QG system is simply a theoretical response to the disruptions of geostrophic and thermal wind balance caused by thermal and differential vorticity advection, with the QG response acting to restore hydrostatic and geostrophic balance. Disturbances in the height field (that are reflected in the relative vorticity) move by vorticity advection, weakening or intensifying as a consequence of differential thermal advection.

Generally speaking, QG theory predicts rising motion ahead of cyclonic disturbances and descending motion behind. In QG theory, however, □ is assumed to be, at most, a function of pressure, whereas in reality, □ varies in space and time. Rising motion favors a decrease in stability below the level of peak ascent, usually somewhere in mid-troposphere, whereas sinking motion increases the stability below the level of maximum descent.[4] Therefore, ETCs should exhibit some asymmetry in their vertical motion patterns beyond that derived from QG theory: ascent should be more intense than descent. Further, Emanuel et al. (1987) have shown (in a non-QG context) that if ascent is saturated and descent is unsaturated, the difference between moist ascent and dry descent can be parameterized by using a dry static stability for descent and a weaker, moist static stability for ascent. Doing so results in ascent being localized and relatively intense, whereas descent is weaker and more widespread (Whitaker and Barcilon 1992; Fantini 1995). These concepts generally are consistent with the observed behavior in mid-latitude cyclones (e.g., Whitaker et al. 1988) which certainly include non-QG processes; upward motion is typically stronger than downward motion and more localized (at times, ascent is concentrated in narrow bands associated with fronts, of course).

ETC development, (e.g., as described in Palmén and Newton 1969; Ch. 11; Uccellini 1990; Bosart 1999) is strongly dependent on the vertical motions. Vertical motion, in turn, depends on □ (even in the QG system considered above), so the environmental static stability is an important factor in cyclogenesis and its associated frontogenesis (Roebber 1993). Moreover, convection acts

[4] This also can be inferred from a static stability tendency equation; e.g., Panfosky (1964; p. 105 ff.).

to stabilize the troposphere, in general, so convection can modify the "environment" seen by cyclones, as well as the other way around. ETCs constantly are advecting and changing the static stability, as well, representing an important *nonlinear* interaction. Besides their impacts on (and modification by) the static stability field, ETCs are also quite important in transporting heat and moisture (Lorenz 1965; Ch. IV). Thus, the modulation of convection through modification of its environment by ETCs is quite substantial, and the interaction with DMC is complex and likely hard to forecast accurately.

Most (but not all) DMC involves parcels originating from near the surface. Although the observed 3-dimensional distributions of heat and moisture are complex in detail, it is generally the case that both heat and moisture tend to increase equatorward and decrease with height. Poleward flows bringing what is generally warm air poleward often include low-level moisture as well.[5] Thus, strongly meridional low-level flows, in general, transport moisture for convection and modify the global sensible and latent heat distributions significantly.

The environmental vertical wind shear structure has been shown to be a very important factor in determining the severity of any resulting convection (Newton 1963; Klemp and Wilhelmson 1978; Weisman and Klemp 1982, 1984; Brooks et al. 1994a). The *geostrophic* vertical wind shear is associated with thermal advection [an important factor in both Equations (1) and (2)]. In the Northern Hemisphere, warm advection favors a veering wind profile (at least to the extent that the vertical wind shear is dominated by the thermal wind contribution), as well as being associated with synoptic scale ascent. It has been shown (e.g., Browning 1964; Rotunno 1993) that strong vertical wind shear (of order 10^{-3} s^{-1} and larger) is a major factor in promoting the supercellular forms of convection that are the most prolific producers of severe weather. Some

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[5] Obviously, there are geographical circumstances where this may not be the case. Notable exceptions include Europe and Australia, where large, arid subtropical and tropical land masses are present equatorward of mid-latitudes. Meridional flows in such places typically transport dry air poleward; moisture is brought into midlatitude regions by predominantly easterly or westerly low-level flows that have a long fetch over warm, open waters. Such flows are also strongly modulated by ETCs, of course.

shear profiles also can favor the organization of convection into more intense and persistent forms (Rotunno and Klemp 1982; Rotunno 1993).

On the other hand, strong vertical wind shear has been viewed as an *inhibiting* factor for convection, in the sense that it tends to reduce the intensity of updrafts (Asai 1970, Brooks and Wilhelmson 1993). That strong vertical wind shear can interact with convection to enhance it in some circumstances was recognized in Newton's (1963) review paper, but this concept has been extensively refined by means of numerical cloud model simulations (Weisman and Klemp 1982; 1984). Not only does vertical wind shear promote new convective cell development via the interaction between existing updrafts and the shear (Rotunno and Klemp 1982), but it also affects the local distribution of hydrometeors produced by convection (Brooks et al. 1994b). It has been shown that the interaction between wind shear and convection can result in a contribution to the updraft as large as that of buoyancy (Weisman and Klemp 1984).

For synoptically-evident,[6] major outbreaks of tornadic severe weather (Doswell et al. 1993), favorable wind shear is widespread. In fact, baroclinic instability (linked to the development of ETCs) is closely related to the vertical wind shear through the thermal wind relationship. If this synoptic-scale vertical wind shear occurs in combination with potential instability, the stage is set for severe convection. Unfortunately, vertical wind shear-derived parameters can vary substantially within the synoptic and mesoscale environments (Davies-Jones 1993). Although the synoptic-scale environment can be characterized by some average vertical wind shear parameter, this is virtually certain not be representative of a particular convective storm's local environment. The mesoscale structure of the wind field can be understood in a QG context only to a limited extent; that is, only on the *largest* scales that might be considered to be "mesoscale".

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[6] The term "synoptically-evident" was used in Doswell et al. (1993) to imply a synoptic pattern associated with tornado outbreaks. These involve strong synoptic-scale advections and the presence of substantial CAPE in the presence of strong flow and vertical wind shear. It is by no means "evident" in advance which days are going to produce tornado outbreaks, but it is often possible to say retrospectively that there was a strong synoptic-scale signal suggesting the possibility of such an outbreak.

2. Potential vorticity (PV) perspectives

The seminal paper describing "potential vorticity thinking" is that by Hoskins et al. (1985). Detailed descriptions of potential vorticity (PV) can be found in textbooks (e.g., Holton 1992; Bluestein 1992). The intensification of cyclones is associated with cyclonic PV maxima that generally have ascent (descent) on their downshear (upshear) sides, even in weakly baroclinic environments, as discussed by Bosart and Lackmann (1995). Raymond and Jiang (1990), Raymond (1992) and Montgomery and Farrell (1993) have showed this theoretically, as well as observationally (e.g., Davis and Emanuel 1991). By being conserved for adiabatic, inviscid flow, an immediate benefit of PV diagnostics is that when PV is *not* conserved following the flow, it indicates important diabatic or viscous processes, with deep, moist convection often being an important contributor (Dickinson et al. 1997).

Standard texts (e.g., Holton 1992; Bluestein 1992) have shown that the QG form of PV (QGPV, or q) is a linearized form of the full Ertel PV for a frictionless, adiabatic flow, and is defined by

$$q \equiv \frac{1}{f_o}\nabla^2\Phi + f + \frac{\partial}{\partial p}\left(\frac{f_o}{\sigma}\frac{\partial\Phi}{\partial p}\right). \tag{3}$$

This quantity is conserved following the geostrophic flow, such that

$$\frac{D_g q}{Dt} \equiv \frac{\partial q}{\partial t} + \mathbf{V}_g \cdot \nabla q = 0\,. \tag{4}$$

Bluestein (1992, p. 372) has observed that (4) is equivalent to (1). QGPV is a good approximation to the “true” PV.[7] As can be seen easily in (3), q is simply the sum of the geostrophic absolute vorticity and a static stability term.

In its QG form, the PV perspective does not contain any physics beyond traditional QG approaches, but thinking in PV terms can be useful, as discussed in Hoskins et al. (1985), Davis

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[7] For some minor technical reasons, however, the QG form of PV is not a "proper" vorticity, as mentioned in Hoskins et al. (1985), Davies-Jones (1991) and Hakim et al. (1996).

and Emanuel (1991), or Morgan and Neilsen-Gammon (1998), among others. Mobile upper atmospheric troughs can be interpreted as PV anomalies that have their own induced flows extending vertically above and below the level(s) containing the PV anomaly. The *induced* flows act to redistribute PV, creating new anomalies by the advections they develop. Many aspects of midlatitude weather systems can be seen as a consequence of interactions between two (or more) PV anomalies, laterally and/or vertically. The induced flows can be derived by using the *invertibility principle* associated with PV thinking; see Charney (1962) or Raymond (1992), as well as other reference in this section on this topic.

Traveling upper-level systems also can be diagnosed using the concept of the dynamic tropopause (hereafter, DT), described in Hoskins et al. (1985) or Morgan and Nielsen-Gammon (1998). The DT is defined as a surface of constant PV (typically 1-2 x 10^{-6} K m^2 kg^{-1} s^{-1}, where the factor of 10^{-6} K m^2 kg^{-1} s^{-1} is often referred to as a “PV Unit” or “PVU”). On the DT, troughs are depicted as regions of high pressure (a low tropopause) and relatively low values of potential temperature. Jets and jet streaks are manifested as ribbons of closely packed potential temperature and pressure contours, indicative of a sloping DT (see Morgan and Neilsen-Gammon 1998). Moving with a positive PV anomaly, then, air parcels from the lower part of the troposphere are being overtaken by the PV anomaly and ascending along sloping isentropic surfaces.

However, when air parcels participate in DMC, they are no longer subject to the balance constraints of large-scale processes. Rather than rising along gently-sloping surfaces, buoyant parcels have a large vertical component of motion. This can have an important impact on the structures aloft ahead of a PV anomaly, since parcels ascending in DMC are carried rapidly to high levels, disturbing the dynamic balances near the tropopause, at least locally. That is, they create regions of high gradients of potential temperature, implying the creation of unbalanced flow, not unlike jet streaks. Diabatic effects will be seen as non-conservation of PV, of course, as noted in Whitaker et al. (1988) and illustrated in Fig. 1. Above an elevated maximum of diabatic heating (of which DMC is clearly an example), PV decreases and anticyclonic outflow increases, which is quite evident in the figure.[8] Beneath that diabatic heating maximum, PV and cyclonic inflow increase. The result (at least in the upper levels) is analogous to the structures created by Mesoscale Convective Complexes (MCCs - Maddox 1980; Fritsch and Maddox 1981), but on the synoptic scale.

The 13-14 March 1993 "Superstorm" over eastern North America provided an excellent example of the importance of diabatic processes associated with deep convection on explosive cyclogenesis. Dickinson et al. (1997; their Fig. 5) show that at the time of incipient cyclogenesis over the northwestern Gulf of Mexico surface θ_e values exceeded 340 K along the lower Texas coast and adjacent northwest Gulf of Mexico. This high θ_e air at the surface was overridden by air on the DT with a potential temperature < 330 K, indicative of exceptionally deep potential instability in the precyclogenetic environment over the northwestern Gulf of Mexico (Dickinson et al. 1997; their Fig. 3). The net result was the rapid onset of widespread deep convection in the vicinity of the cyclone, as seen in the satellite imagery and cloud-to-ground lightning distribution (Dickinson et al. 1997, their Figs. 6-8). It is possible that this convection was a factor in the ensuing rapid cyclogenesis (see section VII, below).

Bosart et al. (1999; their Fig. 2) showed that derechos (Johns and Hirt 1987) could be associated with long-lived, upper-tropospheric mesoscale disturbances moving eastward in a relatively strong flow aloft on the poleward flank of an intense continental anticyclone. The existence of these disturbances aloft during 13-15 July 1995 was revealed by an analysis of pressure and potential temperature on the DT. The disturbances aloft could be tracked for several days on the basis of an area of higher pressure and lower potential temperature on the DT. Deep convection began where ascent associated with these disturbances was superimposed on very high θ_e values at the surface (> 360 K), producing extreme CAPE values (exceeding 3000 J kg^{-1} along weak surface baroclinic zones.

[8] Note that there are two maxima in the difference field: one is north and northeast of the surface low in a region of strong, saturated ascent over the warm front, while the other is southeast of the surface low, associated with parameterized convection.

3. Upper level jet streaks

Sutcliffe (1939) and Sutcliffe and Forsdyke (1950) recognized that upper-level divergence and the associated vertical motions were closely connected to synoptic-scale cyclogenesis. These ideas have been confirmed repeatedly (e.g., Palmén and Holopainen 1962), and have evolved into what is now known as Sutcliffe-Petterssen development theory (see Uccellini 1990; Bosart 1999).

When observations of the flow in the upper troposphere became available, high-speed westerly flow aloft was discovered, and it was recognized that this high-speed flow was organized into long ribbons that came to be known as *jet streams* (see the textbook by Reiter 1961). Bjerknes and Holmboe (1944) and Bjerknes (1951) developed an explanation for at least some of the *along-stream* variations in upper-level divergence in terms of gradient wind imbalance. However, the observed structures in jet streams include features that have considerable along-flow variation in wind speed, known as *jet streaks* that cannot be attributed solely to curvature effects. It has been proposed (Newton 1959, 1981; Newton and Persson 1962) that inertial oscillations play an important role in the along-flow windspeed variations represented by jet streaks. Although inertial oscillations may be a factor in developing jet streaks, it is likely that other factors (such as persistent convection) also create along-stream changes of the flow speed in the jet streams.

A conceptual model (Fig. 2) of the vertical motions and ageostrophic circulations associated with jet streaks has been developed by numerous authors (Riehl et al. 1952; Beebe and Bates 1955; Shapiro 1981; Bluestein 1993, p. 405; Cunningham and Keyser 1997) and used extensively to diagnose regions of ascent. Ascent, of course, is favorable for cyclogenesis, widespread precipitation, and organized deep, moist convection. Thus, it has been suggested that jet streaks are a feature that is associated with severe convection via their vertical motions and coupled ageostrophic flows (Uccellini and Johnson 1979). The accelerations associated with curved flow and along-flow thermal advection can modify this simple conceptual model significantly (see, e.g., Shapiro 1982; Bluestein 1986; Keyser 1986). At the level of QG theory, the intensity of the cross-jet secondary circulation is proportional to the shear of the ageostrophic wind; the ageostrophic wind is enhanced for compact jet maxima with large along- and cross-flow speed variations.

Major cool season cyclogenesis events involving severe convection seem to show a preference for poleward jet exit regions (e.g., Uccellini 1990). In some instances, overlapping divergence regions in the upper troposphere between the poleward exit region of one jet streak and the equatorward entrance region of another, roughly parallel jet streak can interact to support intense and deep synoptic-scale ascent. This has been associated with explosive cyclogenesis, as in the so-called Presidents' Day storm of 18-19 February 1979 (see e.g., Bosart 1981; Uccellini et al. 1984; Uccellini and Kocin 1987) that, in turn, is seen as setting the stage for severe convection.

Severe forms of DMC can develop within the warm sector of a rapidly deepening cyclone during the cool season, where moist, unstably-stratified air is being advected poleward in low-level jet streams that are coupled to the upper-level jet streaks (Uccellini and Johnson 1979). In the warm season, severe convection can spread well poleward of the warm sector in cyclones. An association between jet streaks and severe convection often can be found during the warm season, as well. During the early spring, intense (or intensifying) extratropical cyclones can bring moist, unstably-stratified air into their warm sectors, just as in the winter. Later in the spring, as hemispheric baroclinity weakens, the extratropical cyclones and jet streams are correspondingly less intense, but static stability may be relatively low, such that a given amount of geostrophic advection [i.e., the rhs of Eqn. (2)] produces stronger vertical motions than in the more stable wintertime.

In the summer, the polar jet stream weakens still more and continues to retreat poleward. Jet streams and the jet streaks embedded within them remain tied to severe convection, however. The retreat of the polar jet stream is typically accompanied by poleward penetration of moisture and instability, and the subtropical jet stream can become important in providing vertical wind shear. Over much of the United States in summer, for example, a subtropical anticyclone dominates the continental plains equatorward of, say, 40° latitude, with active extratropical weather systems and jet streams dominating the northern continental plains and the Great Lakes region. The extratropical synoptic-scale systems of summer are not as intense as their cool-season counterparts but can have easy access to abundant moisture and instability, with a correspondingly large frequency of deep, moist convective storms. Such events can become derechos (Johns and Hirt 1987) embedded in northwesterly flow (Johns 1982) that form on the equatorward sides of jet entrance regions, within a deep layer of warm air advection, with ample moisture and low static stability.

III. Planetary boundary layer processes

The planetary boundary layer (PBL) generally is defined to be the tropospheric layer through within which the effects of the surface are important influences; the depth of the PBL varies considerably in both space and time, but generally is on the order of 1 km. In one sense, the PBL might not be considered a component of synoptic-scale processes. As noted by Stull (1988, p. 2), the time scales of PBL processes can be as small as an hour or less. Obviously, processes on that short a timescale are not appropriate for a synoptic-scale discussion, but the PBL clearly interacts with synoptic-scale systems.

What makes the PBL so relevant in this chapter (that attempts to connect extratropical synoptic-scale processes with severe convection) are the surface sensible and latent heat fluxes. The extratropical cyclone is the dominant process outside of the Tropics for mitigating the meridional gradients in temperature caused by differential solar heating (Palmén and Newton 1969, §10.6). Since most of the retained heat is absorbed at the surface, it is the heat flux from the surface to the atmosphere that dominates the heat transfers governed by ETCs. Some fraction of that incoming heat is used to evaporate liquid water, which thereby becomes *latent* heat available for transport by the synoptic-scale flow.[9] The majority of the diurnal temperature wave is within the PBL, so PBL processes are relevant in understanding the heat and moisture transports in ETCs and their relationship to DMC, which also typically has a strong diurnal dependence.

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[9] The ratio of the sensible heat flux to the latent heat flux is known as the *Bowen ratio* (see Stull 1988, pp. 274 ff.). The surface heat balance requires that the net incoming solar radiation minus any upward heat flux from the soil must equal the sum of the sensible and latent heat fluxes.

1. Simple Ekman theory

A very simple model of the effect of friction on the flow in the PBL can be developed using so-called Ekman theory (e.g., Holton 1992, p. 129 ff.). This simple theory indicates that the effect of friction is to cause air to flow across the isobars towards lower pressure at and near the surface, but the cross-isobaric flow decreases with height such that the effect of friction becomes nearly negligible at the top of the PBL. Based on Ekman theory, a profile of the horizontal wind as a function of height can be derived, such that the wind approaches geostrophic at the top of the PBL. This profile, incidentally, is unstable (see Faller and Kaylor 1966; Lilly 1966). The net result is called *Ekman pumping*, with rising motion associated with cyclones (which should exhibit low-level convergence via Ekman pumping) and sinking motion with anticyclones (which should be characterized by low-level divergence from Ekman pumping). Using this theory, it can be shown that the resulting vertical motion at the top of an Ekman layer is proportional to the low-level geostrophic relative vorticity, and inversely proportional to the so-called Ekman parameter: $\gamma \equiv \sqrt{f/2K}$, where f is the Coriolis parameter and K is the eddy viscosity coefficient (assumed constant in this simple theory). Of course, Ekman pumping is only one of several processes that create vertical motion. At the level of QG theory, Ekman pumping can be incorporated as a bottom boundary condition on ω for the solution of (2). Bannon (1998) has considered Ekman pumping in situations where the free atmospheric flow is not assumed to be steady-state.

The eddy viscosity coefficient K is not really constant with height; in fact, it has been found that K typically increases through a relatively shallow surface-based layer (~100 m in depth) to a maximum somewhat above that, and then decreases rapidly to relatively small values at the top of the PBL (see e.g., Stull 1988; p. 210) Thus, even in uncomplicated conditions, the actual PBL may not fit the purely theoretical Ekman profile very accurately. Further, the eddy viscosity is not constant with *time*, which can have some important impacts on the low-level flow. If it is assumed, only for the sake of simplicity, that the major physical mechanism by which the free atmosphere "feels" the surface is via mixing from convection in the PBL,[10] then the eddy viscosity should be highest at the time of maximum heating during the afternoon. Thus, the isobaric crossing angle (from high to low pressure) associated with an Ekman-like force balance looks something like Fig. 3, where the friction-induced departure from pure geostrophic balance should be a maximum at this time of the day. To the limited extent that the real atmosphere fits this simple model, Ekman pumping should vary during the day, reaching a peak at the time of maximum mixing in the PBL.

2. Diurnal variations and residual layers

Under typical conditions involving solar heating during a sunny day, the PBL has an inversion at its top that ascends as the PBL deepens. During this process, the inversion weakens as it ascends, because potential temperature typically increases with height. This stable layer can act as an inhibitor of deep convection and its erosion by diurnal heating is one reason why deep convection often begins during the afternoon. However, this simple picture of the diurnal modulation of convection has numerous exceptions.

As noted first by Carlson and Ludlam (1968) and developed in some detail by Lanicci and Warner (1991) and Stensrud (1993), surface-based layers can be carried away in the flow from their original locations, resulting in *elevated* residual layers, some of which may be "well-mixed" layers. When such a residual layer is created over dry regions, it tends to be a relatively dry, high-lapse rate mixed layer, the superposition of which over another, potentially cooler surface boundary layer can result in a very strong inversion, sometimes referred to as a “capping” inversion (often referred to as a “cap” or “lid”) that can prevent the release of potential instability. Graziano and

[10] It is not obvious that viscosity in the PBL is always dominated by the presence or absence of dry convection; the presence of vertical wind shear in the PBL also contributes to eddy mixing. The diurnal growth of a neutrally-stratified boundary layer promotes mixing and so in a sheared environment produces vertical momentum transports. The momentum from higher levels mixed downward tends to reflect the flow aloft and so the isobaric crossing angles observed need not fit those expected from simple Ekman pumping. Thermal advection in baroclinic PBLs also complicates the simple picture of a convectively-dominated PBL (Hoxit 1974).

Carlson (1987) have shown that when the "lid strength index"[11] is more than about 2 C, deep convection is likely to be suppressed.

The presence of a capping inversion can inhibit convection at some locations, but it also can *promote* it elsewhere through the seeming paradox of *preventing* the release of potential instability. Convective inhibition results in "storage" of PBL parcels with high CAPE. When the unstable parcels at low levels flow out from under the margins of an elevated mixed layer, thereby permitting DMC development, they are said to be *underrunning* the cap. When some process erodes the cap locally, perhaps by forced ascent along some boundary, DMC ensues. Irrespective of what process eventually enables DMC to occur, its intensity is associated with parcels whose CAPE was created and maintained by PBL processes that perhaps can be far removed from where the convection actually occurs. The challenge of knowing when, where, or even if the cap will be overcome is one of several factors that make forecasting severe convection so difficult (see Ch. 11 in this volume).

As the solar heating begins to decrease in the afternoon, at some point the insolation falls below the outgoing long-wave radiation level, and so the net radiation becomes negative. This radiation loss near the surface promotes surface cooling. The resulting surface-based stable layer effectively de-couples the surface from the air above it, thereby reducing the mixing within the stable layer. A residual well-mixed layer (see Carlson and Ludlam 1968; Stensrud 1993) will remain in place above the stable layer, even as continued cooling at the surface slowly deepens the surface-based stable layer (see Stull 1988; his Fig. 1.7 for a schematic of this process). By preventing the free atmosphere from "seeing" the surface, development of this surface-based stable layer can cause a substantial reduction in eddy viscosity.

During the day, when mixing is large, it is possible to imagine how the friction force (directed opposite to the wind) and the resulting Coriolis force combine to balance the horizontal pressure gradient force, as shown in Fig. 3. Imagine that the mixing is "turned off" impulsively,

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[11] Graziano and Carlson (1987) defined this index as $\left(\theta_{sw1} - \bar{\theta}_w\right)$, where θ_{sw1} is the saturation wet-bulb potential temperature at the warmest point in the inversion, and $\bar{\theta}_w$ is the vertically-averaged

such that the friction force vanishes.[12] Then, as shown in Fig. 4, the loss of friction means an unbalanced pressure gradient force, a component of which is along the wind. Part of the pressure gradient force temporarily balances the Coriolis force at the instant the friction disappeared. The unbalanced part of the pressure gradient force, however, increases the wind which, in turn, increases the Coriolis force. Therefore, the wind increases in speed and turns to the right (in the Northern Hemisphere), producing a rotation of the resultant wind about the geostrophic, the so-called *inertia circle*. The rotation period ($2\pi/f$) is known as the *half-pendulum day*; at 45° latitude, this period is about 16 h.[13] Of course, long before a complete rotation has been made, the heating cycle again begins to re-couple the surface and the air above it. About 4-8 h (depending on the latitude) after the de-coupling begins, the geostrophic departure vector is closely aligned with the geostrophic wind, producing a supergeostrophic wind maximum at that time. This process results in the so-called *nocturnal boundary layer wind maximum* (NBLWM), first elucidated by Blackadar (1957) and since discussed in many papers (e.g., Bonner 1966; Holton 1967; Thorpe and Guymer 1977; Stensrud 1996a). This process operates whenever the diurnal cycle of surface heating is strong and is suppressed when the diurnal heating cycle is weak, as on cloudy days.

3. Baroclinic planetary boundary layers

The foregoing is still an argument that assumes barotropy. There is an additional *baroclinic* process that can be superimposed upon the NBLWM (Holton 1967). Differential heating is often observed over sloping terrain; for example, the dry elevated high plains of the United States typically heat up faster during the day than regions to the east, and cool off faster at night. This causes a diurnal cycle in the horizontal temperature gradient superimposed on the synoptic-scale gradient. The result is a diurnal cycle in the thermal wind vector, such that the poleward flow

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value of wet-bulb potential temperature between 30 and 80 mb above the ground.

[12] The reduction of eddy viscosity to negligible levels does not occur instantaneously, of course; this simplification is simply for illustrative purposes.

should increase with height at night, at least within the baroclinic zone created by this differential heating. Thus, this mechanism is roughly synchronous with the NBLWM and can create significant horizontal variations in the NBLWM flow, resulting in a low-level jet *stream* (LLJS). LLJSs are herein distinguished from NBLWMs and the ambiguous term "low-level jet" is avoided specifically to make this distinction (as did Stensrud 1996a). It appears that relatively narrow ribbons of strong low-level flow (LLJSs) arise in a variety of contexts (e.g., see Browning and Pardoe 1973; Uccellini and Johnson 1979; Nagata and Ogura 1991; Doyle and Warner 1993; Chen et al. 1998).

Under baroclinic conditions, furthermore, thermal advection within the boundary layer can create significant departures from simple Ekman theory (see, e.g., Hoxit 1974; Bannon and Salem 1995), In the Northern Hemisphere, the thermal wind relation indicates that the geostrophic wind veers (backs) with height under conditions of warm (cold) thermal advection. As noted in Maddox (1993), diurnal changes in the PBL wind profile can have a large impact on the hodograph (especially in the critical part near the surface) and, hence, on the potential for severe convection.

Further, given the progression of synoptic-scale cyclones from west to east across mid-latitudes, the low-level flow switches repeatedly from poleward to equatorward and back again to poleward (see Lanicci and Warner 1991) with the passage of each cyclone and the approach of the next. The tendency to develop a narrow ribbon of strong poleward flow ahead of approaching ETCs is enhanced by leeside troughing processes (Uccellini 1980; Lanicci and Warner 1991) in the plains region of the United States. Therefore, the LLJS often is tied dynamically to processes operating in the upper troposphere (Uccellini 1980), as well as being modulated by PBL-associated processes that have diurnal cycles.

The LLJS and NBLWM have long been recognized as major factors in promoting nocturnal convection (Means 1944, 1952; Pitchford and London 1962; Bonner 1966; Maddox 1983) during the summer in the plains of the United States. Typically, they prolong the life of

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[13] At 30° latitude, the period is very close to exactly 24 h and so the inertial oscillation is in phase with the heating cycle at that latitude.

convective systems that were initiated during the day, extending the threat of severe weather well into the night. They also play a role in enhancing poleward heat transport (both sensible and latent heat) in ETCs. For a review of the factors in low-level winds, see Stensrud (1996a) and more information about the relevant processes can be found in Bluestein (1993, p. 391 ff.). We already have observed that meridional heat transport for purposes of mitigating the unequal solar heating can be viewed as a major reason for the existence of extratropical cyclones; to this extent, then, PBL processes must be considered an important element of synoptic-scale processes. As we have only been able to describe briefly here, the wide range of processes that govern the PBL's structure and evolution (e.g., solar heating cycles, cloud cover, antecedent precipitation, fronts, drylines, the character of the underlying surface, etc.) are critical in determining when and where stored CAPE is released in its severe manifestations.

IV. Climatology of deep, moist convection

It is of interest to develop a synoptic-scale sense of the climatological occurrence of DMC. A way of viewing the existence of convection within midlatitudes is that convection develops whenever the redistribution of heat by synoptic-scale processes is not sufficient to mitigate the imbalances resulting from differential heating at the surface. ETCs produce prodigious meridional and vertical heat transports; Palmén and Newton (1969, p. 301 ff.) estimate that the *poleward* heat transport per cyclone across 45°N on an April day is about 10^{12} kJ s^{-1} and they note that about six such disturbances around the globe at any one time would suffice to compensate for the associated meridional imbalance in radiative heating. They also estimate the *upward* heat transport across 500 hPa per cyclone to be about 2 x 10^{11} kJ s^{-1}. Again, they note that this amount is roughly comparable to that needed to balance the excess heat input at low levels.

Whereas the typical convective cell transports but a miniscule fraction of the amount of heat transported by an ETC, the large vertical motions of DMC can transport heat much more rapidly than the relatively weak vertical motions of ETCs. The temporal and spatial distribution of convective storms can be viewed as depicting those times and locations when non-convective heat transport processes (especially in the vertical) simply have not been adequate to mitigate the excess heat (both sensible and latent, of course) accumulating at low levels. Thus, as convective cells become more numerous and intense, the implication is that more such rapid vertical transport is needed.

The concatenation of moisture and instability necessary for DMC is quite obviously tied to this process of excessive accumulation of sensible and latent heat in the lower troposphere. Clearly, convective transports are primarily vertical, but they also affect horizontal transports by carrying heat upward to upper-tropospheric levels where they can be redistributed horizontally by the relatively strong winds aloft. The coupling of convection to synoptic-scale processes is still being explored (see, e.g., Gutowski and Jiang 1998).

The release of the excess sensible and latent heat may not be exactly where it was created because the associated CAPE was "stored" in the atmosphere (Emanuel 1994) until it could be released. Convective inhibition is the main reason for this storage, and so synoptic-scale processes can transport the excess heat to a location well away from where it entered the atmosphere. Release of the instability is intermittent and may not occur at the time and location of the destabilization processes that created it. This makes parameterizations of convection difficult, since its occurrence is tied to a number of ingredients that can interact in nonlinear ways, rather than a single parameter.

1. The global spatial distribution of convective precipitation

On a world wide basis, an average of about one meter of precipitation falls annually. This precipitation is actually distributed quite unevenly, of course, with the bulk of it falling in the Tropics along the so-called Intertropical Convergence Zone (ITCZ), the South Atlantic and South Pacific Convergence Zones (Vincent 1994), the western Pacific oceanic warm pool (Ramage 1975), Amazonia and Latin America, central Africa, across the oceanic storm tracks of midlatitudes, and on the windward slopes of mountains. These regional maxima also appear in the global climatology presented by Huffman et al. (1997), based on satellite image datasets obtained in conjunction with the Global Precipitation Climatology Project. Indirect measures of precipitation (e.g., based on satellite imagery) can be compared with other recent global climatologies (Jaeger 1976; Legates 1987; Legates and Wilmont 1990), as well. Agreement is generally good among these different climatological distributions, although uncertainties remain in the absence of in situ measurements over the vast oceanic areas.

Hsu and Wallace (1976) have summarized the annual variation of global precipitation. They have shown that precipitation over the continents peaks during the warm season in middle and low latitudes, and tends to follow the sun in the Tropics, except in the deep Tropics, where there is little annual variation. Other exceptions are associated with tropical monsoon rainfall, where rainfall is heavily concentrated in the warm season.

Although these global precipitation climatologies do not reveal the distribution of rainfall from DMC directly, the vast majority of the rainfall within the Tropics and during the warm season outside of the Tropics is dominated by the DMC contribution. Orville and Henderson (1986) mapped the global distribution of midnight lightning (i.e., *local* midnight) from September 1977 to August 1978, showing that midnight lightning exhibits a strong preference for the continents in low and middle latitudes. This suggests that *maritime* tropical DMC is only weakly electrified, a hypothesis that has been confirmed recently by Zipser (1994).

Laing and Fritsch (1997) have mapped the global distribution of Mesoscale Convective Complexes (MCCs, a subset of Mesoscale Convective Systems; Maddox 1980; Smull 1995). Their distribution is shown in Fig. 5. Figure 5 shows that MCCs in North and South America occur preferentially to the lee of mountain ranges in midlatitudes, where LLJSs are common [see Fig. 4 in Laing and Fritsch (1997)]. The distribution in the Old World is predominantly in the Tropics and Subtropics. Given that considerable DMC occurs in regions other than those favored by MCCs (e.g., Amazonia), organized mesoscale convective systems seem to be associated with a some particular combination of thermodynamic and synoptic-dynamic processes, perhaps modulated by topographic effects (see Ch 9, this volume).

Wallace (1975) examined the diurnal variation of precipitation and thunderstorm frequency across the United States. His results confirmed the presence of a warm season rainfall maximum over much of the United States, with the exception of coastal New England and much of the coastal West. The High Plains region to the lee of the Rocky Mountains has a notable warm season precipitation peak, with much of the rainfall associated with nocturnal DMC activity. Thunderstorms over the Rocky Mountains, Florida, and the East Coast tend to occur more in the late afternoon and early evening. As shown in Riley et al. (1987), the nocturnal storms over the Plains tend to have two sources: (1) afternoon storms that develop initially over and near the Rocky Mountains and then move eastward, and (2) storms that develop locally over the High Plains in associations with a variety of mesoscale weather systems. These findings have been verified recently in a comprehensive investigation of summertime precipitation over the United States by Higgins et al. (1997).

Globally, DMC is not uniformly observed owing to little or no data over the oceans and the sparsely populated regions around the globe. The recent availability of various remote sensing systems provides views that approach being truly global in scope (especially satellite-borne sensors). In fact, this is a major aspect of the ongoing Tropical Rainfall Measuring Mission (TRMM; see Simpson et al. 1988). From these observations, DMC occurs most frequently over land, with the important exception of the ITCZ band near the equator. The predominance of DMC over land surfaces could be the result of the lower heat capacity of soil versus water. A given amount of input radiation produces a larger near-surface temperature increase over land than over water, increasing the amount of instability. The result is that the land typically develops larger potential instability than the oceans (and large lakes). When sufficient low-level moisture is present, this means more frequent DMC over land (assuming that some lifting mechanism is present). However, it is not entirely obvious why continental updrafts in the Tropics are typically stronger than their maritime counterparts, since according to Lucas et al. (1994), they have not found large differences in CAPE.

The equatorial band of convection is usually found on the "summer" side of the equator. Since a major proportion of the surface in the tropical band (between the "Tropics" of Cancer and Capricorn at 23.5˚N and S, respectively) is maritime, much of the excess insolation goes into warming tropical seas. Tropical synoptic weather systems do not usually involve much vertical wind shear (as noted in Ch. 7 of this volume) and a typical maritime sounding in the Tropics develops a near neutrality to DMC through much of the troposphere (e.g., Fig. 6), as discussed by Xu and Emanuel (1989) or Williams and Renno (1993), among others. Tropical maritime convection is therefore relatively uninhibited but also is not characterized by much CAPE, either. Tropical continental regions are famous for the presence of rain forests, with a predominance of convective rains. DMC plays a very large role in the synoptic-scale tropical heat transport, which can be viewed successfully as a "Hadley cell" on average (Lorenz 1965).

Another exception to the dominance of continents over oceans in the distribution of DMC is over the surface currents of warm water flowing poleward on the eastern boundaries of continents (e.g., the Gulf Stream or the Kuriosho Current). As sources of both low-level sensible and latent heat, these currents also show up as regions of enhanced frequency of DMC, notably in the cool season, at a time when the continents tend to be dominated by conditions not conducive for DMC. Much of this convection is associated with strong synoptic-scale cyclones.

Regions of complex terrain often reveal local regions favoring the frequent development of DMC, even when relatively moisture-poor. Acting as an elevated heat source, especially in the warm season, complex terrain can be associated with considerable convective weather, although the resulting DMC does not necessarily produce much precipitation. For example, the Rocky Mountains of North America display a pronounced peak in summer convective frequency (López and Holle 1986) despite being relatively arid. High-based thunderstorms that produce little or no precipitation often initiate fires that can be a significant societal problem even if they are not usually considered to be “severe” thunderstorms. In Australia, for example, thunderstorms that produce little precipitation represent a significant fire hazard. Australian operational severe thunderstorm forecasters regularly review fire danger and issue fire weather forecasts (Love and Downey 1986; Williams and Karoly 1999).

In subtropical areas, or in regions with extensive regions of warm oceans (e.g. the Mediterranean Sea), the land-sea breeze circulations can be important in developing DMC even in synoptically-unfavorable regimes. This role for land-sea breeze circulations can be enhanced for islands and peninsulas (Pielke 1974). Thus, for example, subtropical peninsular Florida is noted for its frequency of warm-season thunderstorms even though the synoptic flows in the region remain generally weak. Relatively minor synoptic-scale fluctuations still can modulate the frequency and location of DMC, however (see Blanchard and López 1985).

2. The global temporal variation

As with the spatial distribution, the temporal distribution of DMC tends to be associated with processes that modulate conditional instability: notably, the diurnal and seasonal heating cycles. In the regions where most DMC is associated with complex terrain or sea breeze circulations, the greatest DMC frequency tends to be coupled to, but lagging somewhat behind, the peak solar heating. However, there are at least three notable departures from this average behavior. First, strong synoptic-scale processes can become the dominant factor in developing DMC and these may or may not occur in phase with the daily heating cycle. Thus, in the cool season in the southern United States around the Gulf of Mexico, for example, DMC is not so dominated by the diurnal cycle because DMC can be tied to intense synoptic-scale processes.

Second, at the terminus of a LLJS, the moisture-bearing poleward flows typically impinge on a thermal boundary (either a front or an outflow boundary from previous DMC), favoring ascent and convection initiation. Since this is enhanced by the typical NBLWM process, such convection can develop after sunset. Although the initial convection often develops late in the afternoon, it can be sustained well after dark in places where the LLJS (enhanced by the NBLWM) encounters a thermal boundary (e.g., Trier and Parsons 1993). This favors nocturnal convective systems (Laing and Fritsch 1997), a process that occurs in many places around the world (recall Fig. 5). Severe weather with these convective systems tends to occur before or shortly after dark, whereas their precipitation may continue for many hours after sunset (Maddox 1983).

Third, recent evidence (e.g., Gray and Jacobson 1977; Janowiak et al. 1994) indicates that over the tropical oceans, convective activity tends to peak during the night and early morning hours.

This cycle is out of phase with the diurnal cycle, but its amplitude is not as large as that over land areas (which, with some exceptions as noted, is much closer to being in phase with the diurnal cycle). Nevertheless, the evidence for this modest diurnal variation in oceanic tropical convection is compelling. Brier and Simpson (1969) have suggested an important role for the semidiurnal tide in modulating tropical rainfall, which is predominantly convective. Chen and Houze (1997) have argued that the nocturnal peak in convective activity in the tropical oceans is due to a complex interaction among the diurnal heating cycle, large-scale processes in the Tropics, radiative effects, and the typical life cycles of convective systems.

The seasonal cycle of DMC also tends to follow the seasonal cycle of conditional instability. Naturally, synoptic-scale processes modulate this significantly. DMC is a feature near the cores of many explosively-deepening maritime cyclones (e.g., Bosart 1981; Reed and Albright 1986; Nieman and Shapiro 1993; Dickinson et al. 1997) even though such cyclones are infrequent outside of the cold season. The impact of warm ocean currents on developing DMC in the cold season has already been mentioned. Maritime and subtropical regions have a much less pronounced seasonal cycle, and so may experience DMC events occasionally outside of their respective warm season maxima; synoptic-scale control on such events is typically high.

Occasionally, thunderstorms develop in winter weather systems. These arise in at least three distinct ways. The first way involves the development of a warm sector that has near-warm season characteristics and so the DMC is virtually the same as that developing in the warm season. A second process is when high lapse rates (in relatively dry air) are carried over a cold and stable but relatively moist surface-based air mass, producing “elevated” DMC that simply produces snow (or freezing rain) rather than simple rain. The third process is associated with slantwise ascent and, possibly, deriving its energy from conditional symmetric instability (CSI - see Emanuel 1994, Ch. 12; Schultz and Schumacher 1999). The release of CSI is confined to *saturated* environments but can produce rain, as well as either freezing or frozen forms of precipitation. Thus, two of the three processes producing thunderstorms in association with winter weather systems involve “elevated” DMC; that is, the parcels participating in the DMC are not surface-based because the surface-based air mass is markedly stable (Colman 1990a,b). Winter thunderstorms are never common and so represent a mostly negligible element in the climatology of DMC, although they certainly can be important when they occur; snowfall rates associated with snow thunderstorms can reach magnitudes of 10-30 cm hr^{-1}. Virtually no severe weather other than heavy precipitation is associated with the second and third forms of winter DMC.

V. Climatology of severe convection

A global climatology of severe convection is not generally available, owing to the absence of international commitment to the development of such a database. Even in the United States, which is unquestionably the single nation with the highest frequency of severe convection, a number of deficiencies exist in the climatological record of severe storms. Further, the definition of what constitutes a "severe" convective event is typically arbitrary, whereby some phenomenon is called severe when it meets or exceeds some specified criterion, and the criteria vary around the world. For convective windgusts, a speed threshold is used; for hail, hailstone diameter is the variable. Tornadoes are an exception, in that a convective storm producing any sort tornado is usually considered to be severe. However, even this can become somewhat arbitrary. When a tornado occurs over the water, thereby being called a waterspout, the associated deep convective storms that produce the waterspouts usually are not designated as severe. In the United States, heavy rainfall is not considered "severe," although many other countries around the world *do* have some sort of threshold rainfall (or rainfall *rate*) that is deemed severe.

Nevertheless, we shall attempt to give at least some picture of the global distribution of severe weather, but the reader should be aware of its deficiencies. Our main interest in the climatology of severe convective weather is that this gives some clues about the synoptic-scale patterns that favor severe weather, as implied by Ludlam (1963). Our presentation focuses on hail and tornadoes. Convective windstorms are quite difficult to classify into severe and nonsevere categories, since subjective estimation of windspeed is so poor and the actual windspeed measurements are so sparse relative to the events. Heavy rainfalls are probably the most global in scope of all DMC-associated events, but the lack of consistent criteria and high-resolution raingage networks worldwide make it problematic to depict heavy rainfall climatology in detail. Even in the United States, there is no systematic record kept of important convective rainfall *events* (such as flash floods), other than the routine precipitation climatological data (see Brooks and Stensrud 2000).

Hailstorms are sufficiently important owing to their economic impact worldwide that some records are kept in most nations that have hailfalls at all regularly. Figure 7 shows one picture of the global annual hail day (i.e., a day with one or more hail events) frequency distribution, indicating where the frequency is at least one hail day yr^{-1}. The size of the hail is not included in this record, which apparently is mostly associated with crop losses. The apparent infrequency of hail days over Australia is probably a direct reflection of the extremely low population density (and, hence, observations) away from the coasts in that continent. Nevertheless, Sydney, Australia has been hit recently by two devastating hail events (in 1990 and 1999), in spite of what might well be a truly low annual frequency.

Ludlam (1963) presented a global climatology of "severe squall-thunderstorms" (his Fig. 5), which bears some resemblance to Fig. 7. He indicates that the distribution of midlatitude events appears to be associated with the jetstream axis and its associated vertical windshear. Given what we know about hailstorms, this seems like a plausible hypothesis. As hailstone diameter increases, the likelihood of its occurrence from a supercell also increases (since supercells typically have the strongest updrafts), and supercells are known to associated with the presence of significant vertical windshear (Weisman and Klemp 1982, 1984).

Over the United States, hailstone *size* is estimated (albeit imperfectly; see Sammler 1993) and so it becomes possible to depict the climatological frequency as a function of hailstone size. As shown in Fig. 8, the distribution of large hail (≥2 cm [3/4 in]) in the United States is concentrated in the central and high plains, to the lee of the Rocky Mountains. This is even more evident when considering the distribution of giant hail (≥5 cm [2 in]; Fig. 9). Heavy falls of hail < 2 cm in diameter can have significant economic impact, especially on crops. As already noted, giant hailstones ($\geq$ 5 cm) are mostly associated with supercell thunderstorms (Nelson and Young 1979; Ludlam 1980, p. 256 ff.; Burgess and Lemon 1990). Supercells are capable of producing the full range of severe convective phenomena (damaging convective wind gusts, tornadoes, and heavy rainfalls), in addition to giant hail.

Since the association between severe convective weather events and supercells is quite high, it is likely that the reason for the dominance of the United States in the global climatology of severe forms of convection is its topography, which favors the occurrence of supercells (Ludlam 1963; Doswell 1993) on its west-central plains. This does not mean that supercells are unheard of outside of the United States (see, e.g., Browning and Ludlam 1962; Ludlam 1963; Doswell and Brooks 1993b; Schmid et al. 1997).

In places where supercells are possible, the occurrence of tornadoes becomes much more likely, although by no means guaranteed. In the United States, prior to the 1980s, it was believed that thunderstorms in Colorado were prolific hail producers and it was known widely that hail events in Colorado sometimes were associated with supercells (see Marwitz 1972), but tornadoes were considered rare. Since the 1980s, there has been a rapid growth of tornado reports in Colorado, to the point where the state is often among the national leaders in reported tornadoes each year. A belief in the infrequency of tornadoes can be a self-fulfilling prophecy; if tornadoes are considered unlikely, there can be little motivation for keeping track of the number of tornadoes, especially in sparsely-populated regions where it is difficult to get any information at all about what might have happened. For example, Snow and Wyatt (1997) found that no tornadoes have been documented in Spain since World War I. This is perhaps nominally true but it does not correspond with the known facts; for example, tornadoes in Spain are reported by Ramis et al. (1995; 1997) and in the Balearic Islands, a part of Spain (Gayá et al. 1997) and there are even widely-circulated videotapes of Spanish tornadoes. Nevertheless, there is as yet no formal, national program to document tornadoes in Spain and so none have been "recorded".

Fujita (1973) attempted to discuss the worldwide climatological distribution of tornadoes, but his presentation acknowledges the same lack of consistent international documentation. What is known about tornadoes outside the United States tends to be anecdotal rather than systematic (e.g., Dotzek et al. 1998; Peterson 1998; Xu et al. 1993). Only in isolated nations has there been any attempt to be comprehensive [e.g., Dessens and Snow (1993) reviewed tornadoes in France] and there remain problems with relative short records of erratic quality. Taking into account all of the known factors associated with underreporting of tornadoes, it is probable that the worldwide pattern of tornado occurrences is not all that dissimilar from that of hail, although at a lower frequency than hail events. This appears to be the case in the United States (Fig. 10; cf. Fig. 8). The distribution of violent tornadoes [F4-F5 on the Fujita (1971) scale; Fig. 11] shows the same west-central plains peak as that of giant hail.

Curiously, in the United States, there is considerable difference between the distribution of all severe convective wind reports (including "wind damage" reports, without an estimate of the wind speed, as well as estimated wind speeds exceeding 25 m s^{-1} [50 kt]) from that of hail (Fig. 12; cf. Fig. 8). It appears that convective wind events arise in a large variety of ways, in situations with both weak and strong synoptic-scale processes underway (e.g., Johns and Hirt 1987; Johns 1993) and so are more widely distributed in the United States, with a rather different climatological distribution from tornadoes and hail. Worldwide, unfortunately, there is virtually no documentation of severe convective wind events. The distribution of *extreme* convective wind gusts ($\geq$32 m s^{-1} [65 kt]; Fig. 13) is again similar to that of giant hail, with a peak in the west-central plains.

The superposition of high lapse rate air in the lower mid-troposphere flowing off the high terrain to the west of the plains above the low-level poleward flow of moisture creates the extreme CAPE values seen in classic severe weather soundings (Doswell et al. 1985). Combining this CAPE with sufficient baroclinity to enhance the vertical wind shear means the United States west-central plains experiences the highest frequency of extreme forms of severe convection. The decrease in giant hail, extreme convective wind gusts, and significant tornadoes to the east of the Mississippi River suggests that in the United States, CAPE is typically released before synoptic scale processes can carry it very far. When DMC is inhibited by a "cap", however, it can be transported eastward, resulting in the occasional outbreaks of extreme severe convection east of the Mississippi.

VI. Synoptic factors associated with severe convection

As already noted, DMC requires three basic ingredients: moisture, instability (high lapse rates), and lift. Assuming sufficient moisture and conditional instability are already present, a frequently important forecast issue is the sufficiency of the lift to overcome convective inhibition (CIN). Some uncertainty remains about the precise details of how DMC is initiated. Certainly if there is forced ascent of parcels that have negative buoyancy during the lifting process, then mass continuity requires that there must be mass convergence beneath the ascending currents, often manifested as "boundaries" exhibiting this convergence (e.g., fronts, drylines, and non-frontal windshift lines). Deep convection usually begins as isolated convective cells along boundaries exhibiting low-level convergence or in groups of relatively isolated cells (Bluestein and Jain 1985). As we will discuss below, topographic processes also are important focusing mechanisms for the initiation of DMC.

The vertical motions in convective storms are of order 10 m s^{-1}, whereas synoptic scale vertical motions are typically of order 1-10 cm s^{-1}, a value two to three orders of magnitude smaller than that of the convective drafts. It can be argued that on the scale of a convective cloud (~10 km horizontal length), synoptic-scale vertical motion is quite negligible. Nevertheless, as mentioned earlier, the occurrence of deep, moist convection (DMC) in association with synoptic-scale weather systems is well-known. How does this association arise?

Consider Fig. 14; in both schematic examples, the parcel must be lifted 150 hPa (corresponding to about 1.5 km) to attain the LFC, but in the one case, there is a CIN of 10 J kg^{-1}, and in the other the CIN is 100 J kg^{-1}. Obviously, more energy must be supplied to lift the parcel that distance in the second example. In either case, if the parcel is rising at a synoptic-scale rate of 3×10^{-3} hPa s^{-1}, such an ascent will take on the order of 5×10^{4} s, which is more than half a day. Clearly, this is too long to represent an important contribution to convection initiation in many cases.[14] Presumably, some process operating on "subsynoptic" scales (e.g., fronts, drylines, gravity waves, etc.) must supply the energy needed to overcome the CIN and lift potentially buoyant parcels to their LFCs (e.g., see Wilson and Schrieber 1986). However, the synoptic-scale motions condition the environment by this slow ascent, causing a reduction in CIN by lifting and weakening any convection-inhibiting stable layers.

This also suggests that convection can be initiated in the *absence* of synoptic-scale ascent (or even in regions of synoptic-scale descent), provided some subsynoptic lifting process can initiate DMC, whenever the moisture and instability are already sufficient to create the potential for DMC. Subsynoptic-scale processes (like upslope flows, fronts, etc.) often operate in association with synoptic-scale systems, so the association between cyclones and DMC is not coincidental but there is no unique relationship, either.

1. Topographic influences.

Topographic features play an important role in the evolution of synoptic-scale weather systems and, therefore, in the development of convection. Note that "topography" is a general term for the physical features of a region (including bodies of water, vegetative cover, snow and ice, soil type, and so on) and is not limited to features of the surface height (orography). There are several quite different types of topographic influences and these are discussed in the following sections.

<u>a. Orographic effects</u>

An important factor in the evolution of a synoptic weather system is its interaction with large-scale *orographic* features. Synoptic-scale cyclogenesis does not occur completely randomly about the midlatitudes but rather is concentrated in certain key geographic areas, often in the lee of major mountainous regions (Roebber 1984; Tibaldi et al. 1990). Although lee cyclogenesis is not

[14] This ignores the effect of synoptic-scale ascent on the sounding itself, of course. That ascent would tend to reduce any CIN and make initiation of convection more likely.

completely understood even now (e.g., Mattocks and Bleck, 1986; Egger 1988), the characteristic structures that develop as a cyclone intensifies are dependent on the topography of the region. For example, in North America, there are three main regions of cyclogenesis: to the lee of the Canadian Rocky Mountains in Alberta, to the lee of the American Rocky Mountains in Colorado, and over the eastern coast of the United States near the Gulf Stream. Although the East Coast cyclogenesis region also is to the lee of the Appalachian Mountains, it is not quite so obviously dominated by the lee cyclogenetic effect, since the Gulf Stream makes the East Coast the location of a quasipermanent baroclinic zone, notably during the cool season (Bell and Bosart 1988).

The Alberta and Colorado cyclogenesis regions are most active when the main belt of zonal flow is in their vicinity. Hence, both can be active in the cool season, but the Colorado region is not so active in the warm season, since the westerlies in the warm season can migrate well poleward of Colorado for extended periods (Atallah and Bosart 1996). Whereas cyclones developing in Colorado have reasonably consistent access to the high low-level moisture of the Gulf of Mexico, even in the cool season, the Alberta cyclones can develop without much absolute moisture content. During the warm season, with the poleward progression of the subtropical anticyclones into North America, however, cyclones developing in the Northern Plains of the United States and the southern parts of the Prairie Provinces of Canada can have rich moisture nearby. Clearly, the ability of a cyclone to import substantial low-level moisture into a region is a key contributor to DMC activity.

Another aspect of topographic effects is the development of thermal contrasts. Midlatitudes are the focus for the main thermal contrast between the equator and the poles, of course, but east-west oriented mountains can inhibit the meridional flow of contrasting air masses. The plains of North America are ideal in this regard for the development of conditions conducive to severe weather; there is no mountain barrier between the Tropics and the pole. Cyclones that affect North America can easily bring the ingredients together for severe forms of DMC simply because of the unique geography of the region. That is, moisture and instability can be brought together readily by these ETCs, with the baroclinic zone favoring the presence of vertical wind shear. Further, the lee slopes of the Rocky Mountain barrier promote the development of poleward flows of sensible and latent heat and, at the same time, the high terrain of the west acts as an elevated heat source, as described already.

b. Other topographic effects

Some *non-orographic* topography can play a substantial role in modulating the behavior of cyclones. For instance, the Great Lakes of the United States have been shown to favor cyclonic circulations around in their vicinity during the winter (e.g., Petterssen and Calabrese 1959; Sousounis and Fritsch 1994). On the other hand, Harman (1987) has shown that anticyclones avoid the Great Lakes in the winter and tend to pass over them frequently in the summer. Thus, if a lake, grouping of lakes, or an inland sea is large enough, it can alter the dynamics of extratropical weather systems, as well as the more obvious effect of evaporation of moisture into the airstream passing over the water. The diagnoses by Alpert et al. (1996; see especially their Fig. 2) suggest strongly that cyclogenesis can be the result of a variety of distinct processes and interactions. Topographic effects can interact synergistically with other physical processes (e.g., convection) in complex ways. As a result, the mixture of processes promoting cyclogenesis at any particular time can vary during a particular cyclogenetic event.

In different parts of the world, the ingredients for severe DMC are brought together in different ways. In the western Mediterranean, for example, a long fetch of poleward flow ahead of an approaching cyclone brings in low-level air from the deserts of northern Africa, which is not conducive to severe forms of convection. When the low-level flow has a long *easterly* fetch over the warm waters of the Mediterranean, however, it can be quite rich in moisture (see Doswell et al. 1998). The ingredients remain the same, but the topography changes the details by which the ingredients are concatenated as synoptic-scale systems interact with that topography.

In regions of complex terrain, subsynoptic-scale, solenoidally-driven flows are quite common; Whiteman (1990) and Egger (1990) provide recent summaries of these. Banta (1990) reviews the impact of complex terrain on clouds, and on convective rain in particular. In essence, complex terrain offers a number of different processes that can provide sufficient lift to make DMC possible. Therefore, when sufficient moisture is present, mountainous terrain often supports vigorous convective activity even in the absence of favorable synoptic-scale processes. This is reflected in the climatological frequency of thunderstorms in favored locations in complex terrain noted earlier, wherein thunderstorms become a nearly daily occurrence in selected mountain areas at the height of the warm season.

c. Air mass characteristics

The vertical structure of the atmosphere depends on, among other factors, the character of the various airstreams at different levels moving over a point. Given the existence of substantial vertical shear of the horizontal wind in mid-latitudes, the patterns of vertical thermodynamic structure associated with severe forms of DMC arise in part as a consequence of the superposition of air masses. Fawbush and Miller (1954) presented examples of the "typical" structures. As noted by Carlson and Ludlam (1968) and Doswell et al. (1985), these structures can arise from superposing airstreams with different properties at different levels. The classic "loaded gun" sounding typical of tornadic situations arises from the superposition of a flow of moist air at low levels (see below) and a midtropospheric air mass with high lapse rates that originates elsewhere. In the plains of the United States, midtropospheric high lapse rates originate on the elevated terrain of the Rocky Mountains and are advected as an elevated residual mixed layer (see Carlson and Ludlam 1968; Lanicci and Warner 1991; Stensrud 1993) over the low-level poleward flow of moist air in the southwesterly airstream ahead of an approaching ETC. The weak static stability of this air mass enhances the upward motion ahead of the advancing cyclone, as we have discussed.

Note that the low-level moisture is also a critical ingredient. For the plains of the United States, this moisture is typically brought poleward first at or near the surface in the airstreams ahead of the migrating synoptic-scale cyclones. The presence of adequate moisture in that low-level flow, however, is by no means assured, especially early in the Spring (Crisp and Lewis 1992; Lewis and Crisp 1992). Many issues influence the structure of the low-level moisture field; hence, its forecasting can be a challenge (Thompson et al. 1994).

2. Solenoidal circulations over simple terrain

Many flows associated with solenoidal circulations are subsynoptic in scale, including classic fronts [not adequately described by QG theory; e.g., see Williams and Plotkin (1968)]. Naturally, fronts are an important aspect of ETCs. That fronts are associated with vertical circulations is well-known (e.g., Eliassen 1962; Bluestein 1993, p. 297 ff.), of course. As already noted, vertical motions produced along and near fronts are a ready source of subsynoptic scale lift (of order 10 cm s^{-1} or larger) for initiating DMC. Classical fronts that develop in ETCs are not the only solenoidally-associated flows that can provide initiation mechanisms for DMC, however. The majority of these other thermally-driven circulations probably fall into the time and space scales known as mesoscale (reviewed in Atkinson 1981 [p. 123 ff.], and discussed in Ch. 3 of this monograph), and so are only mentioned briefly here.

Another solenoidally-driven flow that has a large impact on DMC is the land-sea breeze. This process has been recognized as an important contributor to initiating DMC for some time (see, e.g., Pielke and Segal 1986). As with the comparable thermally-driven circulations in complex terrain, the local topography has an important impact on where and when DMC is likely to be initiated. There is a modulation of the process by its interaction with the synoptic-scale flow. The movement of the land-sea breeze is retarded when it is moving against that synoptic flow, and enhanced when moving it. Typical diurnal movement of the boundary defining the land-sea breeze "front" is on the order of 10 km, so the favored zones for DMC initiation are usually confined to the vicinity of the coastline.

Convectively-generated outflow boundaries are yet another solenoidally-driven circulation that can initiate convection. In situations where DMC evolves into large mesoscale convective systems (MCSs), outflow regions grow to meso-□scale (Orlanski 1975) or larger when numerous convective cells contribute to the pool of precipitation-cooled outflow. These large outflows can be detected and followed readily in many surface observation networks, although the details can not be said to be well-resolved outside of a true mesonetwork.

There is growing evidence for a variety of solenoidally-generated flows associated with inhomogeneities in the underlying topography (e.g., Sun and Ogura 1979; Segal and Arritt 1992; Hane et al. 1997). These might be associated with albedo differences, vegetated versus bare soil or irrigated versus non-irrigated regions, soil moisture availability differences due to antecedent precipitation (Ookouchi et al. 1984), variations in snow cover (Johnson et al. 1984), and so on. At times, they have been described as an "inland sea breeze" (Sun and Ogura 1979) in analogy with ordinary sea breezes. As discussed in Ch. 3, evidence is accumulating that these non-traditional sources of solenoidally-driven flows can an important factor in DMC initiation (e.g., Colby 1984).

Finally, the dryline (Schaefer 1986) is a special case where solenoidal circulations may or may not be present. The thermal gradient across the dryline in the United States undergoes a diurnal oscillation, changing directions during the heating cycle. In the morning, on the dry air side, temperatures are colder than on the moist side; by afternoon, this reverses. What makes it challenging is that the moisture content in the afternoon contributes to density in the opposite way to the temperature, apparently resulting in a near-cancellation of the density gradient. Given natural variability, limited sampling, and finite measurement accuracy, it has not yet been shown definitively that the ascent at the dryline during the afternoon is driven by solenoids. For instance, Ziegler and Hane (1993) present a case study where a virtual temperature (and, hence, density) gradient still exists across the dryline in the afternoon, implying that solenoidal circulation might explain the ascent. However, their sample was a single case, and their capacity to sample the structure and evolution of the dryline was limited. Much remains to be shown about how the dryline acts to initiate DMC (Ziegler and Rasmussen 1998)

3. Monsoons

Monsoons have been defined as a seasonal shift in the direction of the surface winds (Huschke 1959). Monsoons have been reviewed several times in the literature (e.g., Ramage 1975; Boyle and Chen 1987; Webster et al. 1998). Although it certainly is an oversimplification, monsoons can be thought of as a *seasonal* analog to the solenoidally-driven diurnal flows.

Basically, the seasonal shifts in surface flow are being driven by continental-scale land-sea temperature differences. A tendency for ascent over the warm continents in the warm season drives a flow toward the continent from the cooler oceans, whereas descent over the continents in the cool season drives a flow outward from the continent. The monsoons do not initiate convection, per se, but they are associated directly with processes that *do* provide subsynoptic scale ascent, such as upslope flows, monsoon-associated weather systems, and embedded solenoidally-driven flows. The most well-known monsoons globally tend to be associated with the Tropics and Subtropics in Asia (e.g., Ramage 1975; Chang and Krishnamurti 1987), but the midlatitudes have monsoon-like flows as well (see, e.g., Lau and Li 1984; Douglas et al. 1993). These seasonal tendencies can enhance diurnal flows that also are thermally-driven and so can play a role in the development or suppression of DMC (e.g., Doswell 1980). Moreover, they can import low-level moisture from over the oceans to participate in DMC over land, and so can be a factor in creating a suitable environment for DMC as well as in the initiation process.

4. Vertical wind shear

Our understanding of the processes by which severe convective events arise has come to include an awareness of the importance of the vertical wind shear structure in the convective storm environment. DMC arising in certain circumstances can produce quasi-stationary convective systems (Chappell 1986) that are common in heavy rainfall events. Supercell storms are a product of the interaction of the convective updrafts within a sheared environment (see, e.g., Rotunno and Klemp 1982; Weisman and Klemp 1986; Rotunno 1993; Brooks et al. 1994a,b). The vertical wind shear is not the only measure of character of the wind profile. Recently, some theoretical work (Davies-Jones 1984; Lilly 1986) has suggested a relatively new measure: the *helicity* of the wind profile. The ground-relative helicity, $\mathcal{H}$, is defined by

$$\mathcal{H}(z) = -\int_{z_o}^{z} \mathbf{k}\cdot\left(\mathbf{V}\times\frac{\partial \mathbf{V}}{\partial z}\right)dz\,, \qquad (5)$$

and is useful as a measure of the curvature of the hodograph. In a *storm-relative* framework, (5) becomes

$$\mathcal{H}_{sr}(z) = -\int_{z_o}^{z} \mathbf{k} \cdot \left[(\mathbf{V} - \mathbf{C}) \times \frac{\partial \mathbf{V}}{\partial z} \right] dz, \tag{6}$$

where **C** is the observed (or forecast) storm motion. Numerous studies (Davies-Jones and Brooks 1993; Davies and Johns 1993; Moller et al. 1994) have suggested that this parameter can be useful in assessing the potential for severe thunderstorms.

It should be noted that helicity is associated with winds changing direction with height, which is especially important when the framework is shifted to a storm-relative one. The Galilean coordinate transformation from one uniform velocity to another (as in going from a ground-relative to a storm-relative viewpoint) is of significance, because the physics of the atmosphere are invariant to such a transformation. Vector wind shear is indeed a Galilean invariant, whereas helicity is not. However, the use of diagnostic variables is not necessarily restricted to those that are Galilean invariant. The value of evaluating the threat of severe weather using storm-relative helicity is well-established (see Davies-Jones et al. 1990; Moller et al. 1994; Rasmussen and Blanchard 1998) in spite of whatever drawbacks might accrue from its violation of Galilean invariance. Since the storm motion is not known in advance of storm formation, this makes evaluation of storm-relative helicity problematic. A number of approximations can be used (see Bunkers et al. 2000) based on existing observations. From a purely theoretical viewpoint, however, storm motion can be viewed as implicit in the governing equations as a sort of "eigenvalue" (Davies-Jones 1998); it is not something entirely external to the problem of DMC. However difficult it may be to forecast, it still is physically relevant.

No matter how the character of the wind profile is evaluated, ETCs impose both temporal and spatial complexity on that vertical wind profile. In addition to processes in the PBL, various processes in the free atmosphere (both geostrophic and ageostrophic) in association with ETCs alter the winds aloft. Notably, thermal advection imposes backing and veering of the wind with height, and isallobaric accelerations associated with moving and evolving ETCs are important in modifying the wind profile. If the thermal wind ($\partial \mathbf{V}_g/\partial z$) is substituted into the integrand of (5) and its definition in terms of the thermal gradient is used, along with some vector identities, it can be shown that

$$\mathbf{k} \cdot \left(\mathbf{V} \times \frac{\partial \mathbf{V}}{\partial z} \right) = \frac{g}{fT} \mathbf{V} \cdot \nabla_z T \; . \qquad (7)$$

Thus, $\mathcal{H}$ is roughly proportional to the integrated thermal advection in the layer z_o to z in midlatitudes where QG approximations are reasonably good (as first demonstrated by Tudurí and Ramis 1997). Warm advection favors veering of the wind with height in the Northern Hemisphere, corresponding to positive ground-relative helicity, which is considered to be a favorable wind profile for the development of severe convection, as already noted. However, it is not the case that veering wind profiles are confined to regions of warm advection.

The importance of the low-level wind shear on severe forms of convection is beyond doubt (e.g., Weisman and Klemp 1982, 1984; Brooks et al. 1994b). However, it can be challenging to forecast the winds accurately. To the extent that an accurate wind profile forecast can be made, it becomes possible to anticipate the influence of synoptic systems on the character of any DMC that develops, especially the likelihood of supercells. Maddox (1993) has noted the potential for interesting diurnal variations of helicity due to PBL effects, but this is only a small part of the picture.

5. Organized convective systems and the low-level jet stream

Mesoscale convective systems are a particularly large and persistent class of DMC. As described by Maddox (1983), the largest and most persistent of these (MCCs) have some characteristics that appear to be tied to the synoptic structures in which they tend to occur. Rather than being associated with vigorous ETCs, these systems often arise in relative quiescent conditions in the middle and upper troposphere but with considerable warm thermal advection at low levels. This thermal advection is enhanced, at least in the plains of the United States, by diurnal variations in the LLJS, as noted in Maddox and Doswell (1982) and discussed earlier.

The structure and evolution of the LLJS in the United States does not take place independently of the synoptic structure, unlike the NBLWM. This has been demonstrated by, among others, Lanicci and Warner (1991) or Djuric and Ladwig (1983), who have shown that the LLJS undergoes a quasi-regular cycle associated with the passage of midlatitude cyclones and anticyclones. Further, Uccellini (1980) has shown that the LLJS is often coupled to upper-level jet streaks. Thus, the synoptic-scale structure is quite pertinent in the distribution and intensity of MCSs, and perhaps other aspects of severe DMC. The topic of MCSs is discussed in considerable detail in Ch. 9.

VII. Feedbacks to the synoptic scale

Most of the content of this chapter has been directed at revealing how the synoptic structure organizes the ingredients for DMC and modulates convective events. However, this addresses only half of the interaction. How does convection *feed back* to synoptic-scale processes? Uccellini (1990) concludes that "The role of convection in enhancing cyclogenesis ... remains unresolved." We concur with this assessment, but observe that evidence of upscale feedbacks certainly can be observed when DMC is widespread and persistent (see e.g., Stensrud 1996b, Dickinson et al. 1997). The climatological minimum of DMC in the winter (in midlatitudes) can be attributed to the enhanced vigor of cool season synoptic systems and the reduced heat input at low levels. In the warm season, synoptic systems have reduced baroclinity and, therefore, are correspondingly less intense. At the same time, the heat input at low levels is considerably larger in the warm season, so convection is needed more often than in the cool season. Convection and ETCs operate *together* to maintain the global heat balance.

In this context, we note that static stability is a physically relevant variable at virtually *all scales* in meteorology. Notably, if the atmosphere is too stable, cyclogenesis is inhibited since static stability measures the "resistance" of the atmosphere to forcing for vertical motions (Gates 1961) and, at the same time, DMC is inhibited. It is widely accepted that ETCs provide the means of alleviating baroclinity by poleward heat transport; their upward heat transport stabilizes the atmosphere as well. Both heat transports serve to dampen cyclogenesis. Tracton (1973), among others, has suggested that when deep convection is near the cyclone center, it can *enhance* cyclogenesis. As noted already, the evidence supporting a positive impact on ETCs is inconclusive.

The 13-14 March 1993 "Superstorm" over eastern North America mentioned in section II.2 also provides an excellent example of how convection feeds back to the synoptic scale (Bosart et al. 1996; Dickinson et al. 1997; Bosart 1999). A dynamical tropopause (DT) analysis revealed the presence of highly nonconservative PV behavior on the basis of an abrupt poleward shift of potential temperature contours on the DT over the southeastern US in the 24 h ending 0000 UTC 13 March 1993. This poleward shift, occurring downstream of a massive convection outbreak over the northwestern Gulf of Mexico (as mentioned earlier in section II.2), could not be explained by simple advection as would be the case if PV were conserved. This PV nonconservation strengthened the potential temperature gradient on the DT downstream of the deepening trough, indicative of a near-doubling of the jet strength in the upper troposphere over a 24 h period (see Dickinson et al. 1997; their Figs. 3 and 4 and Bosart 1999, his Fig. 9). The inability of the operational forecast models to simulate the explosive convection and its impact on the strength of the downstream jet resulted in significant errors in modeling the incipient cyclogenesis (Dickinson et al. 1997l their Figs. 14-22; Bosart 1999, his Fig. 9).

Convective modification of the larger-scale environment has also been noted in conjunction with relatively weak forcing aloft. Bosart and Lackmann(1995) documented the reintensification of tropical storm David (September1979) over land while it traversed the eastern slopes of the central and northern Appalachians. They showed that diabatic heating associated with deep convection on the eastern side of David resulted in amplification of a ridge aloft just to the east of the storm on the basis of a DT analysis. This ridge amplification created an area of positive PV advection over David ahead of a weak trough to the west that culminated in the reintensification of the tropical storm. The slope of the DT across David increased, with the DT higher to the northeast (Bosart and Lackmann 1995; their Figs. 11-14). In effect, the ability of the upstream trough to contribute to cyclogenesis by means of positive PV advection was enhanced by downstream ridging associated with widespread latent heat release from DMC associated with the tropical storm.

In the United States, the existence of strong synoptic scale systems in the spring, combined with increasing conditional instability created by surface heating while the middle and upper troposphere is still cool, produces the most intense and widespread severe convection. In the autumn, after a warm season's worth of DMC, the middle and upper troposphere are relatively warm; the asymmetry in conditional instability makes fall much less likely to produce the intense convective events. However, the autumnal DMC can be very effective at producing heavy precipitation. In midlatitude summer, synoptic-scale processes are weak and the heat input must be balanced *primarily* with convective transports (as in the Tropics).

As an interesting example of geographical variations on this theme, a relatively cool sea dominates the Mediterranean basin in the spring, so its most vigorous convective season tends to be in the autumn. After a summer's heating, the warm, moist boundary layer over the sea basin is the source of most of the region's severe convection. As it turns out, the predominant form of such severe convection in the Mediterranean basin is also heavy precipitation (as in North America during the fall).

Another variation can be found in Australia. A considerable portion of the continent is tropical or subtropical, but no barrier exists between the continent and Antarctica. The southern Hemispheric flow is predominantly zonal, perhaps as a result of fewer mountain barriers and much less landmass. A large, predominantly arid interior promotes deep boundary layers and steep lapse rates, so poleward flow into nontropical Australia is often unstable but dry. A mountain range close to the east coast limits the penetration of moisture into the interior in the prevailing low-level easterly flow. Thus, moisture can be the major limiting factor on severe convection during the spring and fall, when extratropical cyclones brush across the poleward half of the continent. Low-level moisture in available to synoptic-scale systems typically during landfall on Australia’s west coast and perhaps again as they exit the east coast.

In qualitative terms, DMC must have an important role to play in the global heat balance. The well-recognized importance of DMC in the Tropics is a direct consequence of the need for a near-equilibrium between convection and those processes leading to destabilization (Emanuel 1994; p. 479 ff.). As noted in Ch. 1 of this monograph, midlatitudes are not so well-characterized by a sort of quasi-equilibrium, at least on times scales of a day or less. DMC does not always develop in mid-latitudes as soon as destabilization begins; rather, it can be "stored" for hours or even days before DMC commences.

Although an individual convective cloud is too small and represents too brief a disturbance to have much *direct* impact on synoptic-scale processes, it is not uncommon for convection to develop into large mesoscale convective systems (MCSs), and in some situations, a succession of MCSs can exist for several days (Bosart and Sanders 1981; Fritsch et al. 1986). Stensrud (1996b) has shown that the accumulated effects of several days of persistent convection can have a significant impact on the synoptic structure. Whereas the long-term effect of convection must inevitably be an increase of static stability and a decrease in the vertical wind shear, the short-term effect on the environment sometimes can be to *maintain* conditions for DMC. In Stensrud's case study, an increased flow of low-level moisture (latent heat) into the convective area as a result of the convection allowed DMC to persist in spite of the slow stabilization of the lapse rates.

An interesting issue regarding feedback of convection to synoptic scales is the production of perturbations in the upper troposphere (and perhaps the lower stratosphere) from convection. Several studies (e.g., Ninomiya 1971; Fritsch and Maddox 1981; Maddox et al 1981; Maddox 1983) reveal that MCSs can produce significant perturbations in the meso-□ scale wind and height field aloft. Such structures do not usually persist for long beyond the demise of the convective system that produced them, but they do so occasionally. It is not known, in general, to what extent these perturbations influence the synoptic evolution downstream, subsequent to their development.

It is clear that at times, convectively-produced *mesoscale* features, notably the so-called mesoscale vorticity centers (MVCs; e.g., see Bartels and Maddox 1991), occasionally persist well beyond the demise of the convection that produced them. As with perturbations in the upper troposphere, these can persist and even be associated with the redevelopment of a convective system (e.g., Fritsch et al. 1994). MVCs appear to be a product of the preferential enhancement by the Coriolis effect of the cyclonic member of a couplet of vortices produced at the ends of convective lines (Davis and Weisman 1994). Jiang and Raymond (1995) have used a nonlinear balance model to produce a simulation of a convective system that develops a MVC, allowing exploration of such topics as the energy budget of the system.

VIII. Examples of severe convective events from a synoptic perspective

Rather than using only major outbreaks of severe weather to illustrate the synoptic-scale viewpoint, the idea here is to present some events as examples that are rather more subtle and certainly have proven to be more challenging to forecast. Severe convection in major outbreak situations during the climatological frequency maxima for severe convection (see Miller 1972; Doswell et al. 1993) is *relatively* easy to forecast and understand.[15] Many studies involving severe weather outbreaks (typically tornadic) of varying magnitude can be found (e.g., Fujita et al. 1970; Galway 1977; Benjamin and Carlson 1986). However, as noted in Ch. 1 of this monograph, presenting only cases that involve major outbreaks of severe DMC gives an inappropriate bias to the relationship between the synoptic weather pattern and severe DMC.

Severe convection can develop in many different synoptic environments. All such environments have in common the potential for DMC (see Johns and Doswell 1992), but specific severe weather events have different basic ingredients. For example, the conditions favoring strong downdrafts and, hence, potentially damaging convective outflow winds, need not be the same as the conditions favoring strong updrafts (which can favor, e.g., large hail). Thus, hazardous convection can be found in what might seem to be a bewildering array of synoptic-scale conditions. This makes generalizations difficult and leads to some concern for the value of individual event case

[15] However, not every "synoptically-evident" situation produces significant outbreaks of severe weather. Although major outbreaks of severe convection are often associated with vigorous ETCs, not every vigorous ETC results in severe convection.

studies, since to some extent every case is unique. Our approach here must necessarily be limited. Case studies are presented not because they exemplify the entire range of severe convection events, but because they illustrate the variety of synoptic structures that can be found with such events. The examples we have provided by no means exhaust that variety.

The format used for each short case study will not be identical. Space does not permit an extravagant presentation of the details for the case. Rather, the intent is to provide a few key figures that illustrate the main points associated with each case. All but the last of our cases have been documented to a greater or lesser extent in publications, either in the refereed literature or in NOAA Disaster Surveys. Interested readers can consult the documentation for more details.

1. Tornadic systems

a. 28 Aug 1990 -- Plainfield, IL, an isolated tornado event

Johns (1982) has documented many aspects of severe convection associated with midlevel flows exhibiting an equatorward component, rather than the more well-recognized situations involving flows having a poleward component in midtroposphere. This devastating tornado developed in such a case, striking Plainfield at about 1815 UTC and lasted about 30 min; it was rated F5 on the Fujita scale. A survey of the event can be found in the National Weather Service (1991a) and was discussed by Fujita (1993) and Seimon (1993). As shown in Fig. 15, the event occurs on the margins of a synoptic-scale ridge in mid-troposphere, just ahead of a weak short-wave trough. However, the region is characterized by extreme lapse rates in the lower mid-troposphere, giving rise to very large CAPE (Fig. 16), with estimates ranging from nearly 6000 J kg^{-1} to around 8000 J kg^{-1}, depending on which parcel is chosen for ascent (Doswell and Brooks 1993a). CAPE values exceeding 3000 J kg^{-1} are uncommon, as documented by Doswell and Rasmussen (1994). This CAPE clearly has arisen because of convective inhibition, preventing the destruction of the high lapse rates and allowing them to be carried relatively far east of the west-central plains severe weather maximum.

The Plainfield event represents a challenge to our current concepts of environmental influences on tornadogenesis, in that the tornado was of "violent" intensity but occurred in what appears to be an environment with relatively weak wind shear and helicity (see Korotky et al. 1993; Doswell and Brooks 1993a). As noted by Brooks et al. (1994b), the actual storm environment in any given situation is always open to question, especially in cases that are not "synoptically-evident." Further, it is clear that we have so few examples of extreme CAPE-associated supercells, that our understanding of convective storm dynamics is not adequate to explain every event.

b. <u>27 Mar 94 -- Palm Sunday 1994 tornado outbreak</u>

This event was the subject of an entire session at the 18th American Meteorological Society's Severe Local Storms Conference (e.g., see Hales and Vescio 1996). It has been characterized as not exhibiting the "classic" features of an outbreak; notably, a prominent mid-tropospheric cyclone was not observed. Nevertheless, the National Weather Service (1994) concludes that synoptic-scale processes were important in creating an environment favorable for the outbreak. During the time leading up to this event, the synoptic evolution created a widespread region with considerable CAPE and substantial vertical wind shear (Fig. 17). This evolution took place without involving strong surface cyclogenesis (Fig. 18). Within the broad region of favorable instability and wind shear, mesoscale processes played a large role in developing and organizing convection (see Kaplan et al. 1998; Langmaid and Riordan 1998). The presence of a jet stream with embedded subsynoptic-scale windspeed maxima certainly may have been an important factor in initiating the outbreak of tornadic storms via transverse circulations (Kaplan et al. 1998).

There were two tornadic storms responsible for most of the damage and casualties. One supercell developed out of thunderstorms that began in eastern Mississippi before 1500 UTC, and the storm tracked across Alabama, producing four tornadoes, dissipating as it moved into Georgia after 1800 UTC. The second supercell evolved from thunderstorms in west-central Alabama before 1500 UTC, remaining south of the other major supercell. This second storm also produced at least four long-track tornadoes as it moved across Alabama and on into Georgia, dissipating in northwestern South Carolina after 1800 UTC.

Although it is not uncommon for mesoscale processes to be important, outbreaks are associated classically with major cyclones (e.g., Fujita et al. 1970). The absence of such a prominent cyclone in this case makes it clear that the ingredients for a significant outbreak of severe weather can be assembled in many different synoptic-scale structures. The fact that it is *typical* for outbreaks to occur in association with cyclogenesis does not mean that the absence of cyclogenesis precludes major outbreaks of tornadic severe convection.

2. Nontornadic systems

a. 17 Aug 94 -- Lahoma, OK, an isolated intense wind and hail storm

This storm also was the topic for most of an entire (poster) session at the 18th American Meteorological Society's Severe Local Storms Conference. Janish et al. (1996) note that it was not well-anticipated, perhaps owing to the relatively quiescent appearance of the synoptic environment (Figs. 19-20). The most obvious feature in mid-troposphere is the approaching thermal trough (Fig. 19), which is exaggerated by using an isotherm interval of 2 C. It apparently was produced by a relatively rare form of supercell storm first documented by Brooks and Doswell (1993) that is notable for the intensity and duration of strong convective outflows over a broad swath. Based on a limited sample, Brooks and Doswell conclude that these events are characterized by: extreme CAPE, low mid-tropospheric relative humidity, and significant storm-relative helicity in the 0-3 km layer, factors shared with many other supercells. However, they surmise that the factor leading to the dominance of strong outflow is relatively weak mid-tropospheric storm-relative flow. In their conceptual model, this weakness in storm-relative flow suggests that mid-tropospheric mesocyclones would tend to be filled with precipitation, rather than having most of the precipitation carried downstream to fall in the storm's forward flank precipitation cascade. This case seems to fit the conceptual model, with high CAPE (not shown) and weak storm-relative windflow in mid-troposphere (Fig. 21).[16] It also provides a compelling example of the potentially devastating character of such non-tornadic severe convective wind events, in spite of a seemingly innocuous synoptic-scale setting.

This event, which became severe soon after 1900 UTC and ended at about 2100 UTC, produced a swath of large hail (up to 12 cm) and strong winds (gusts exceeding 50 m s^{-1} were reported) that was 6-12 km wide over a path about 90 km long. This represents an effected area approaching 1000 km^2, which means that should such an event strike an urban area, the potential for substantial, non-tornadic damage exists. Apparently, events of this magnitude are relatively rare; over all of North America, Brooks and Doswell (1993) estimate an occurrence rate of perhaps a few times per year, at most.

b. 07 Jun 82 -- Kansas City, MO, a derecho

Johns and Hirt (1987) were the first to document the widespread windstorm event now known as a derecho. This case, in which severe weather began (in northwestern Kansas) at about 0530 UTC on 7 June and continued until after 1400 UTC, has been presented in detail by Rockwood and Maddox (1988) as an example of the complex interactions that lead to convective events. It also illustrates the importance of a detailed analysis in some situations, since many features of significance (e.g., as illustrated in Fig. 22) are not easily detected in a smoothed analysis, typical of conventional synoptic-scale analyses done at a national center.

The synoptic setting cannot be described as benign, but neither can it be said to be clearly portentious of the event to come. The features of importance are not readily diagnosed using the operational objective analyses. Although the Q-vector divergence (Fig. 23) implied synoptic-scale

[16] The veering of the wind through a deep layer also suggests deep warm thermal advection, which is associated with ascent, quasigeostrophically.

descent in the area where storms first developed (in extreme southwestern Nebraska),[17] the thermodynamic conditions indicated considerable potential for strong storms, with CAPE of about 3000 J kg^{-1} (Fig. 24) if convection could be initiated. In this case, Rockwood and Maddox show that the initiation of convection required a complex sequence of mesoscale events that would be difficult to anticipate. Once initiated, however, an important convective event ensued; the widespread damage in Kansas City associated with the storm disrupted activities for several days, and there were numerous injuries.

3. Heavy precipitation systems

a. <u>14 Jun 90 -- Shadyside, OH, a challenge</u>

As a measure of the importance of heavy rainfall as a severe convective event, this flash flood produced more than 25 fatalities (National Weather Service 1991b). Although the National Weather Service had issued a flash flood watch for the area at 1941 EDT (2341 UTC), no flash flood *warning* was issued for the event. It is typical of many flash flood events in that it occurred with the passage of a weak short-wave trough in mid-troposphere (Fig. 25) on the margins of a large anticyclone. Very high surface moisture was evident on the poleward side of a non-frontal convergence boundary (Fig. 26), that originated as an outflow from previous convection. Considerable rainfall preceded the Shadyside flash flood; indeed, the preceding May had well above average precipitation across the state of Ohio. This sort of situation is described reasonably as a "mesohigh" event (using the conceptual model developed by Maddox et al. 1979), but it is certainly not a perfect fit to that pattern.

Rainfall began over the basins that flooded at about 2345 UTC and lasted for about 2 hr. The convection was quasistationary over two small catchments with steep, rocky terrain (although there was no history of flash flooding in the catchments). The rainfalls were not measured by any

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[17] Note that inferring vertical motion from the rhs of the Omega Equation (Eqn. 2) can be perilous. In this case, the pattern had considerable vertical coherence and is correspondingly likely to represent the vertical motion pattern reasonably well.

raingage, but estimated peak amounts were in the range of 75-100 mm, most of which fell in about one hour. Such peak rainfalls are by no means remarkable in themselves; the hydrological setting (steep terrain, antecedent rainfalls) clearly played a large role in this event's societal impact, as it often does. The available radar data (the WSR-88D radars were not operational for this case) in real time proved an unreliable tool for detecting the event and, in spite of favorable synoptic conditions, the magnitude of the event went undetected. In spite of having relatively low flash flood guidance values of about 50 mm in 3 hr, there was no obvious indication from the available data that such an event was underway. Regrettably, it took about 4 hr for information about the flooding at Shadyside to reach the National Weather Service office.

b. <u>09 June 72 -- Rapid City, SD, a western event</u>

The circumstances of a dominating midtropospheric anticyclone with a weak short-wave trough (Fig. 27) are found again for this event. Dennis et al. (1973), Maddox et al. (1978), and Doswell et al. (1996) have documented the event, which killed more than 230 people. In this case, the orography of the region seems to have been an important factor in focusing the convective development. At low levels, easterly flow behind a surface front (Fig. 28) was bringing high moisture values upslope, in the presence of strong lower mid-tropospheric lapse rates (Fig. 29a) at Rapid City. By the evening of the event, the arrival of the moist easterly flow at low levels and the increase in tropospheric humidity are quite evident (Fig. 29b). The arrival of this moisture, combined with orographic lifting and the approach of a slow-moving shortwave trough in midtroposphere, resulted in strong thunderstorms, beginning about 0000 UTC. The thunderstorms were moving relatively slowly, apparently owing to the weak winds throughout the middle and upper troposphere on this day, as well as the importance of orographic lifting. The presence of a weak, negatively-tilted midtropospheric ridge apparently favored the slow movement of the storms and the northward track of the weak shortwave trough. Peak 24 hr rainfall amounts are estimated to be about 380 mm, most of which fell within about 4 hr on the steep slopes of the Black Hills to the immediate west of Rapid City.

Maddox et al. 1978) specifically noted many similarities of the Rapid City case to another flash flood event, in the Big Thompson Canyon of Colorado four years later. The Big Thompson flash flood killed more than 140 people, thus having the dubious distinction of being the last flash flood in the United States to produce more than 100 fatalities.

4. *A cold season severe weather case*

On 21-22 January 1999, a substantial outbreak of severe thunderstorms and tornadoes (Fig. 30) developed, including several tornadoes rated F3 on the Fujita scale. The first severe weather began about 2000 UTC on 21 January, diminishing after 0430 UTC on 22 January. What makes this event interesting is the *season* in which it occurred; tornado outbreaks in the depths of winter are certainly not common. Although vigorous extratropical cyclones (Fig. 31c) occur routinely during the winter, they don't typically include both relatively high mid-tropospheric lapse rates (Fig. 31b) and substantial low-level moisture (Fig. 31a). An indication of the effect of the synoptic-scale ascent can be seen in the 6-hr evolution of the sounding at Shreveport, LA (Fig. 32). Clearly, there was considerable erosion of what was originally a strong capping inversion, resulting in widespread severe convection.

The pattern shown in Fig. 31 includes a negatively-tilted trough in midtroposphere, with a strong jet stream upstream of the trough axis and difluent contours ahead of the trough. This is by no means a necessary structure for severe weather outbreaks, but it may not be coincidental that this pattern is similar to that of 3-4 April 1974, the infamous "Superoutbreak" described by Fujita (1974). Such a pattern favors ascent on the cyclonic shear side of the jet exit region, and can put strong flow aloft over the warm sector of the developing surface cyclone, producing favorable wind shears. When the right lapse rate and moisture conditions can be created within a major winter storm, outbreaks of severe convection (see Galway and Pearson 1979; Kocin et al. 1995) can occur.

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FIGURE CAPTIONS

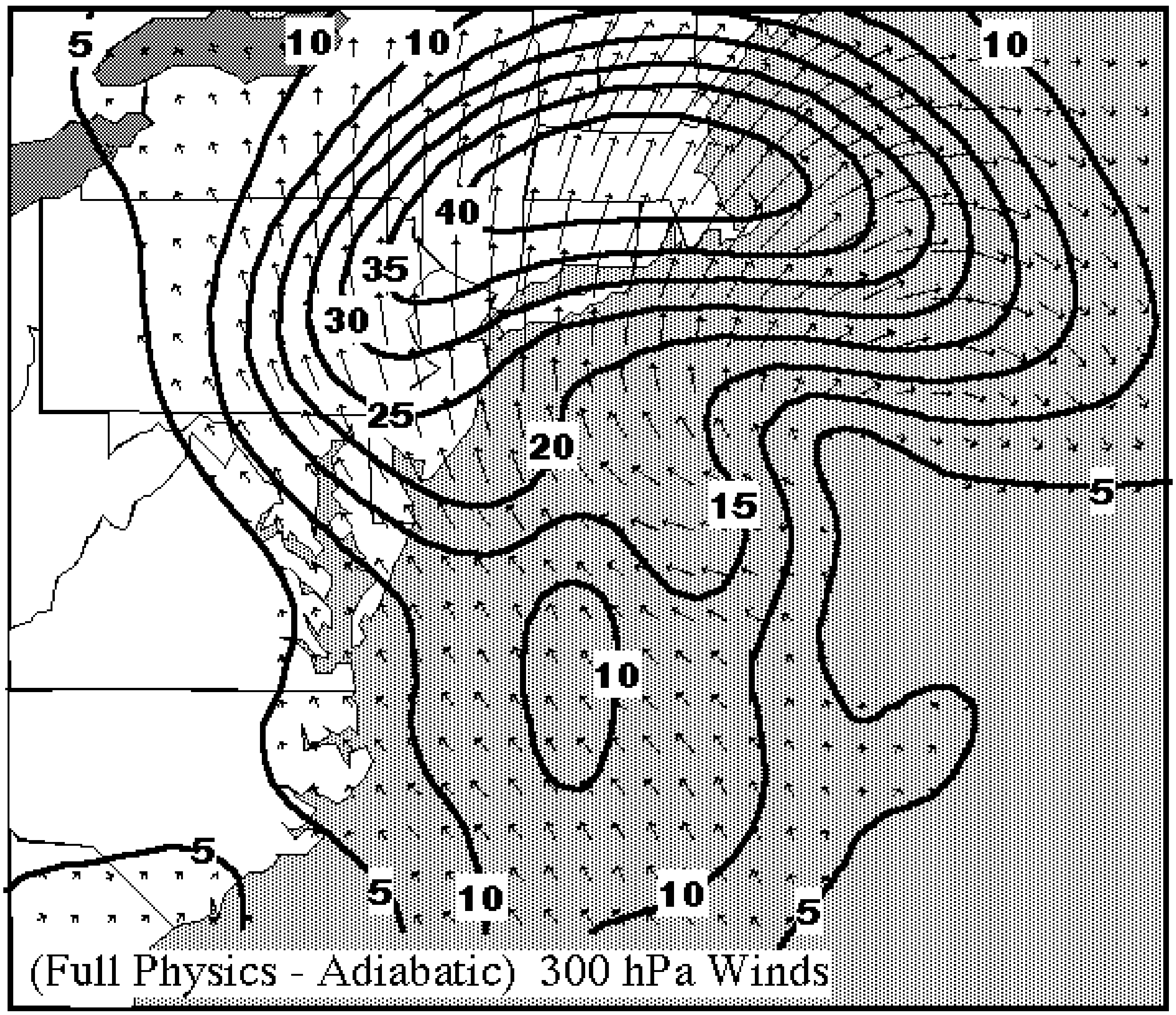


Figure 1. 300 hPa wind differences associated with subtracting an adiabatic numerical model run from a "full physics" version that includes parameterized convection, valid at 1200 UTC on 19 February 1979. Contours show the isotachs of the wind differences in m s^{-1}, while the arrows show the direction and the speed; the boldface "L" depicts the full-physics run surface low pressure center location. The figure is based on data supplied by L. Uccellini for the case described in Whitaker et al. (1988).

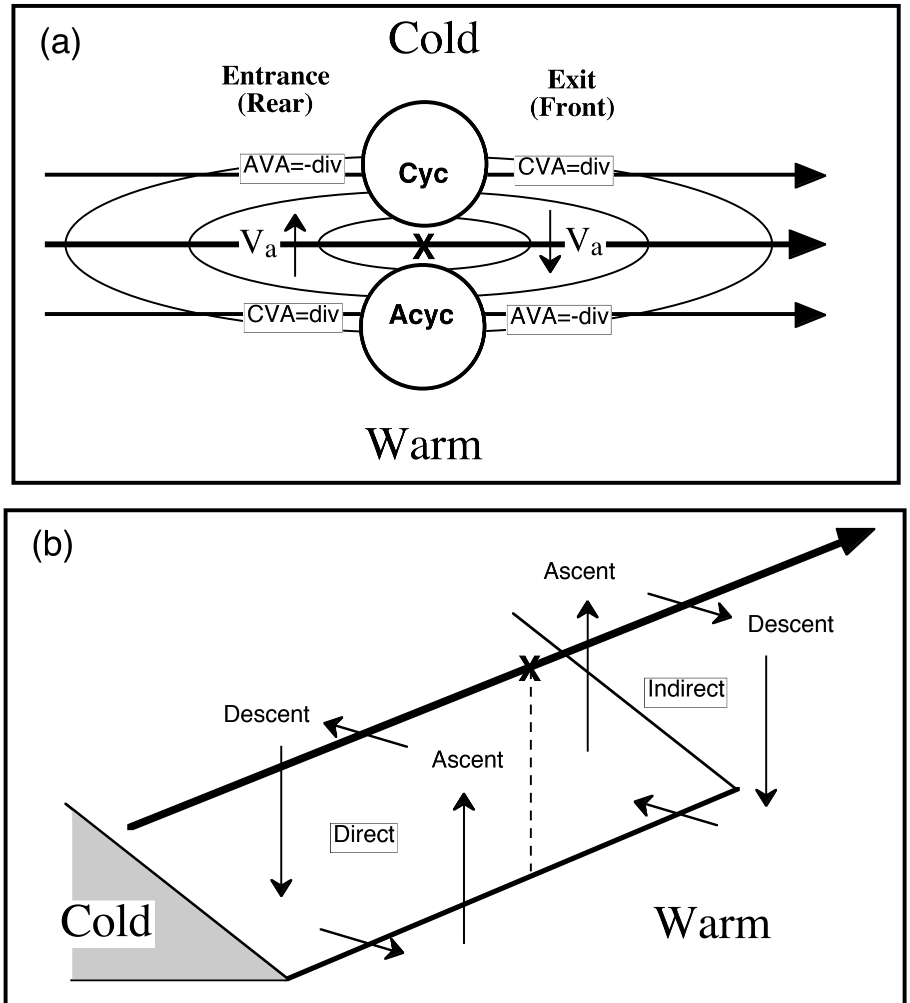


Figure 2. Schematic model of a Northern Hemispheric jet streak in straight flow; cold air is indicated to the left of the jet streak and warm air to the right. In (a), the isotachs (thin solid lines) depict the jet streak; the thick solid lines with arrows indicate the flow, an "X" marks the windspeed maximum. The circle marked "Cyc" shows a cyclonic vorticity maximum and that marked "Acyc" is a comparable anticyclonic vorticity maximum. Cyclonic vorticity advection is located by "CVA" and "AVA" locates anticyclonic vorticity advection, while "div" stands for divergence and "-div" for convergence. The vectors labeled $\mathbf{V}_a$ are the transverse ageostrophic wind vectors at the height of the jet streak, implied by the along-stream variations in the wind speed. In (b), the conceptual circulations are depicted in 3 dimensions. The vertical circulation [including the ageostrophic cross-jet flows, as in (a)] in the exit region is indirect (warm air sinking and cold air rising), while that in the entrance region is direct (warm air rising and cold air sinking). The "wedge" of cold air is indicated schematically, with its leading edge under the jet axis.

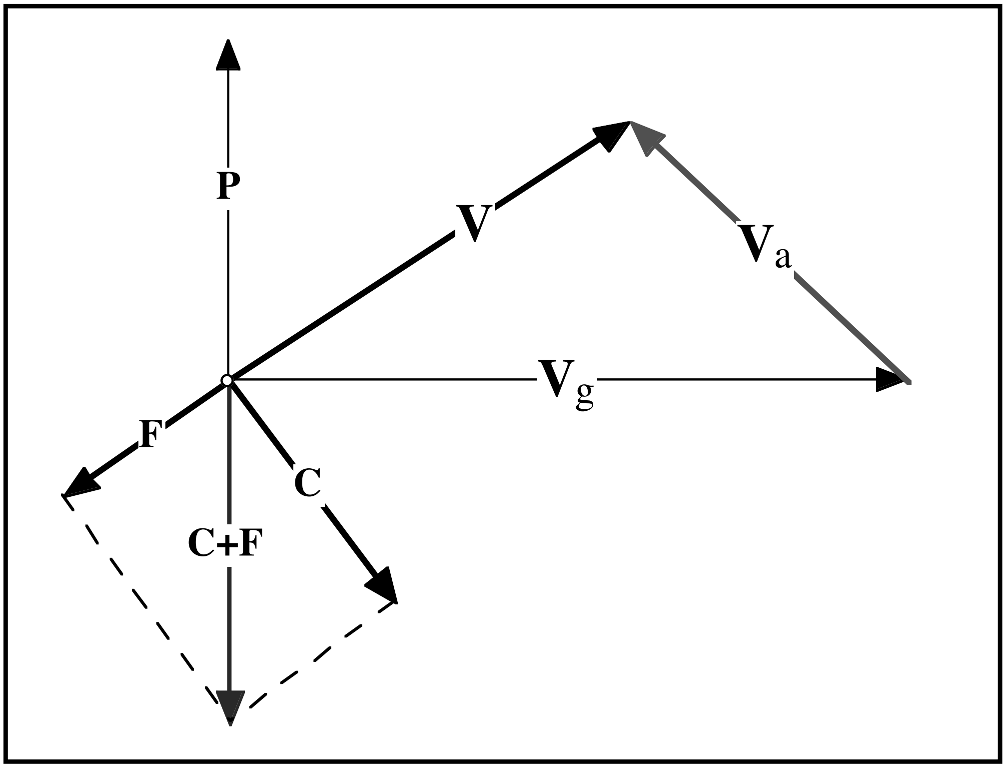


Figure 3. Force balance for a wind **V**, including the Coriolis force **C**, pressure gradient force **P**, and a simple friction force (opposite to the wind direction) **F**. $\mathbf{V}_g$ is the geostrophic wind and $\mathbf{V}_a$ is the ageostrophic wind.

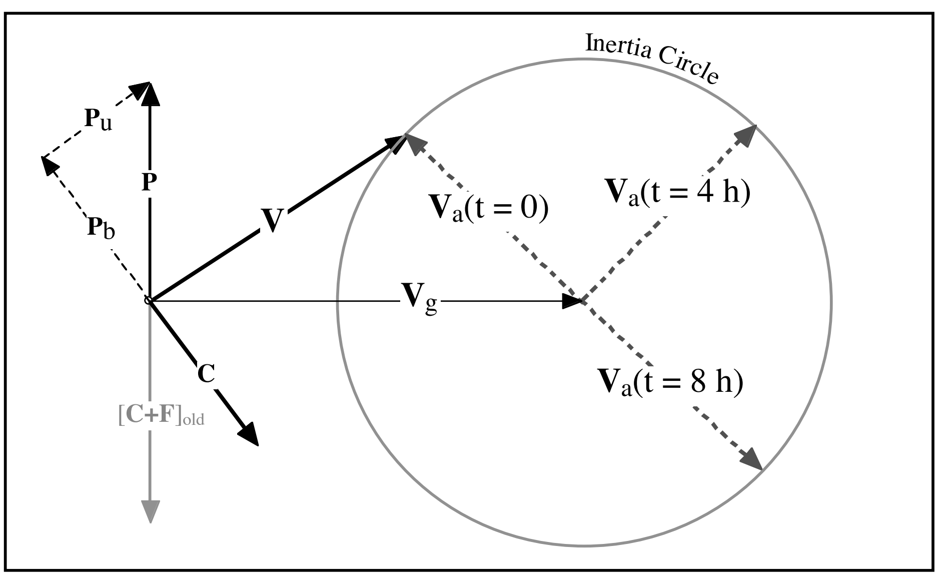


Figure 4. The force balance immediately after "switching off" the friction force, showing that the pressure gradient force **P** includes a component $\mathbf{P}_b$ in balance with the Coriolis force **C** and a component $\mathbf{P}_u$ that is unbalanced; **F** is the friction force vector. Also shown are the "old" vector sum of **C** and **F** (as in Fig. 3) that has vanished when the friction was turned off, as well as the inertia circle, showing how the ageostrophic wind $\mathbf{V}_a$ rotates anticyclonically about the geostrophic wind $\mathbf{V}_g$ at selected times (labeled).

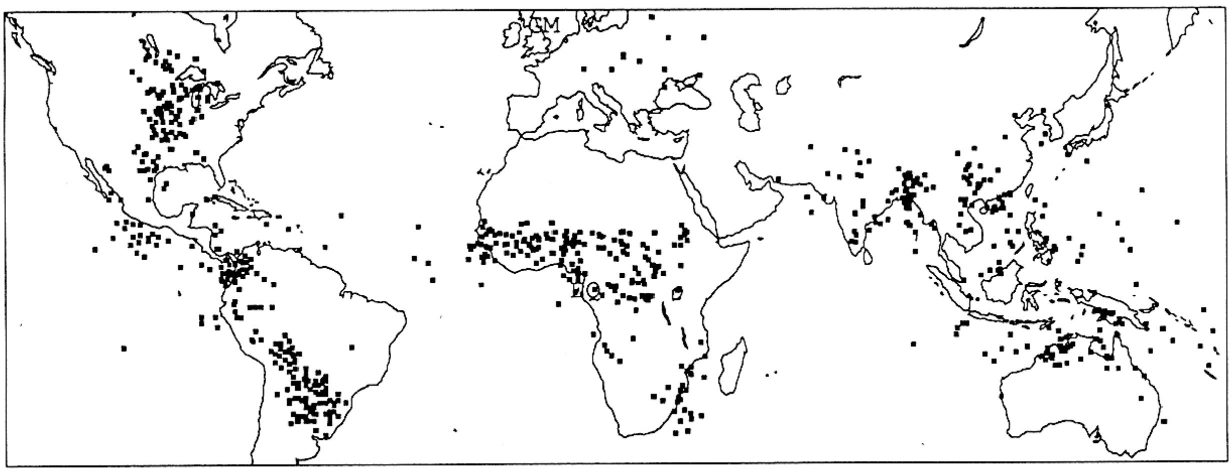


Figure 5. The global distribution of mesoscale convective complexes (MCCs), based on satellite imagery as described in Laing and Fritsch (1997). Small squares indicate location of MCC at time of maximum extent; adapted from Laing and Fritsch (1997).

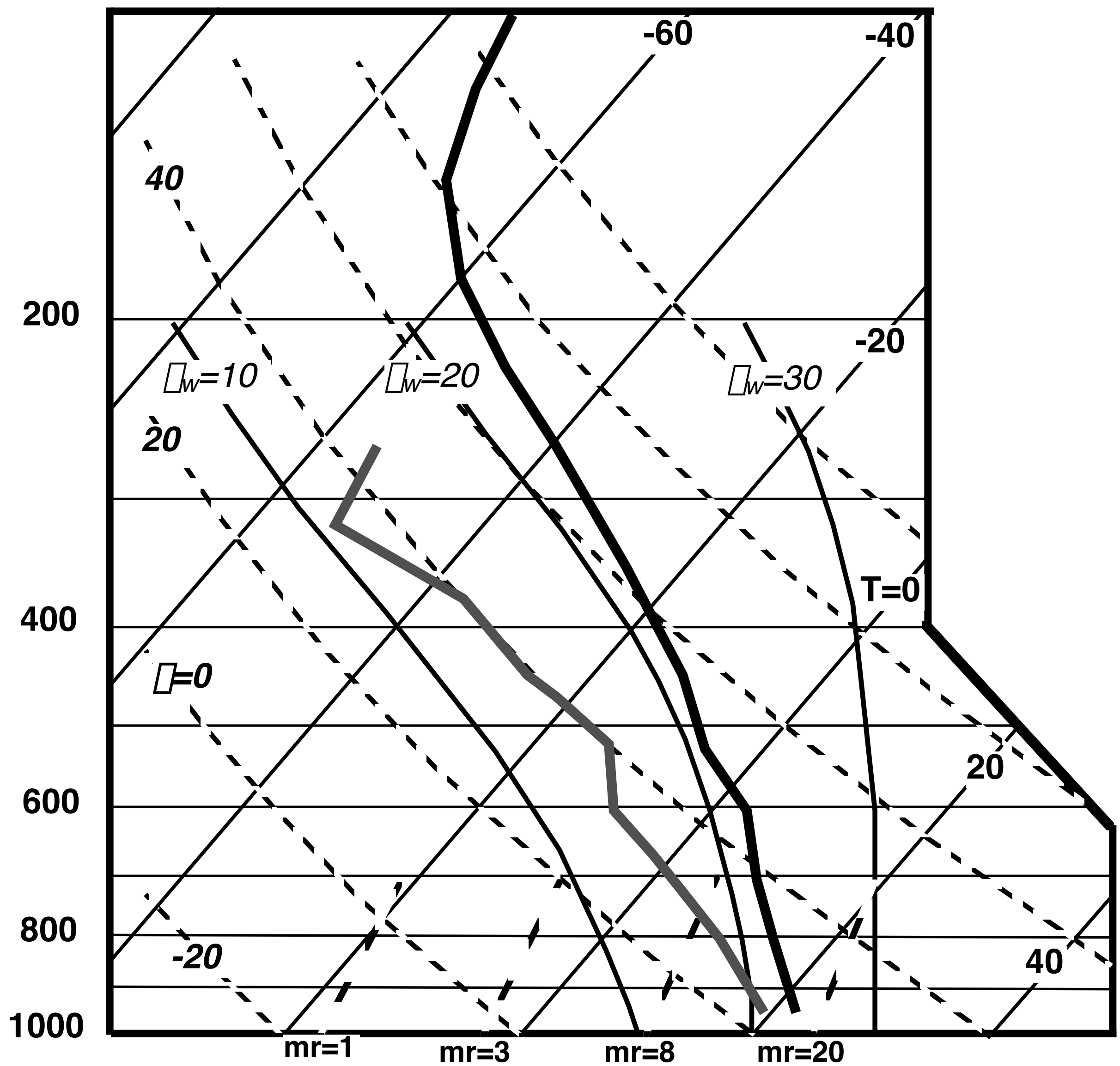


Figure 6. C.L. Jordan's (1958) mean West Indies tropical sounding plotted on a skew T, log p diagram.

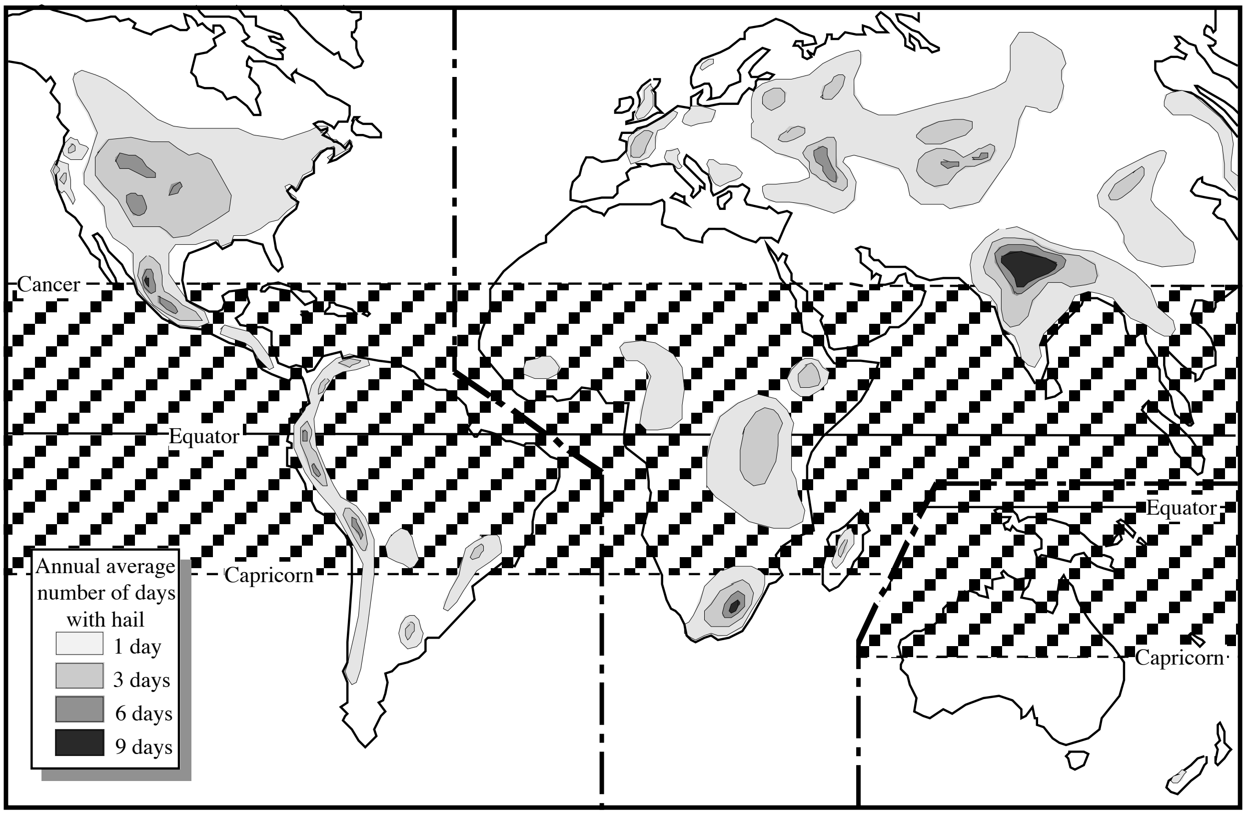


Figure 7. Depiction of the global frequency of hail, after Frisby and Sansom (1967).

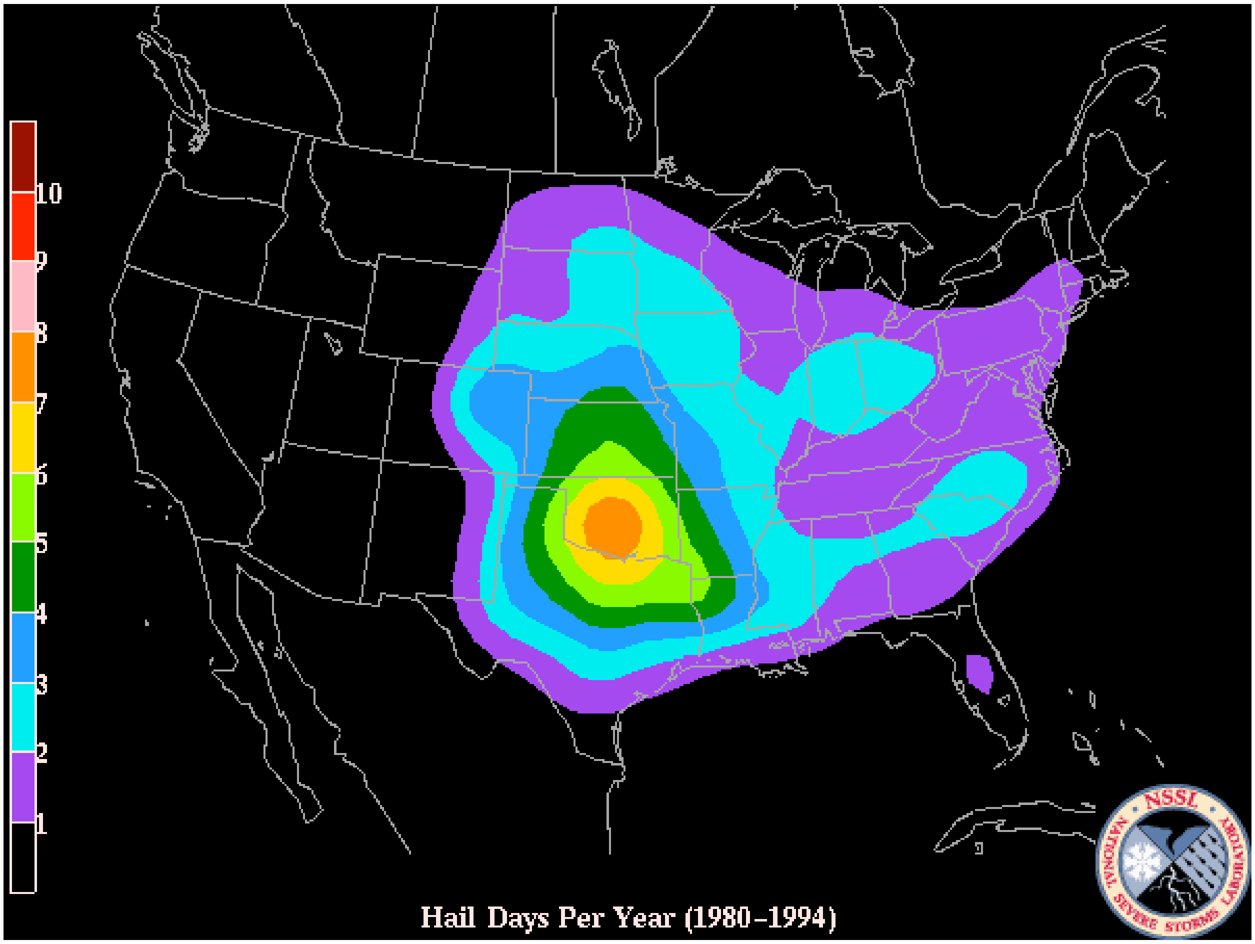


Figure 8. Annual frequency (see the key) of hail reports with diameters ≥ 3/4 in (2 cm) within an 80 km grid square, based on data for the period 1980-1994.

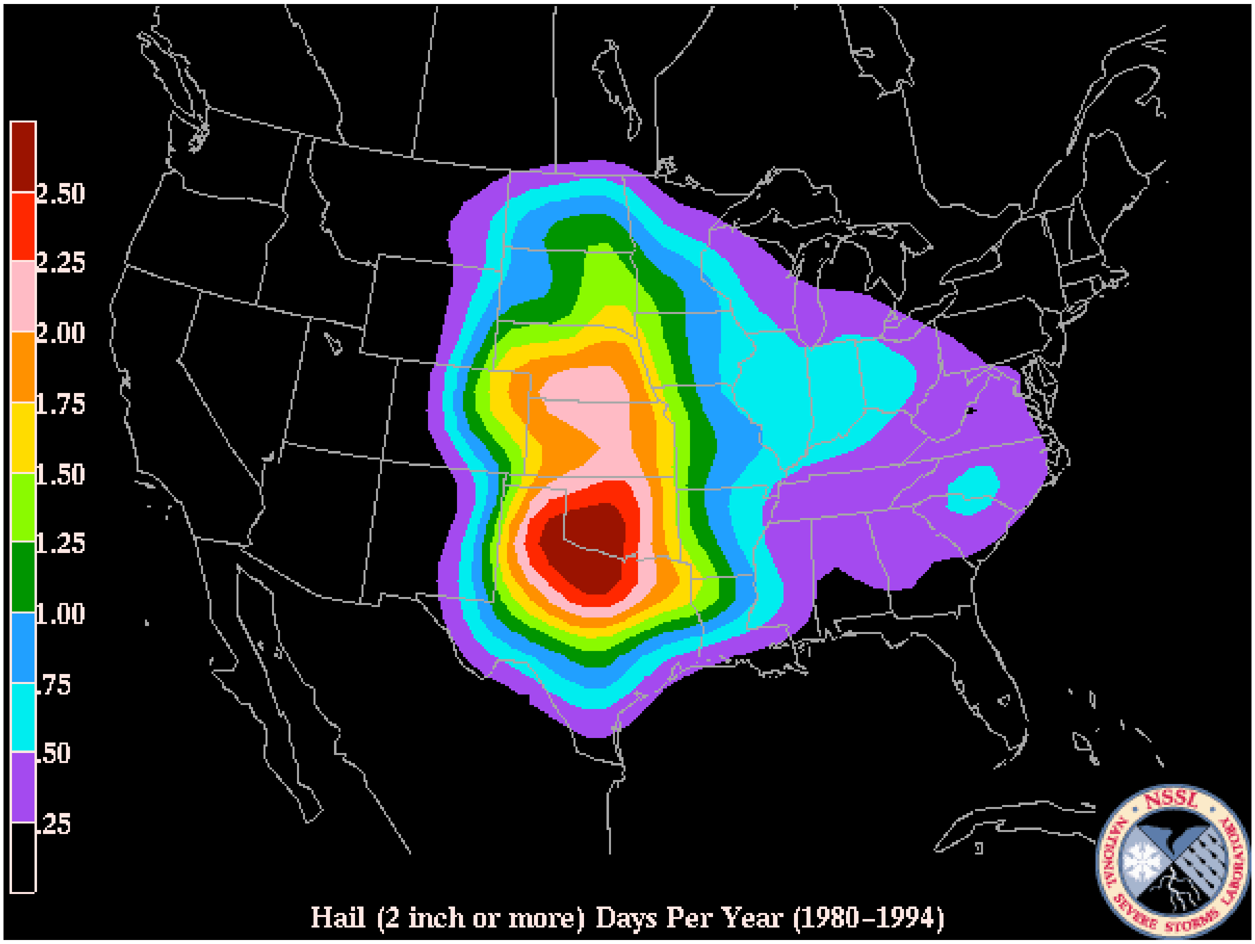


Figure 9. As in Fig. 8, except for diameters ≥ 2 in (5 cm).

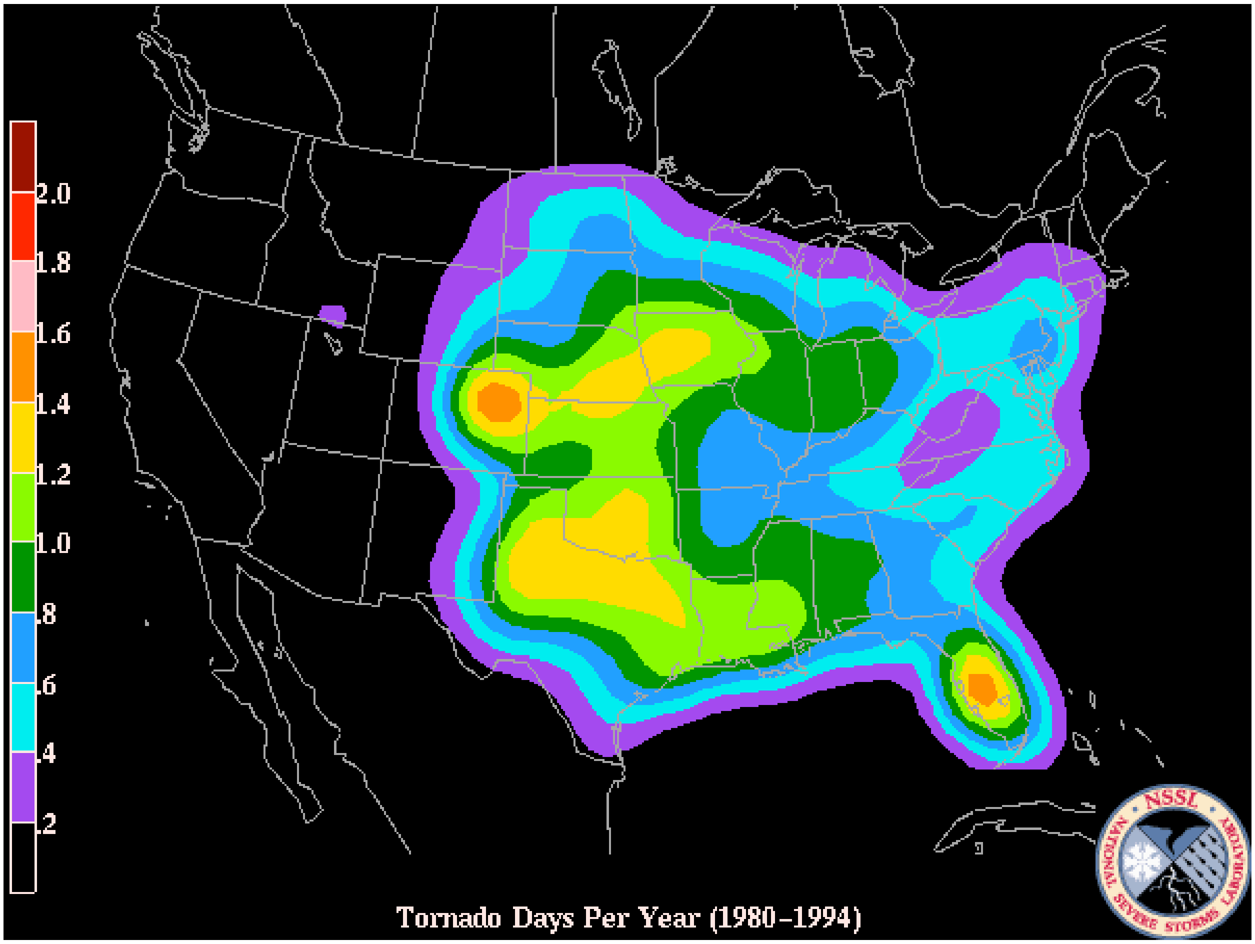


Figure 10. As in Fig. 8, except for any tornado report.

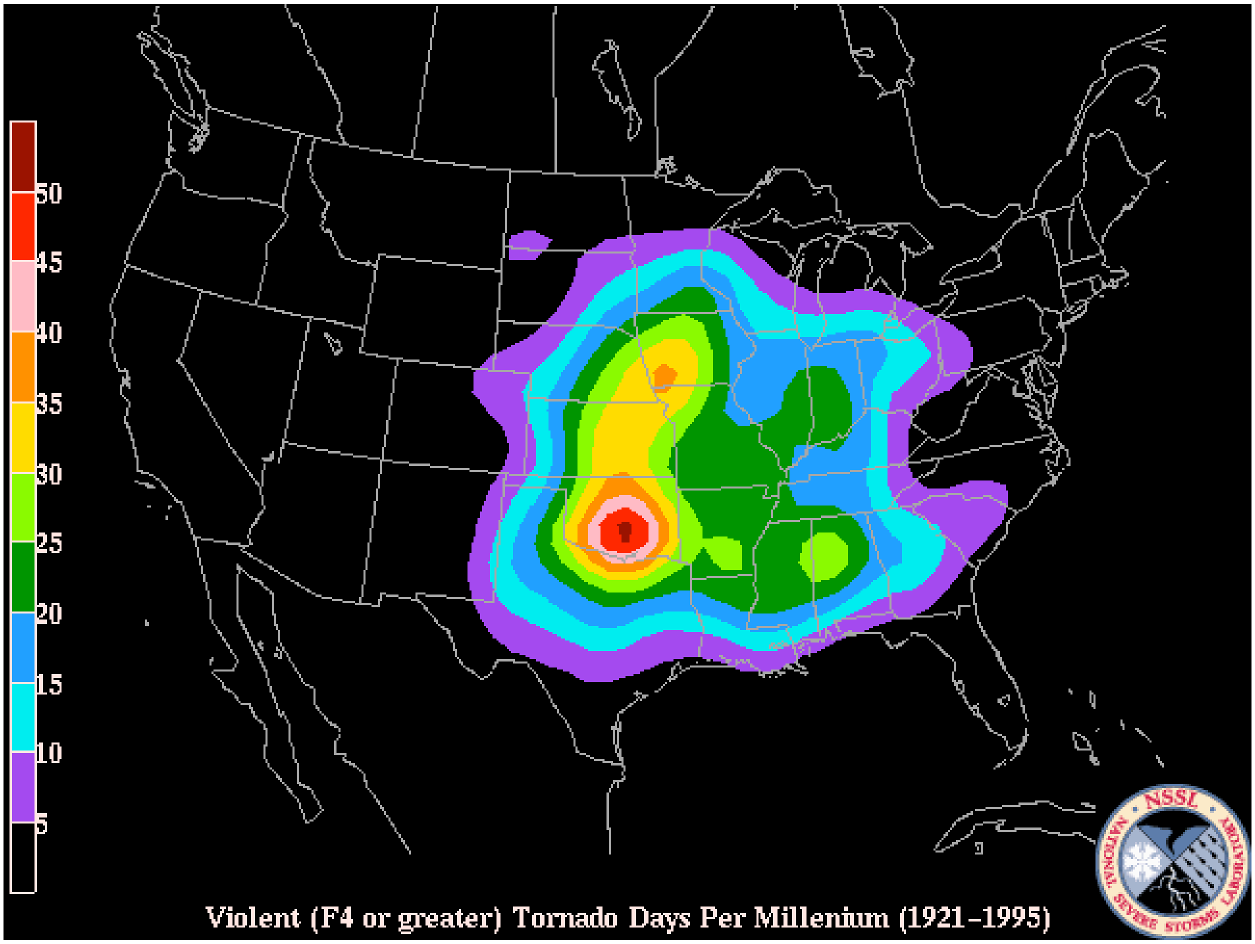


Figure 11. Frequency of violent tornadoes [F4-F5 on the Fujita (1971) scale)] normalized to per thousand years, based on data for the period 1921-1995.

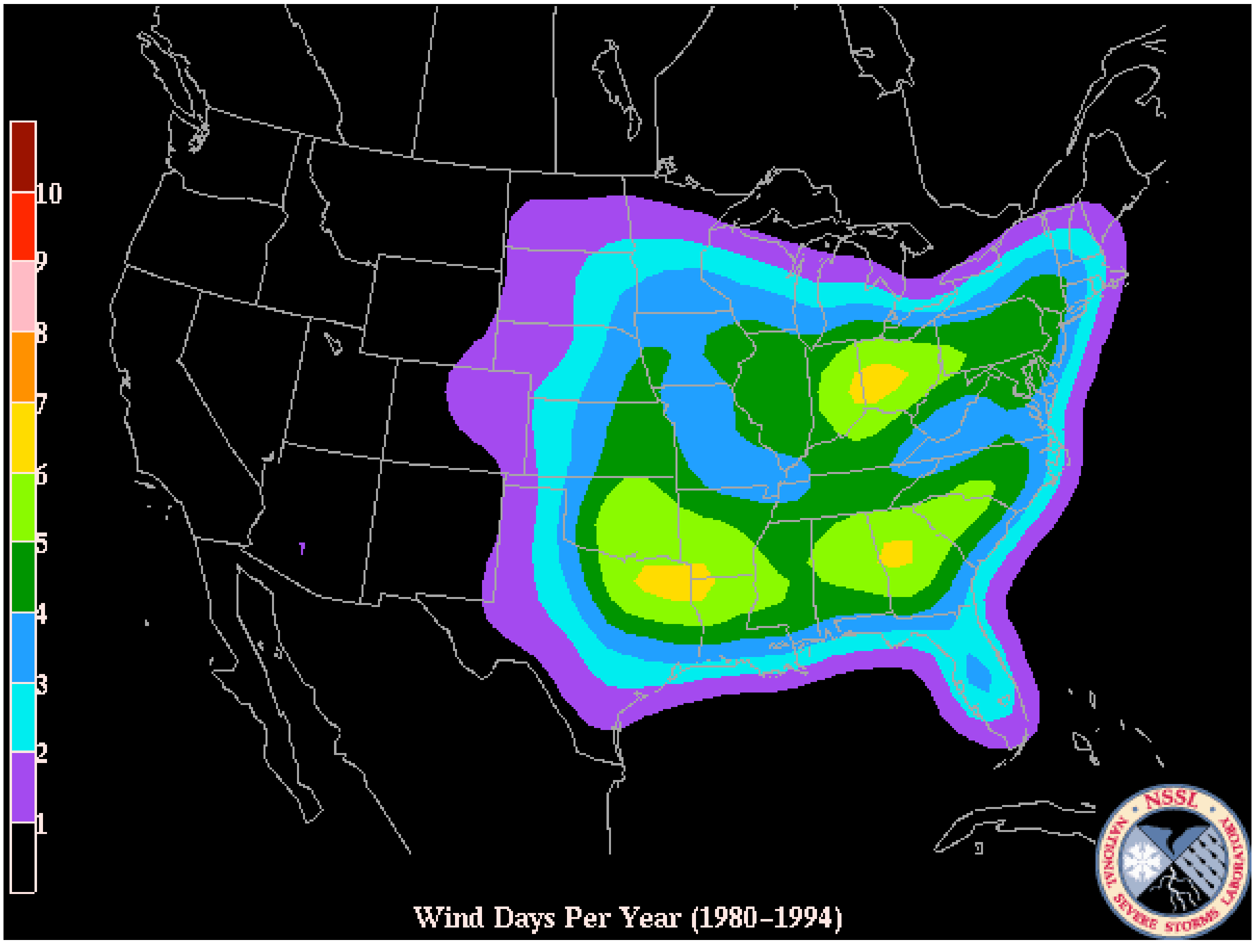


Figure 12. As in Fig. 8, except for any convective wind gust report.

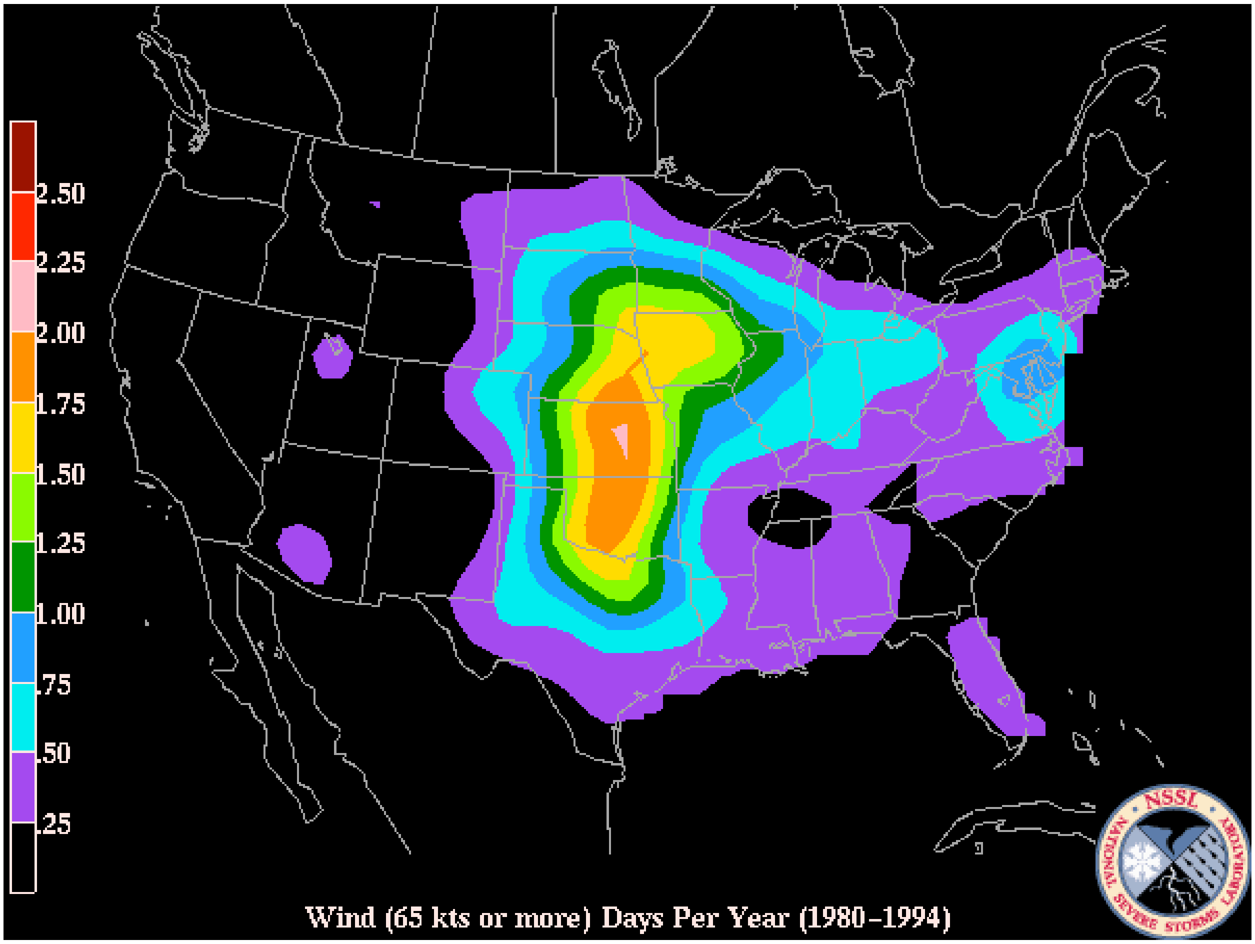


Figure 13. As in Fig. 8, except for wind gusts exceeding 65 kt (32 m s^{-1}).

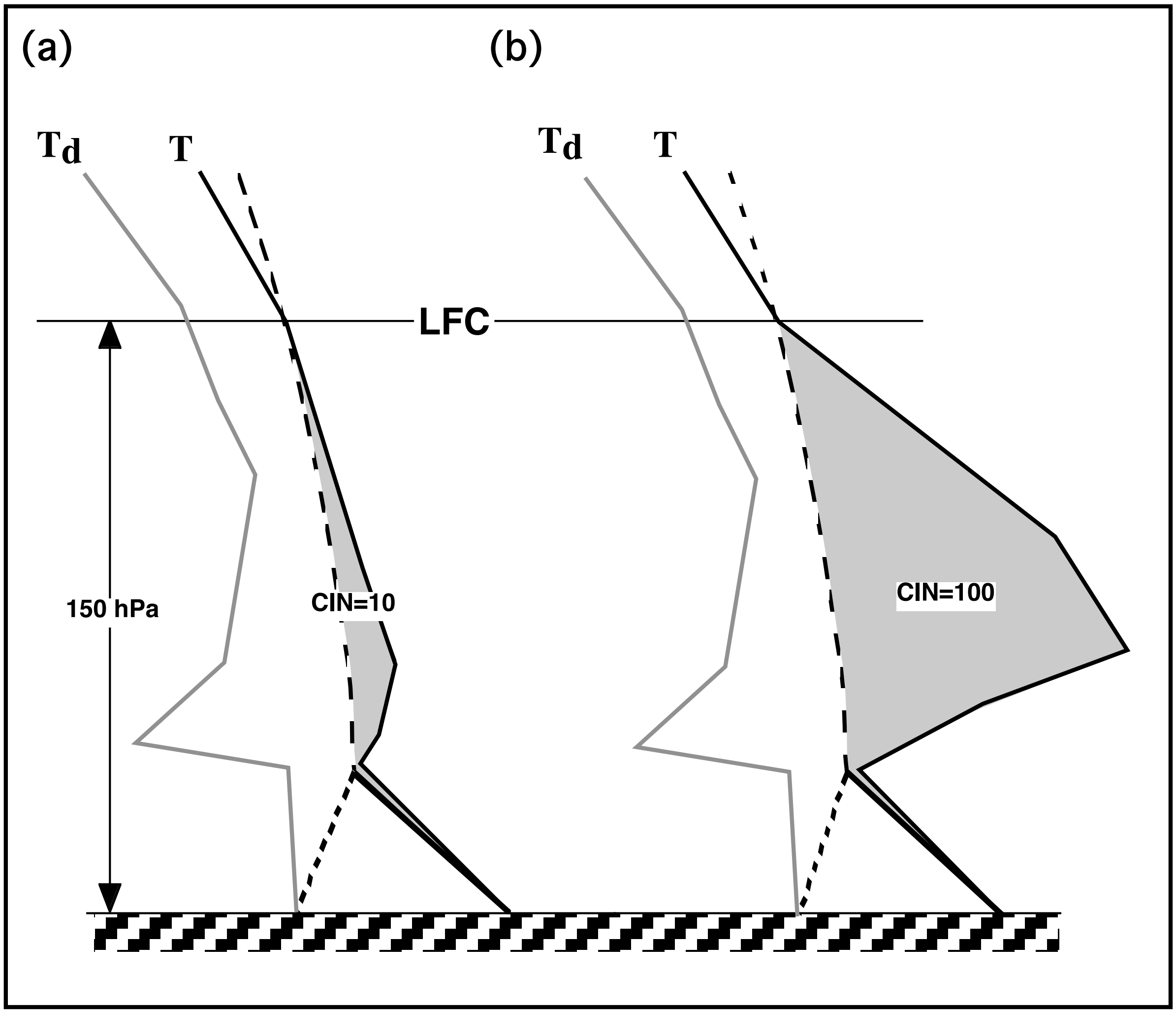


Figure 14. Schematic comparing two soundings, each with a Level of Free Convection (LFC) 150 hPa above the surface. The trajectory followed by the surface parcel is suggested with the solid, dashed, and hatched lines that show the dry adiabatic ascent, the moist adiabatic ascent, and the mixing ratio of the surface parcel, respectively. The environmental temperature (T) and dewpoint temperature (T_d) soundings are shown, and the associated CIN is indicated by the stippling.

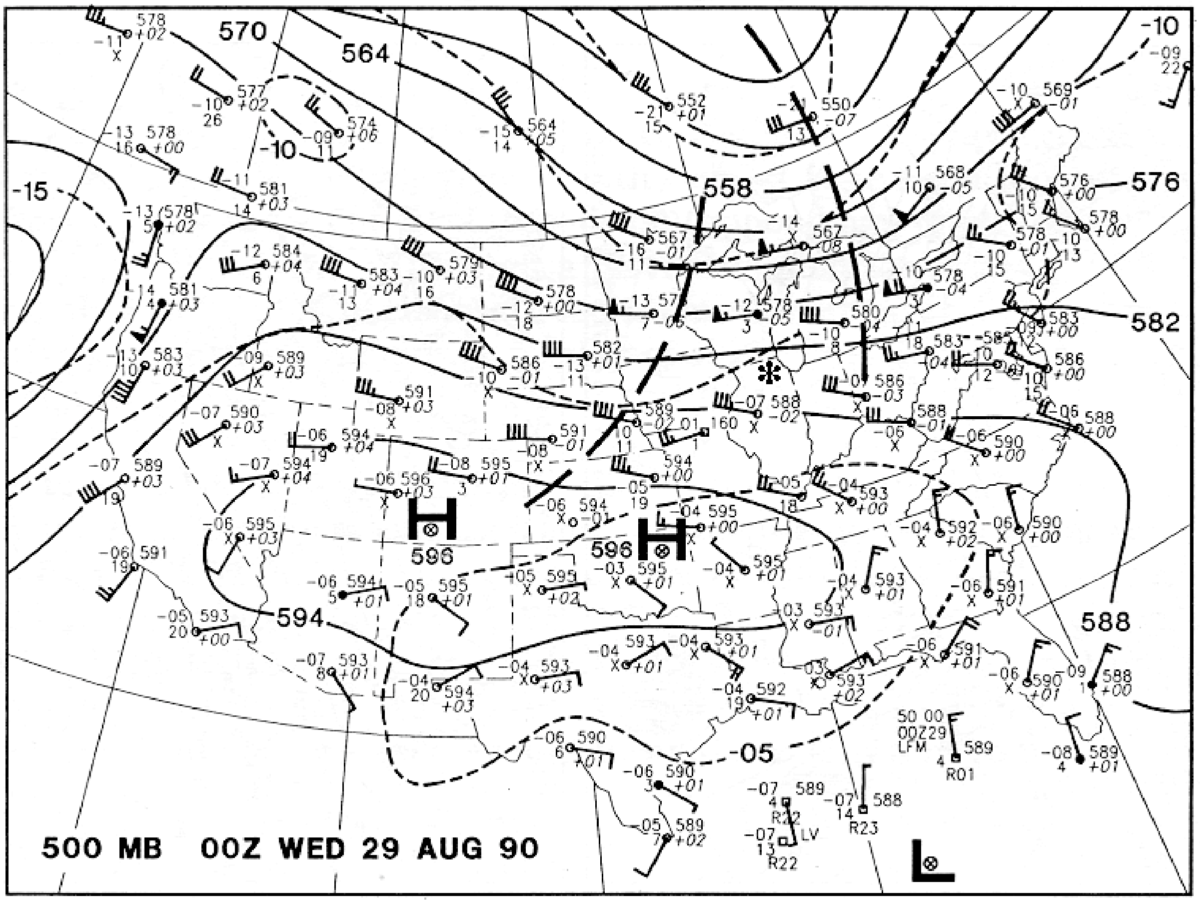


Figure 15. 500 hPa chart at 0000 UTC on 29 August 1990. Plainfield, IL is indicated with the "*" symbol; thick dashed lines denote shortwave trough axes.

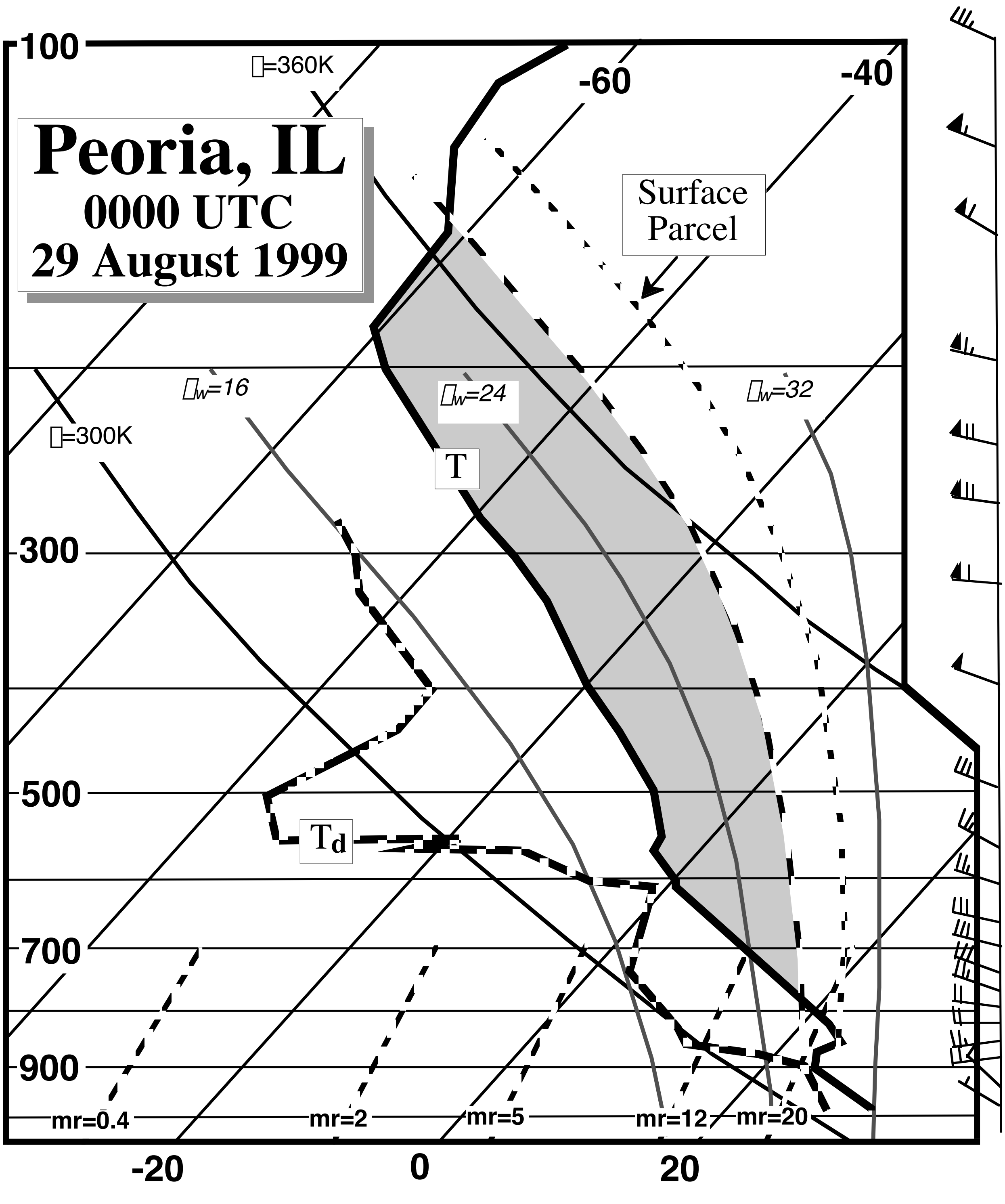


Figure 16. Sounding at Peoria, Illinois at 0000 UTC on 29 August 1990, including plotted wind profile. The positive area (CAPE) is stippled, for a surface-based mixed layer parcel (hatched curve). Also shown is the ascent curve for a surface parcel (finely-hatched curve).

(a)

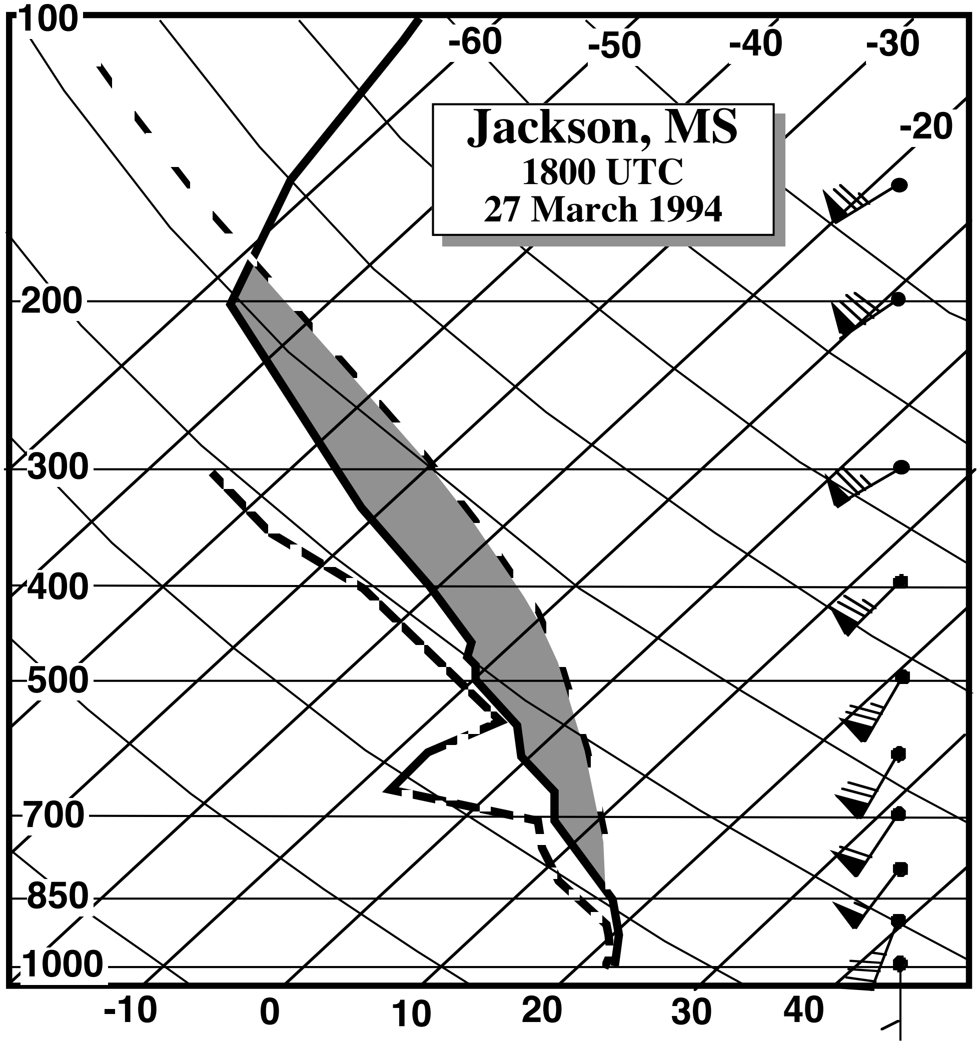
Lifted Index = -6
Total Totals = 52
K-Index = 40
SWEAT Index = 440
Cap = 0.2

CAPE = 2348 J kg^{-1}
CIN = 2 J kg^{-1}

Lifted Parcel Level = 988 mb
Equilibrium Level = 42,000 ft
Max Parcel Level = -999 ft

Lifted Condensation Level = 361 ft
Freezing Level = 13, 638 ft
Wet Bulb Zero = 11, 752 ft

Precipitable Water = 1.96 in



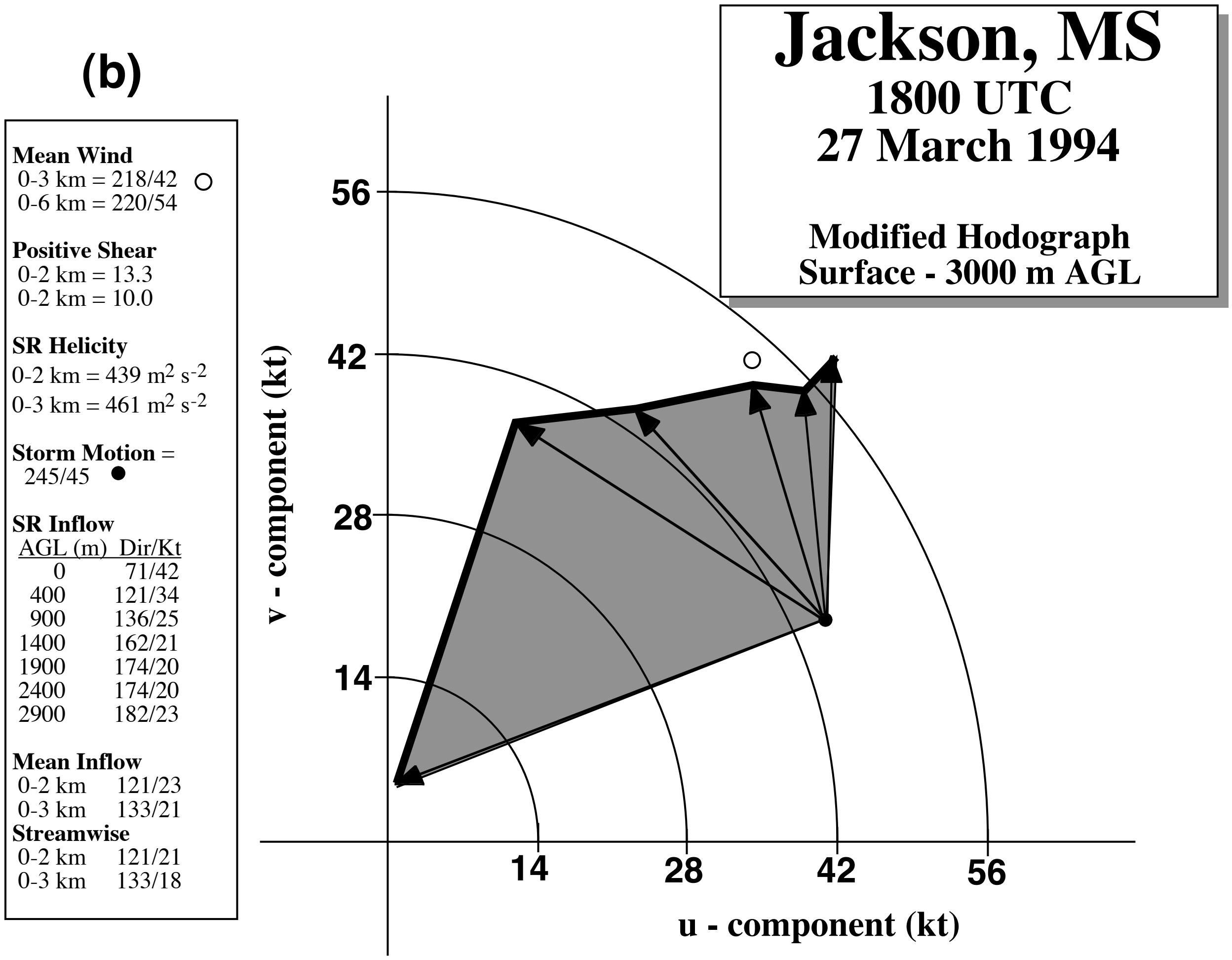


Figure 17. (a) Sounding (plotted on a skew-T, log-p diagram) and (b) hodograph for Jackson, Mississippi at 1800 UTC on 27 March 1994. The hodograph was modified as in National Weather Service (1994).

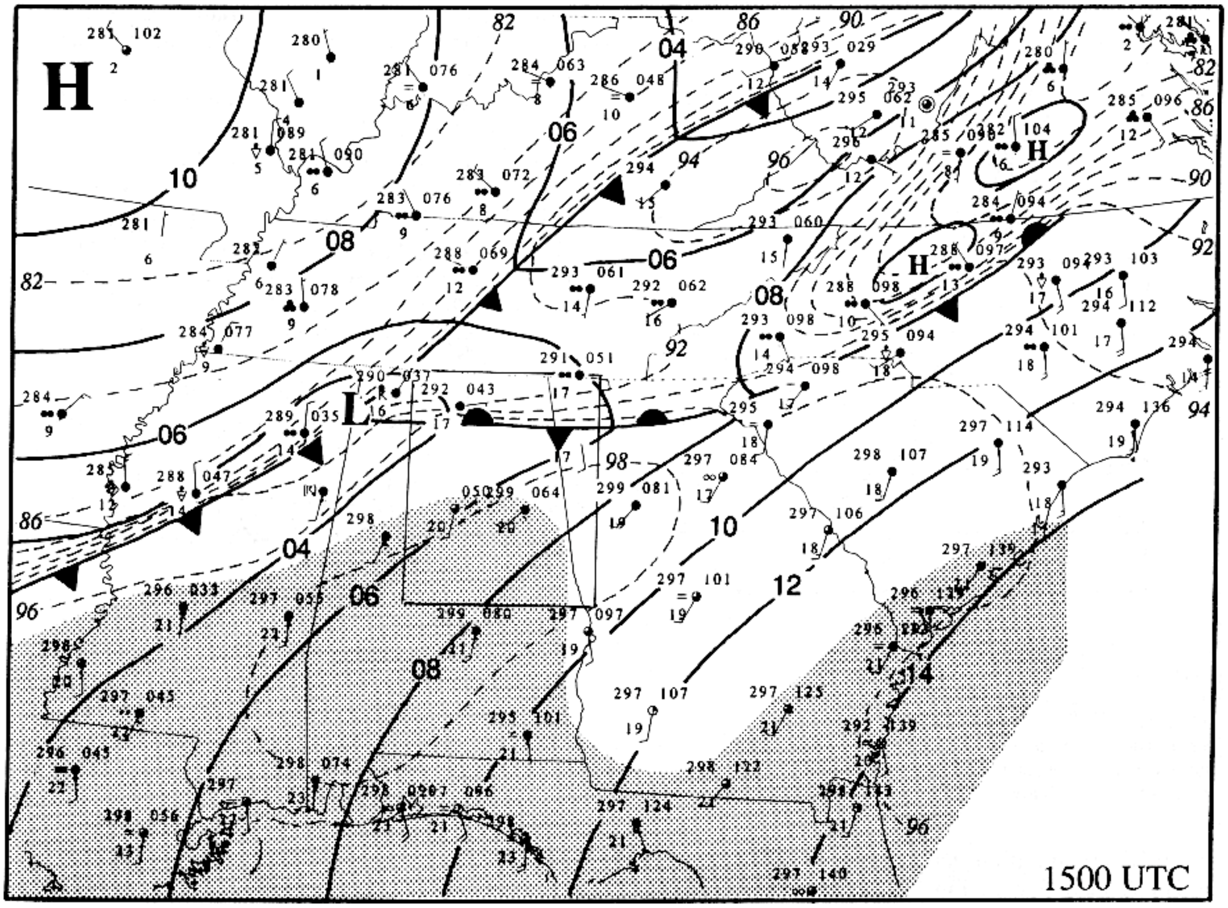


Figure 18. Surface analysis for 1500 UTC on 27 March 1994 (from Langmaid and Riordan 1998); isobars and frontal symbols are conventional, isotherms are dashed lines (deg F), and shading denotes areas with dewpoints ≥ 20 C (68 F).

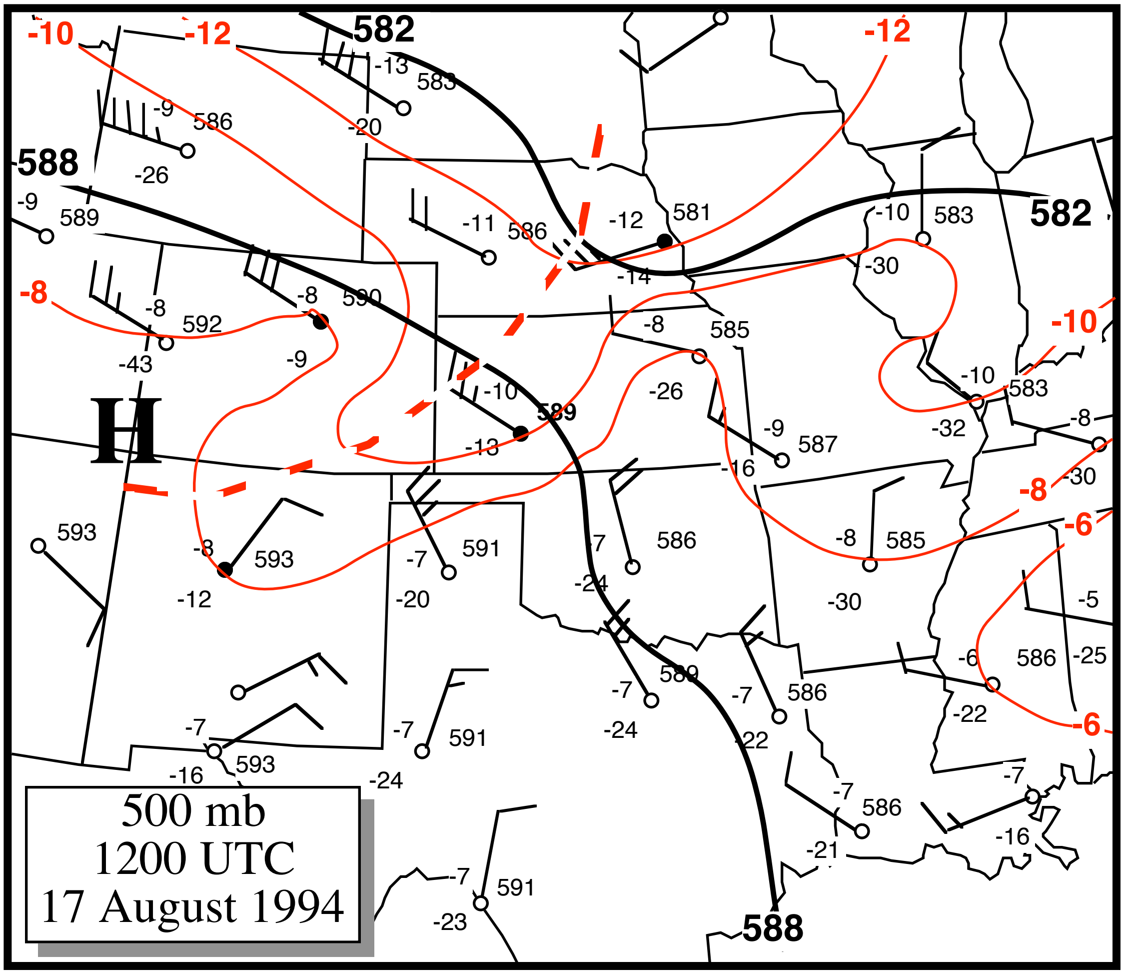


Figure 19. Analysis at 500 hPa at 1200 UTC on 17 August 1994; thick solid lines are isohypses (dam), thin gray lines are isotherms (deg C), the thick hatched line denotes a thermal trough.

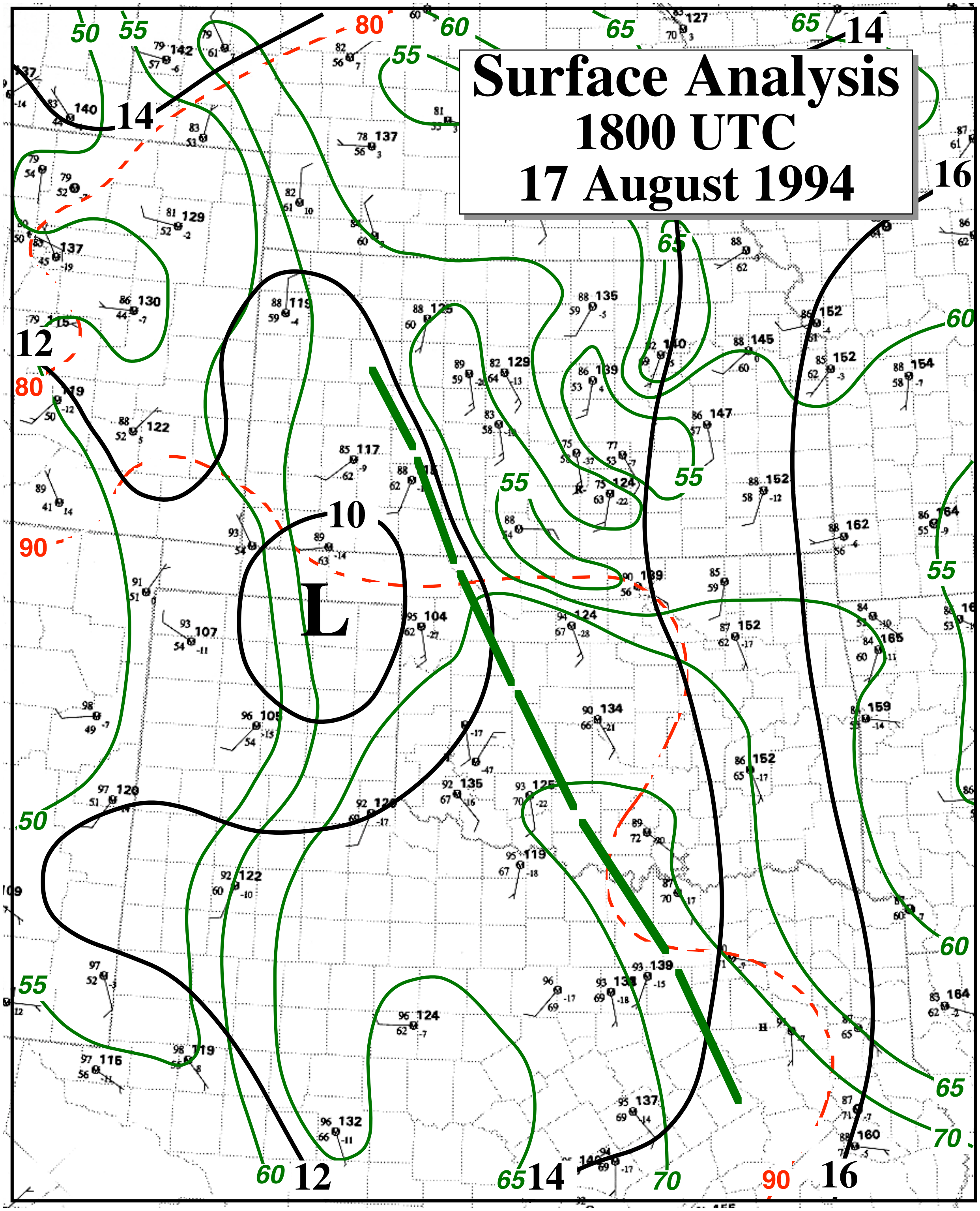


Figure 20. Surface analysis at 1800 UTC on 17 August 1994; thick solid lines are isobars (hPa, labeled conventionally), thin hatched lines are isotherms (10 deg F interval), thin gray lines are isodrosotherms (5 deg F interval), and the moisture axis is depicted by the thick dashed line.

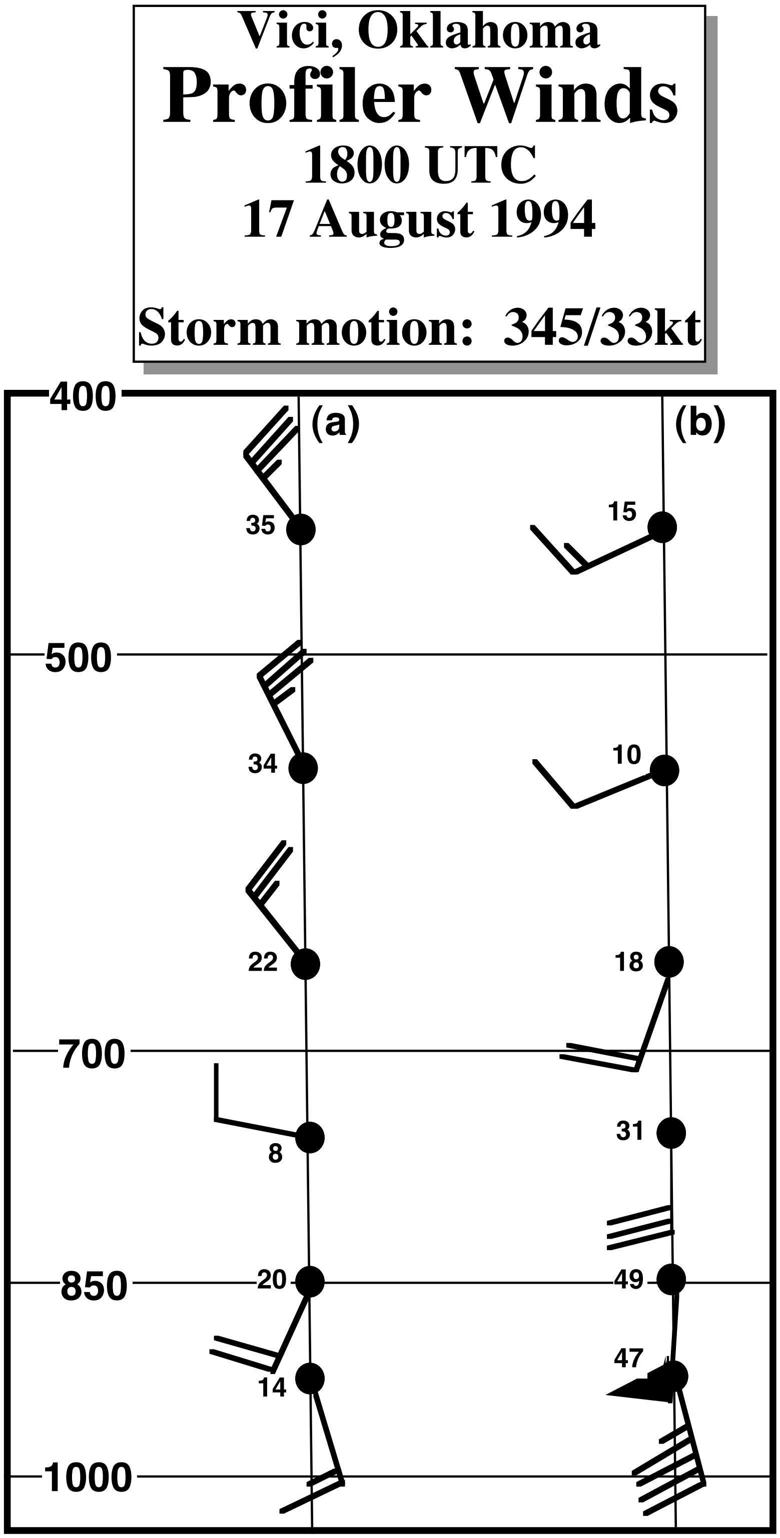


Figure 21. (a) Ground-relative, and (b) storm-relative wind profiles from the Vici, OK profiler (in northwestern OK) at 1800 UTC on 17 August 1994; wind barbs are conventional.

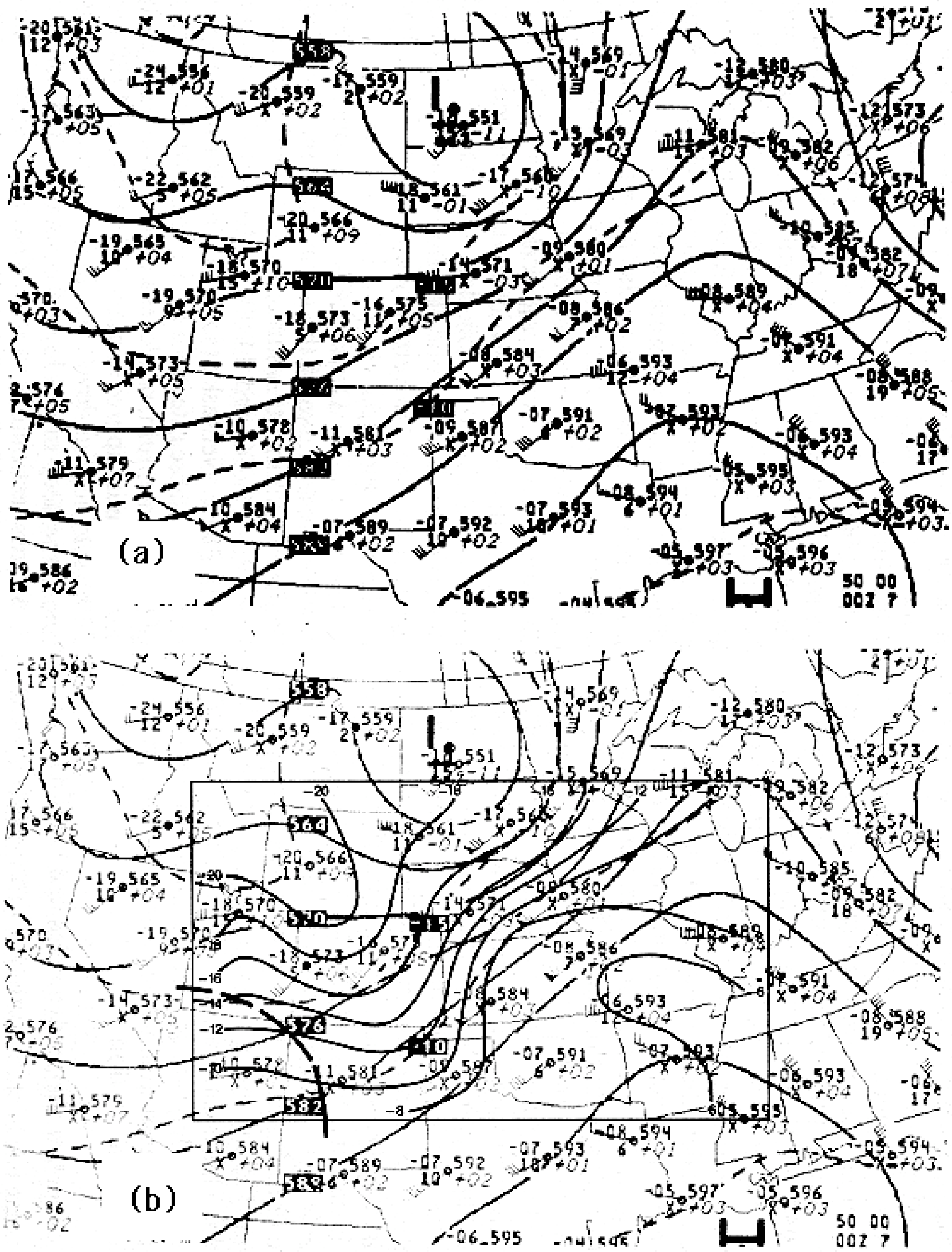


Figure 22. A comparison of (a) basic NMC (National Meteorological Center, now known as the Environmental Modeling Center) analysis at 850 hPa for 0000 UTC on 7 June 1982, and (b) as reanalyzed by Rockwood and Maddox (1988).

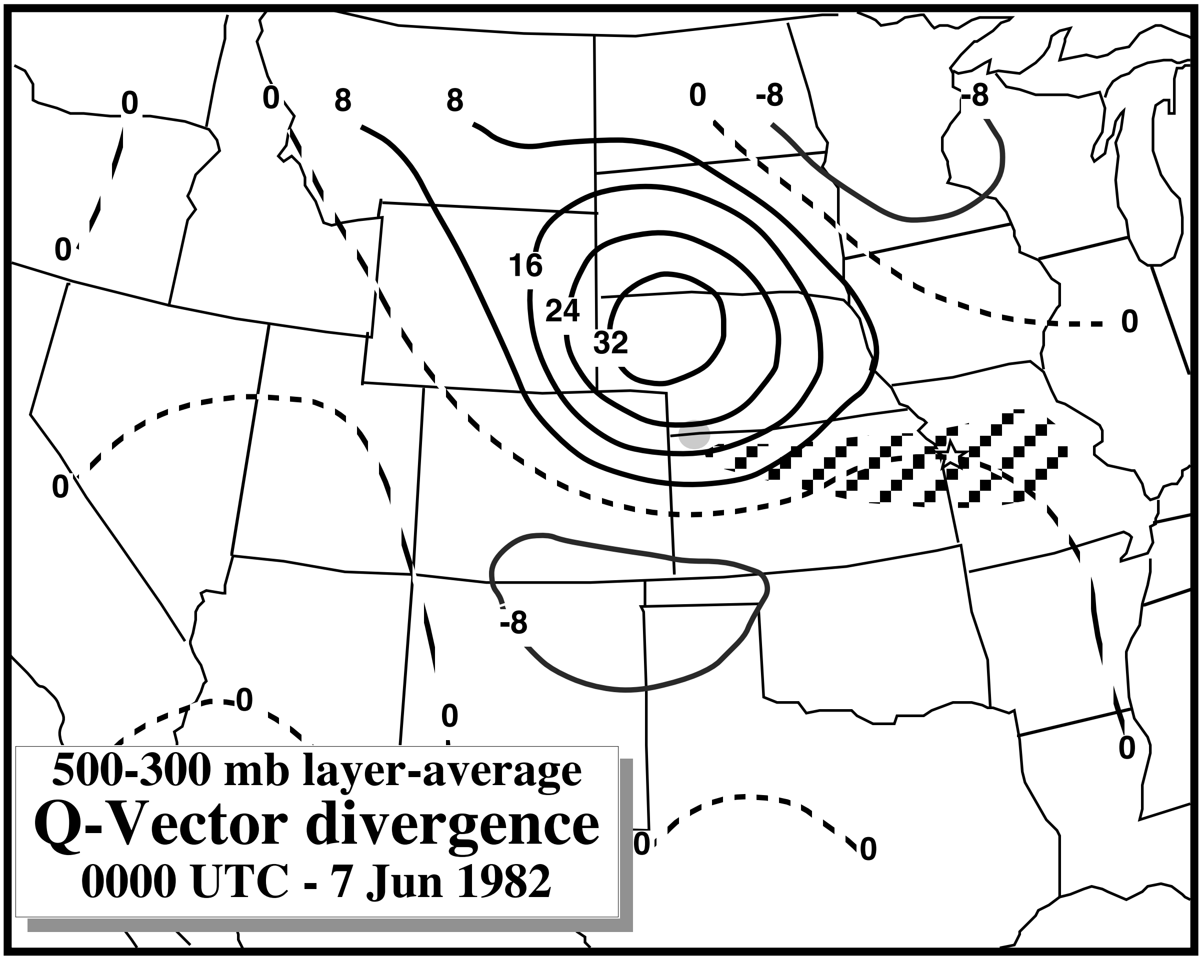


Figure 23. 500-300 hPa layer averaged Q-vector divergence field at 0000 UTC on 7 June 1982, showing forcing for vertical motion; positive values imply downward motion, isopleths are in units of 10^{-17} s^{-3} hPa^{-1} (after Rockwood and Maddox 1988). The calculation follows that described in Barnes (1985). Shading indicates location of first storm development, hatched area shows the region of reported severe weather, and the Kansas City, MO area is indicated by a star.

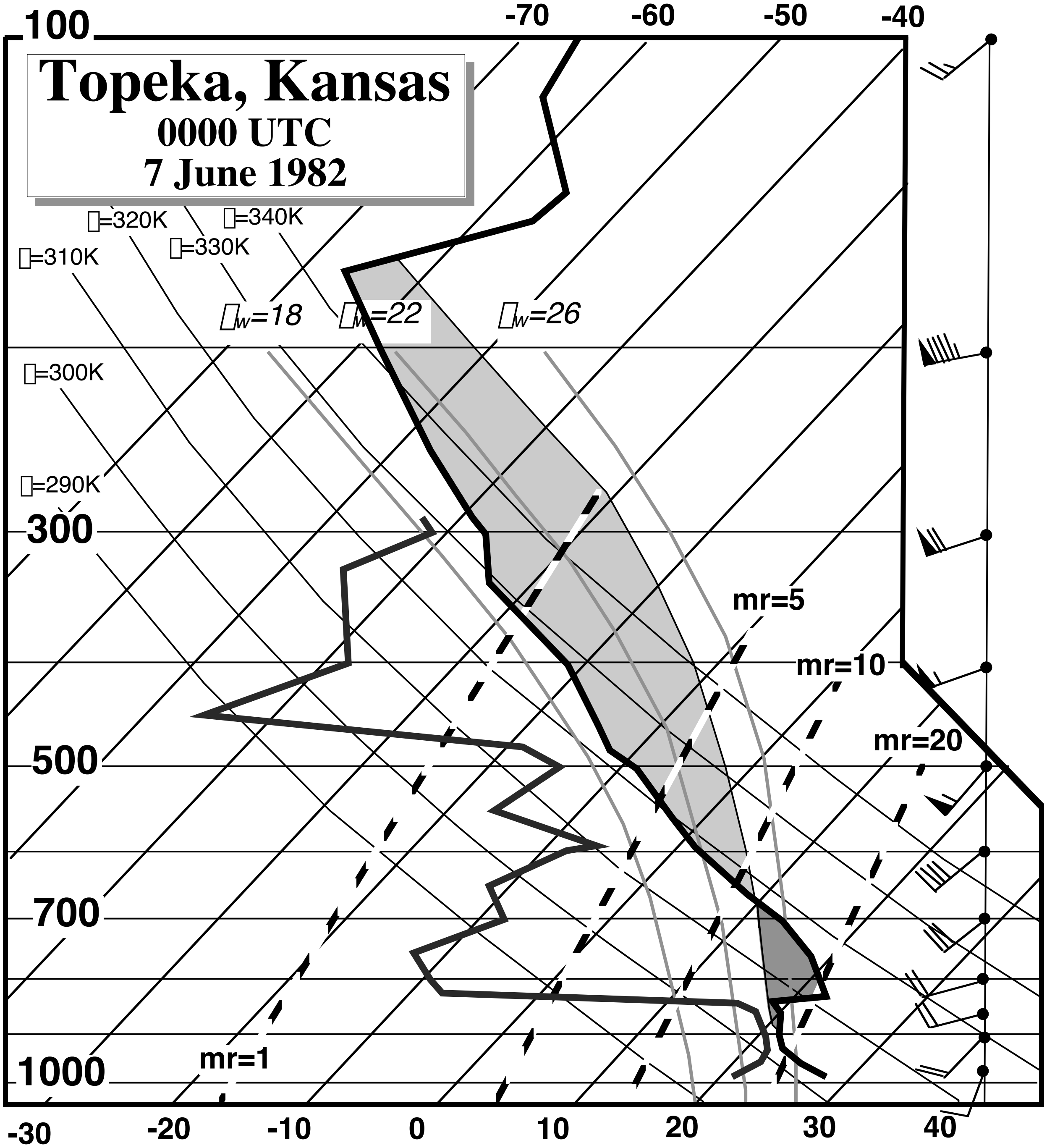


Figure 24. Topeka, Kansas sounding at 0000 UTC on 7 June 1982, plotted on a skew-T, log-p diagram, also showing plotted wind profile. The dark shaded area is the negative area, and the light shaded area is the positive area, for a parcel lifted from the surface layer.

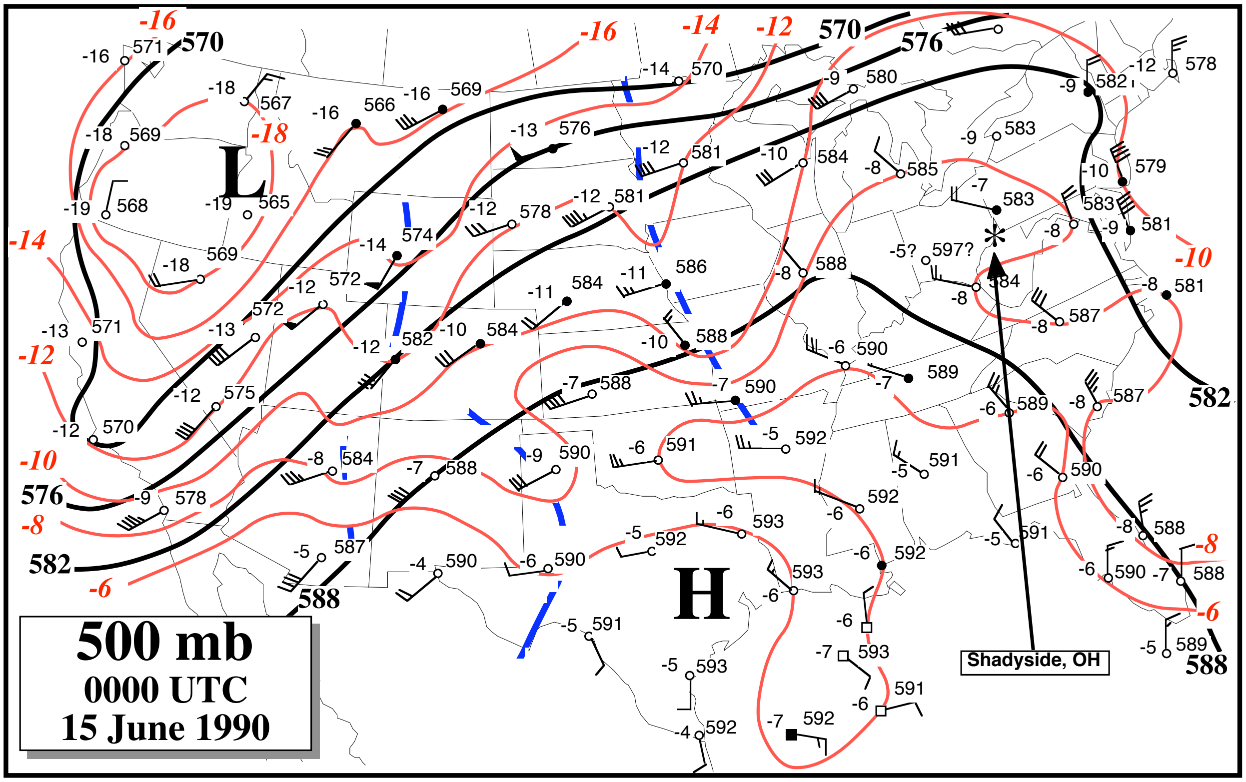


Figure 25. 500 hPa analysis at 0000 UTC on 15 June 1990; thick solid lines are isohypes (dam), thin grey lines are isotherms (deg C), thick hatched lines denote thermal troughs; Shadyside, OH is indicated by the “*” symbol.

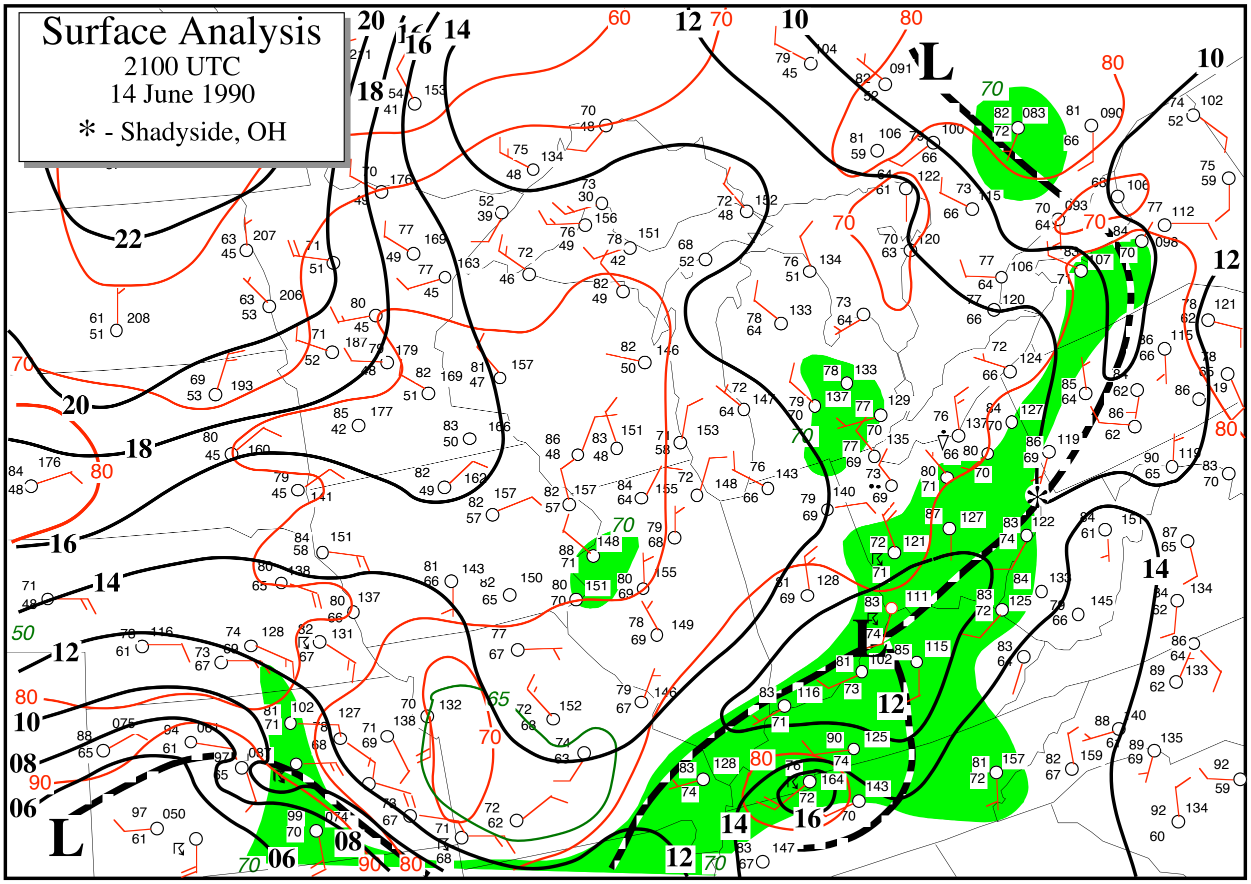


Figure 26 Surface analysis at 2100 UTC on 14 June 1990; thick solid lines are isobars (2 hPa interval), thin gray lines are isotherms (10 deg F interval) and shading denotes areas with dewpoints ≥ 70 deg F; Shadyside, OH is indicated by the "*" symbol. Thick stippled lines denote surface boundaries; station model is conventional.

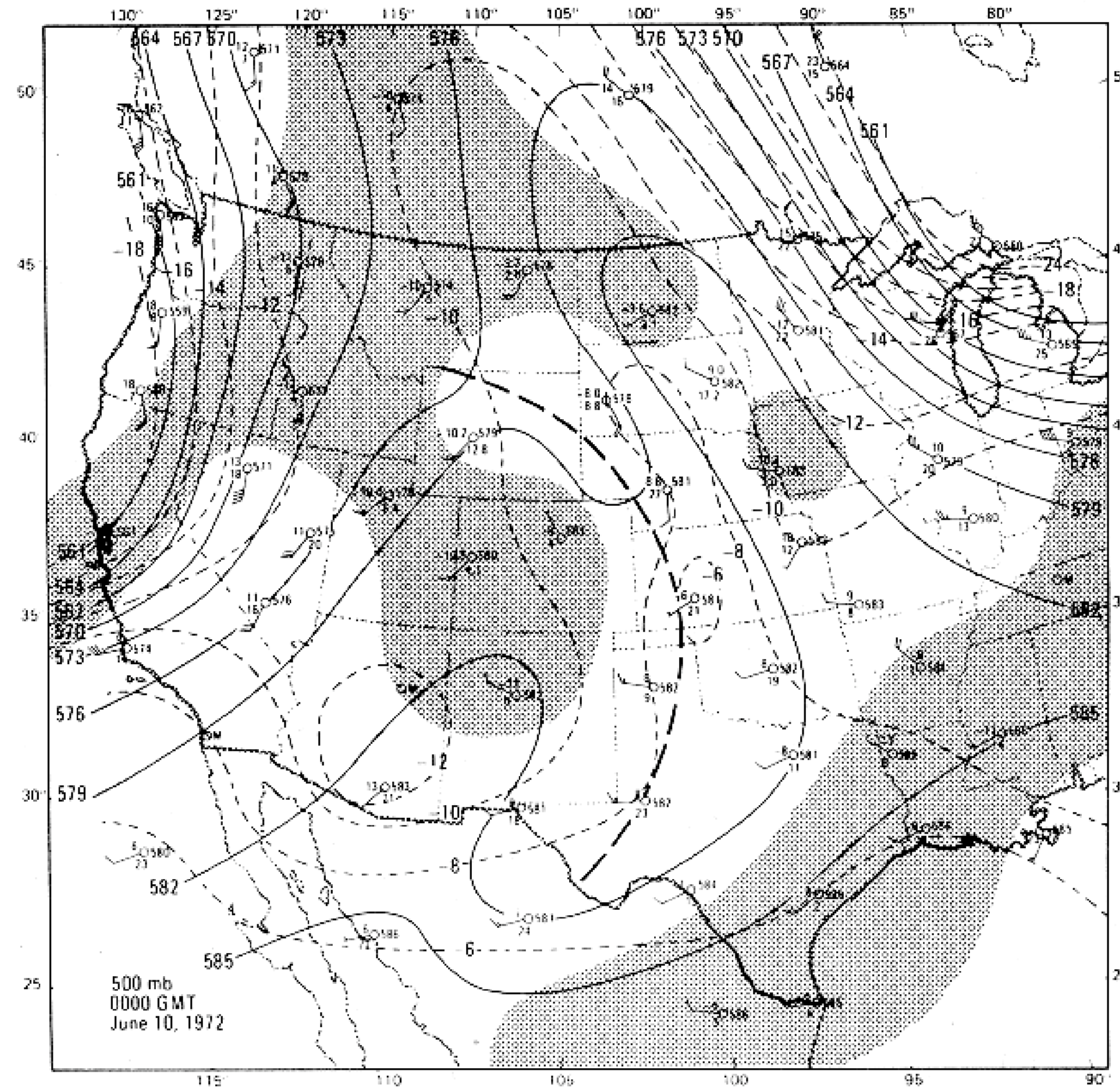


Figure 27. 500 hPa analysis at 0000 UTC on 10 June 1972 (after Maddox et al. 1978), with heavy solid lines showing isohypses (dam), gray hatched lines showing isotherms (deg C) and the heavy hatched line depicting a shortwave trough axis; shading denotes regions with dewpoint depressions of 6ºC or less. The location of Rapid City, South Dakota is indicated by the "*" symbol.

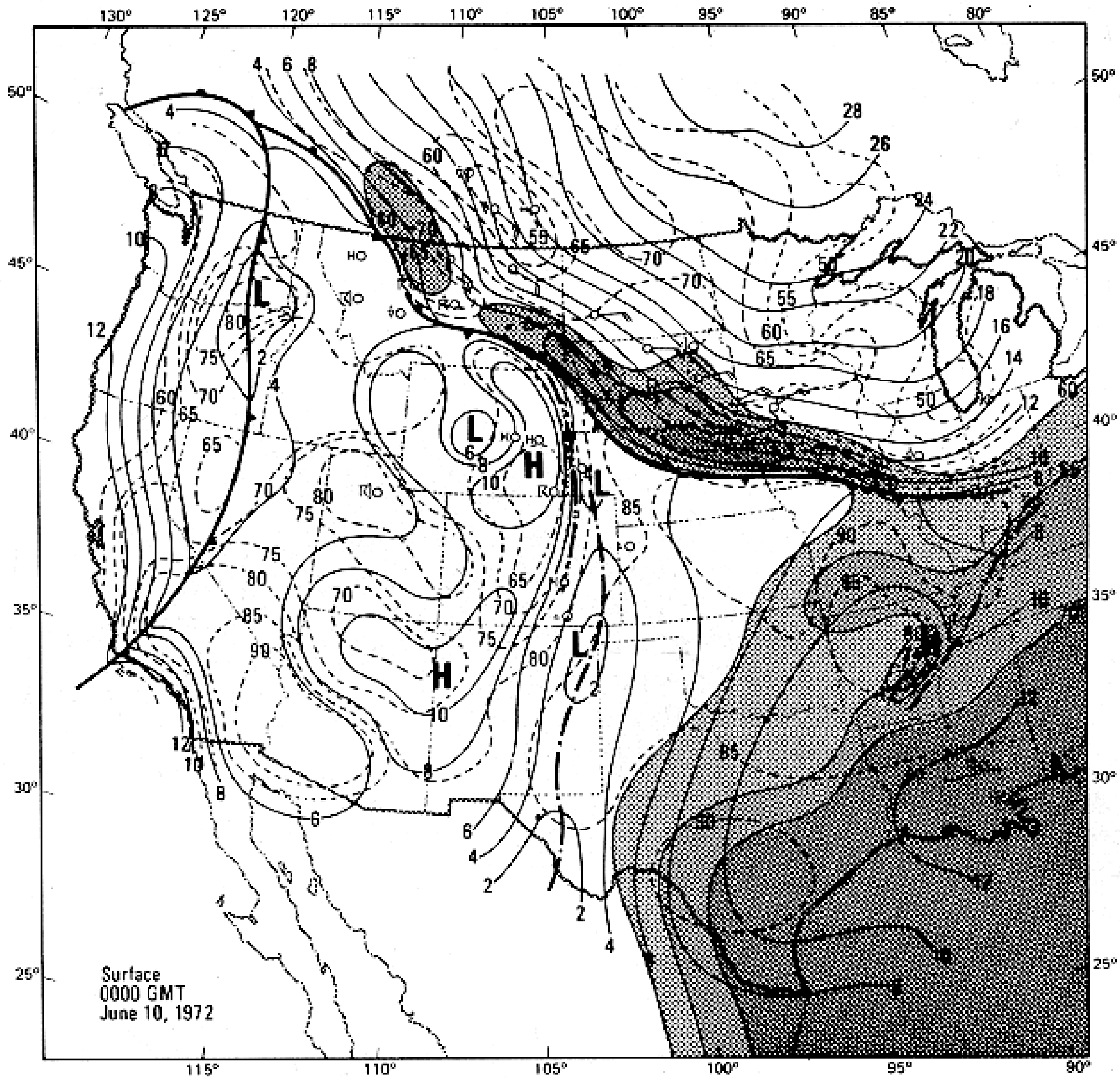


Figure 28. Surface analysis at 0000 UTC on 10 June 1972 (Maddox et al., 1978). Frontal positions, pressure centers and isobars (interval of 2 mb) are solid lines. Isotherms (interval of 5ºF) are dashed lines; isodrosotherms $\geq$ 60ºF are shaded, with solid lines (interval of 5ºF). The location of Rapid City is indicated by the white "*" symbol.

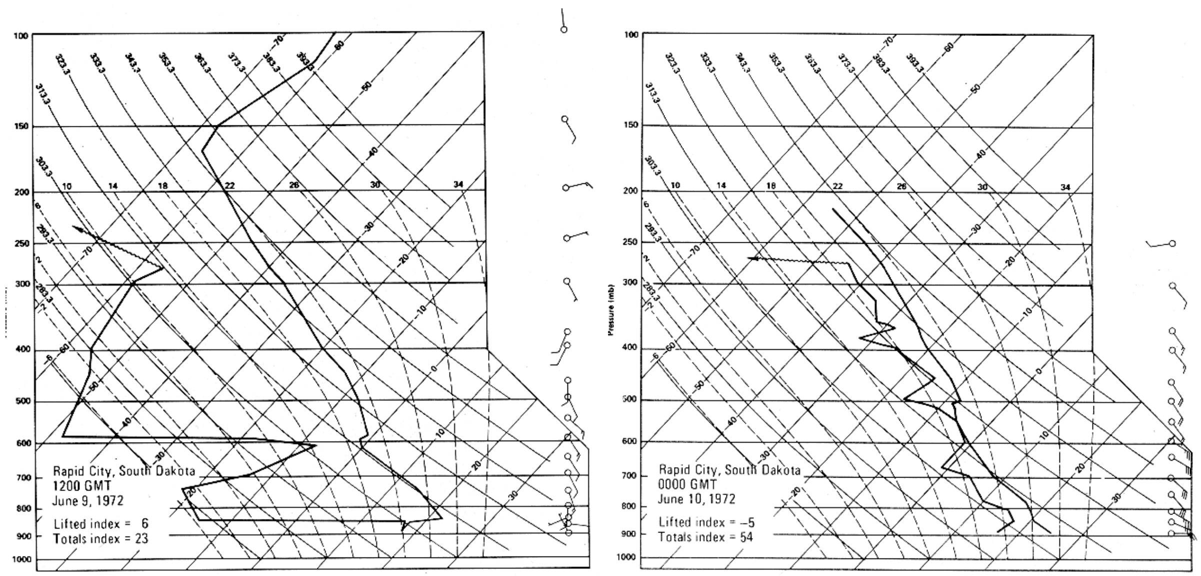


Figure 29. Rapid City, South Dakota soundings at (a) 1200 UTC 9 June 1972 and (b) 0000 UTC 10 June 1972, plotted on a skew-T, log-p diagram.

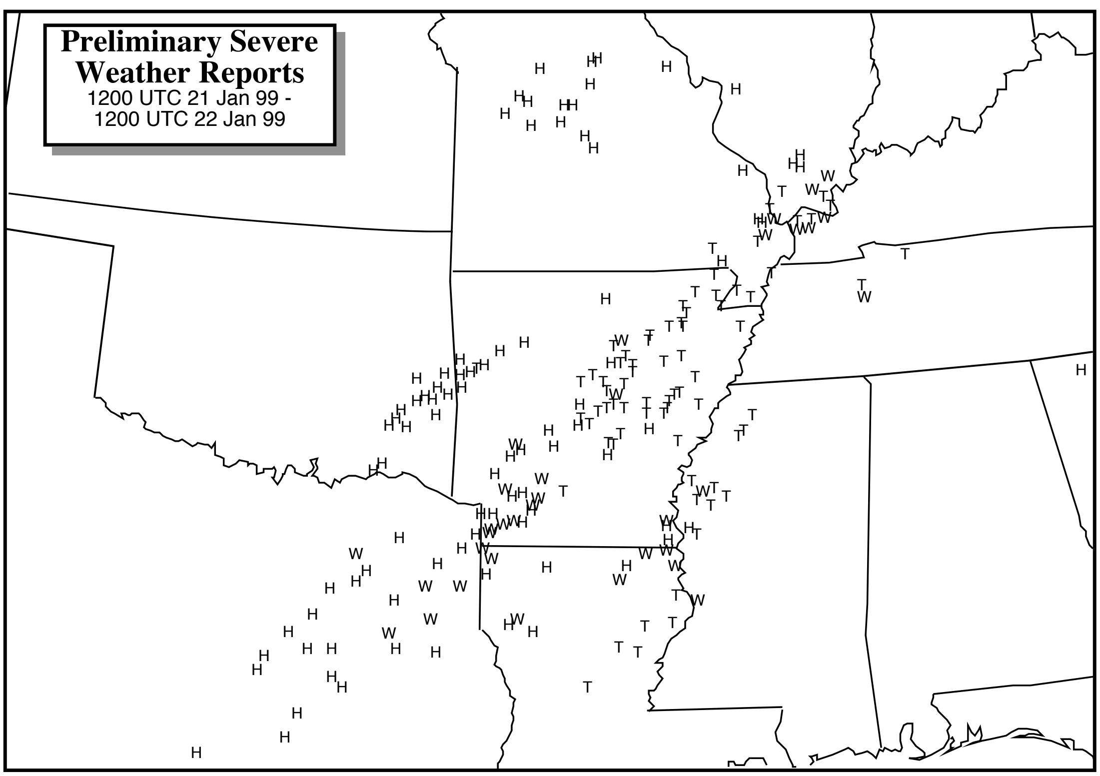
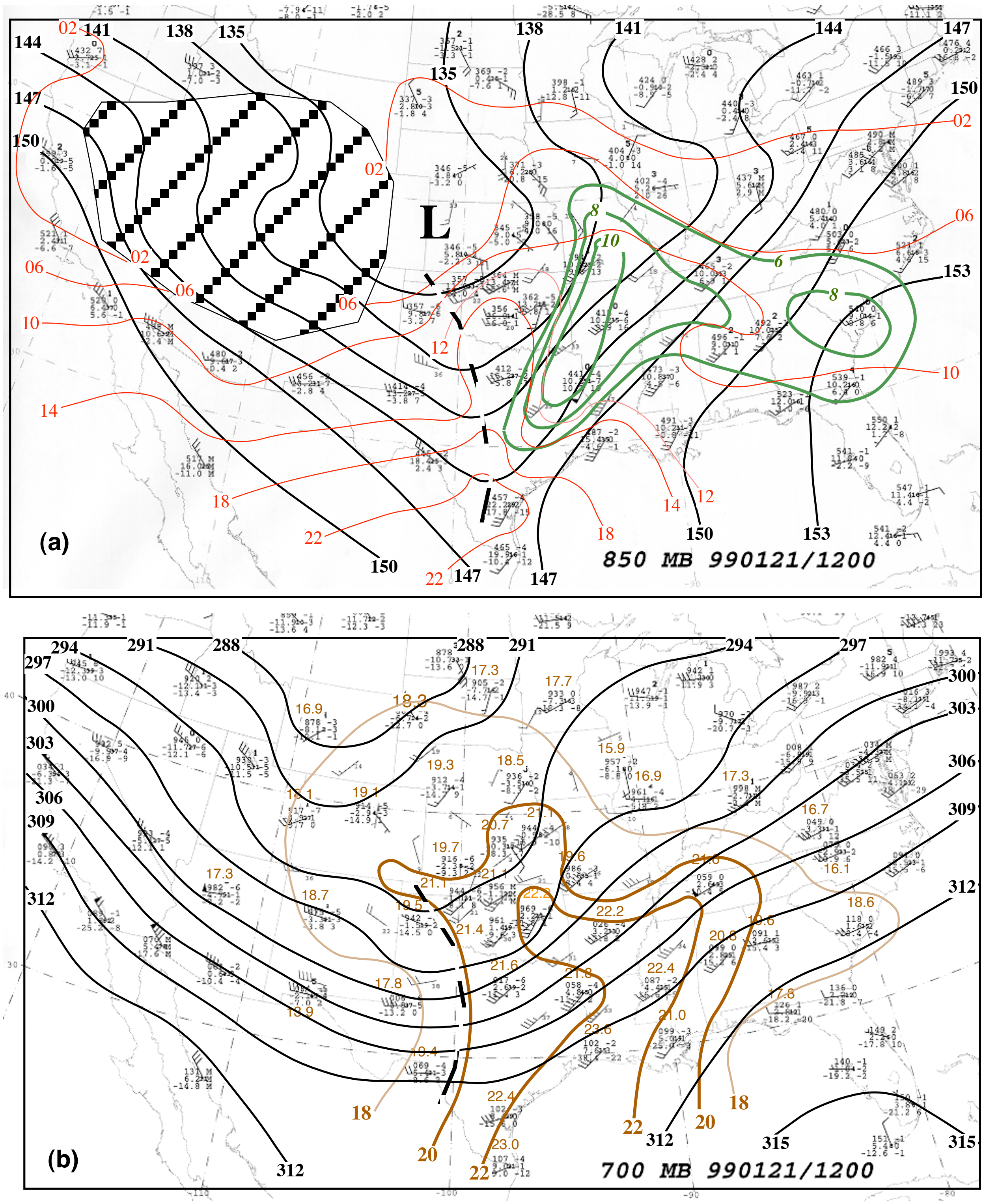


Figure 30. Plot of the preliminary severe weather reports from 1200 UTC, 21 January 1999, to 1200 UTC, 22 January 1999. Hail, convective wind gust, and tornado reports are denoted by “H”, “W”, and “T”, respectively.


(a)
L
850 MB 990121/1200
(b)
700 MB 990121/1200

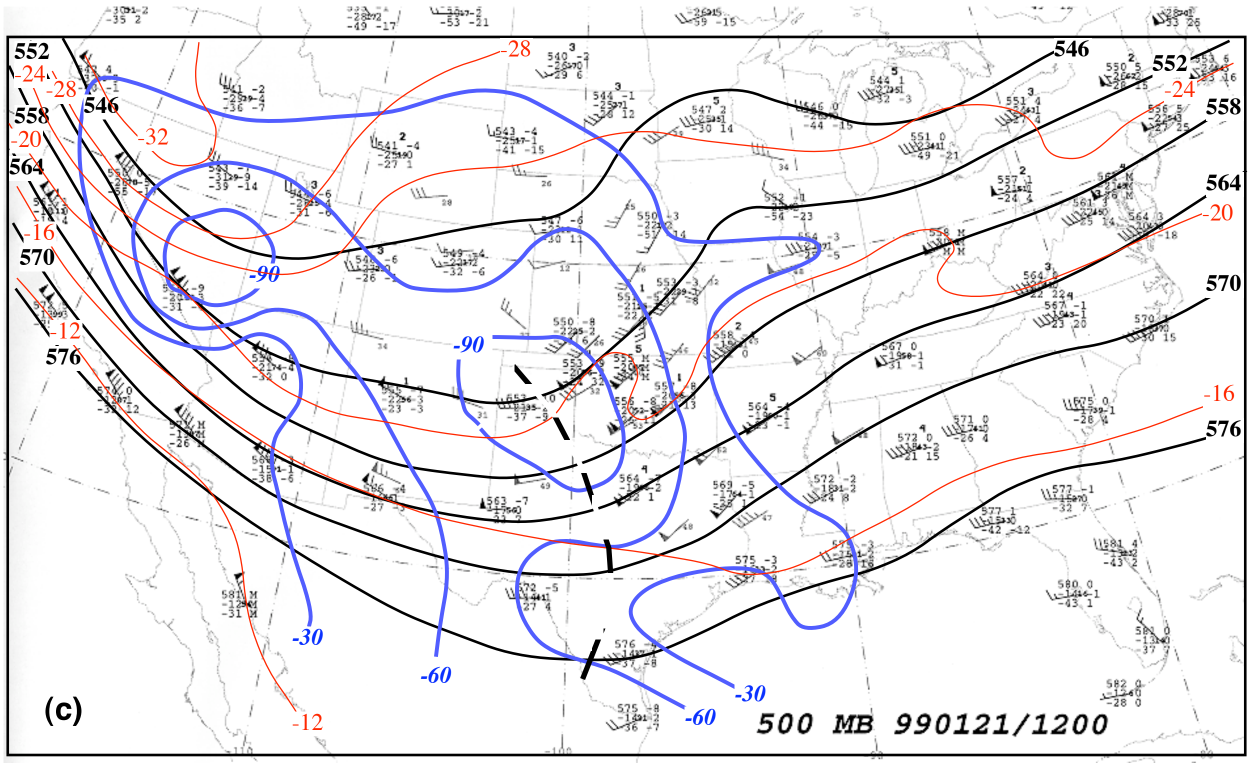


Figure 31. Analyses at (a) 850 hPa, (b) 700 hPa, and (c) 500 hPa. At 850 hPa, thick solid lines depict isohypes (in dam), thin solid lines are isotherms (deg C), and thick gray lines depict isodrosotherms (deg C), with a trough line depicted by a thick hatched line; at 700 hPa, the convention is the same, except the thick gray lines depict isopleths of the temperature difference (deg C) between 700 and 500 hPa; at 850 hPa, the convention is the same, except the thick gray lines are isopleths of geopotential height changes (m). The hatched area on the 850 hPa is excluded from any thermal analysis since the associated stations are generally below the earth's surface. The station model for all charts is conventional, except that the heights are plotted on the upper left side and dewpoint temperatures are used instead of dewpoint depressions; plotted numbers on the right side of the station model are 12-h changes.

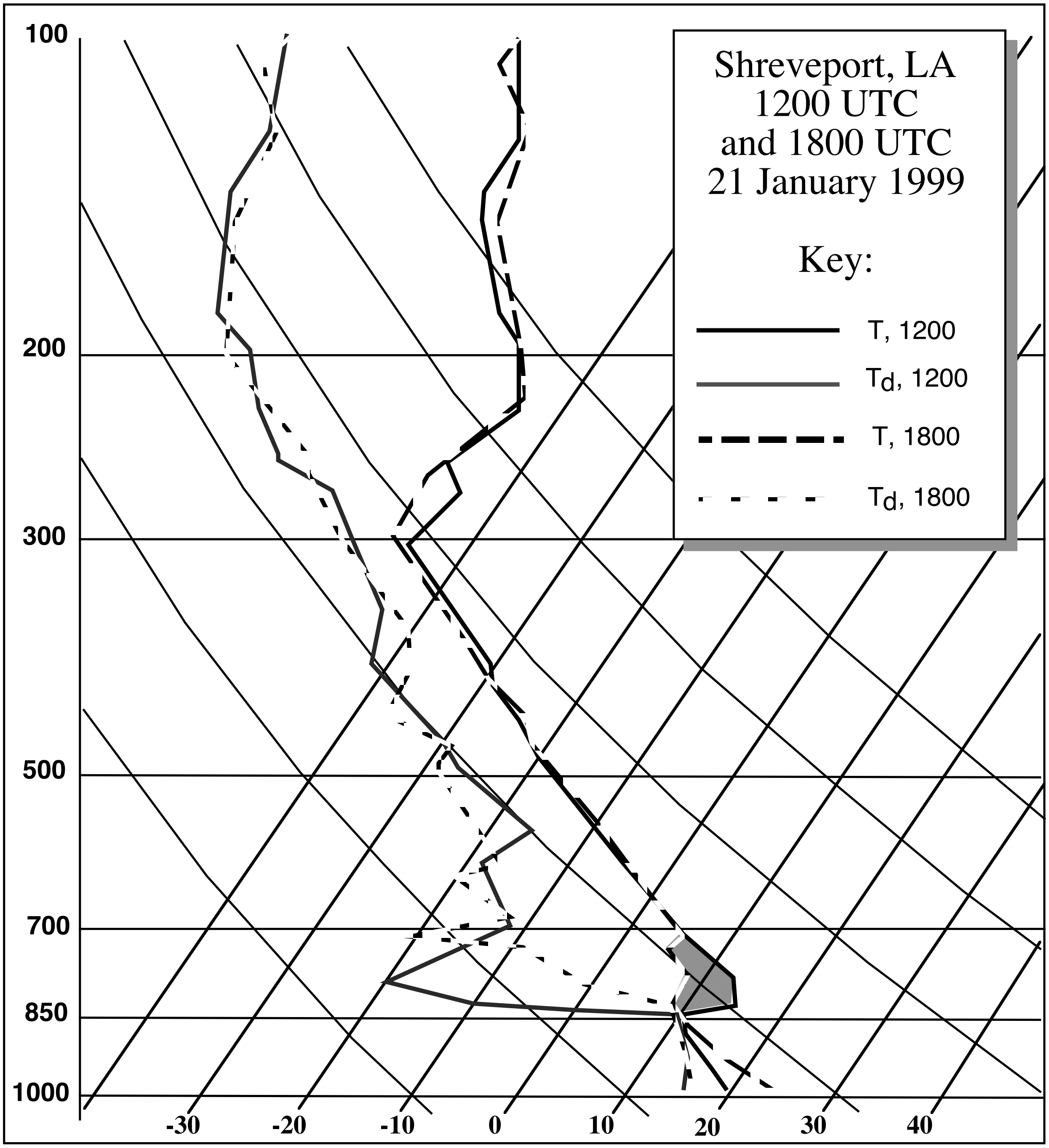


Figure 32. Soundings for Shreveport, LA at 1200 and 1800 UTC on 21 January 1999 plotted on a skew T, log p diagram. Shaded area depicts the temperature change in the 700 and 850 hPa from 1200 to 1800 UTC.