Use of the Parcel Buoyancy Minimum ($B_{\text{min}}$) to Diagnose Simulated Thermodynamic Destabilization. Part I: Methodology and Case Studies of MCS Initiation Environments

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ABSTRACT

A method based on parcel theory is developed to quantify mesoscale physical processes responsible for the removal of inhibition energy for convection initiation (CI). Convection-permitting simulations of three mesoscale convective systems (MCSs) initiating in differing environments are then used to demonstrate the method and gain insights on different ways that mesoscale thermodynamic destabilization can occur.

Central to the method is a thermodynamic quantity $B_{\text{min}}$, which is the buoyancy minimum experienced by an air parcel lifted from a specified height. For the cases studied, vertical profiles of $B_{\text{min}}$ using air parcels originating at different heights are qualitatively similar to corresponding profiles of convective inhibition (CIN). Though it provides less complete information than CIN, an advantage of using $B_{\text{min}}$ is that it does not require vertical integration, which simplifies budget calculations that enable attribution of the thermodynamic destabilization to specific physical processes. For a specified air parcel, $B_{\text{min}}$ budgets require knowledge of atmospheric forcing at only the parcel origination level and some approximate level where $B_{\text{min}}$ occurs.

In a case of simulated daytime surface-based CI, destabilization in the planetary boundary layer (PBL) results from a combination of surface fluxes and upward motion above the PBL. Upward motion effects dominate the destabilizing effects of horizontal advections in two different simulated elevated CI cases, where the destabilizing layer occurs from 1 to 2.5 km AGL. In an elevated case with strong warm advection, changes to the parcel at its origination level dominate the reduction of negative buoyancy, whereas for a case lacking warm advection, adiabatic temperature changes to the environment near the location of $B_{\text{min}}$ dominate.

1. Introduction

One of the most difficult aspects of deep, precipitating convection to quantify and predict is its initiation. This is because atmospheric processes of varying scales influence both the thermodynamic vertical structure in which convection occurs and how it is initiated. Recent research has emphasized the latter, partly because of the increasing availability of radar, satellite, and high-resolution surface data used to detect finescale flow features (e.g., drylines, sea-breeze fronts, remnant convective outflows, internal gravity waves, and horizontal rolls) along which convection initiation (CI) often occurs. The widths of these phenomena are $L \sim 1$–10 km. Horizontal convective rolls (HCRs) are one of the above examples that influence CI not only by the lifting they impart, but also because of the thermodynamic contrasts within them (e.g., Mueller et al. 1993; Weckwerth et al. 1996). The role of mesoscale ($L \sim 100$ km) environmental temperature/moisture advections and vertical motions on organized convection is perhaps even more widely recognized (e.g., Maddox 1983; Crook and Moncrieff 1988). However, because of our inability to sample these effects at appropriate resolution over mesoscale regions, their influence on the thermodynamic vertical structure, crucial to CI, remains poorly quantified.

Though detailed observations are typically lacking, current research numerical models are capable of both explicitly simulating deep, moist convection and representing the mesoscale environmental thermodynamic variations often associated with CI. In the current paper we develop a method, based on parcel theory, that allows us to evaluate the influence of different physical processes on the evolution of the mesoscale thermodynamic vertical structure. Case studies with a convection-permitting

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numerical model are used to demonstrate the method and examine how these physical processes modify thermodynamic environments allowing the initiation of observed mesoscale convective systems (MCSs) in three different midlatitude warm-season meteorological situations.

The widely used parcel theory of convection neglects vertical accelerations due to pressure forces, as well as buoyancy effects due to condensate loading and entrainment of environmental air. Since these effects typically have a deleterious impact on updraft strength, the convective available potential energy,

$$\text{CAPE} = \int_{p_{\text{LFC}}}^{p_{\text{LNB}}} R(T_{vp} - T_{vc}) d\ln(p),$$

derived from parcel theory provides only an approximate upper bound on the strength of the convection. The CAPE defined in (1) is the integrated buoyancy, $B = R(T_{vp} - T_{vc})/p$, from the level of free convection (LFC) to the level of neutral buoyancy (LNB) arising from differences between the parcel ($T_{vp}$) and environmental ($T_{ve}$) virtual temperatures.

Although the CAPE provides an estimate of how strong convection can become, it offers little guidance on whether it will actually occur. Indeed, expansive areas of positive CAPE often occur during the midlatitude warm season (e.g., Dai et al. 1999), but areas experiencing deep convection at any given time are much smaller. Given positive CAPE, the magnitude of the convective inhibition,

$$\text{CIN} = \int_{p_{\text{LFC}}}^{p_{\text{LNB}}} R(T_{vp} - T_{vc}) d\ln(p),$$

which is the vertically integrated negative buoyancy between the parcel origination level $p_{\text{LFC}}$ and its LFC, is a better indicator of whether deep convection is likely to occur. Convective inhibition (CIN; Colby 1984) is illustrated schematically by the shaded region in Fig. 1 for a planetary boundary layer (PBL) air parcel.

The minimum vertical velocity of an air parcel, $w_{\text{min}}$, required to overcome the CIN may be estimated by substituting the form of the vertical momentum equation derived from parcel theory, $dw_{p}/dt = w_{p}(dw_{p}/dz) = B$, into (2) and taking the square root, which yields $2(-\text{CIN})^{0.5}$. Thus, even for a seemingly small magnitude of CIN = $-50 \text{ J kg}^{-1}$, a strong updraft of $w_{\text{min}} = 10 \text{ m s}^{-1}$ is required to initiate deep convection. The infrequent occurrence of vertical velocities of this magnitude outside of preexisting convection indicates the importance of physical processes acting over longer time scales to eliminate or substantially reduce the CIN.

For a given $T_{vc}$ vertical profile, CAPE depends only on the air parcel’s moist static energy, $H = c_p T + L \delta q + gz$, where $c_p$ is a specific heat at a constant pressure, $L$ is the latent heat of vaporization, and $g$ is gravity. This implies $\delta q_p (\text{g kg}^{-1}) \approx 2.58 T_{ve} (\text{K})$ for equivalent moist static energy perturbations to the parcel. The contributions of parcel temperature and moisture perturbations to CIN changes are more complicated and, unlike for CAPE, do not depend uniquely on moist static energy changes. In contrast to CAPE, which is completely specified by initial parcel pressure, environmental virtual temperature, and parcel vertical displacements along a moist adiabat where $H$ is constant, contributions to CIN also depend on unsaturated lifting beneath the parcel’s lifted condensation level (LCL). Thus, for a fixed increment of parcel $\delta H$, the relative importance of $\delta T_p$ and $\delta \delta q_p$ to CIN changes depends on both the parcel’s relative humidity at its source level and the environmental static stability (Crook 1996).

Previous diagnostic studies have used CIN and related thermodynamic quantities to assess physical factors influencing CI. For example, Ziegler and Rasmussen (1998) used soundings modified by a simple numerical transport model to illustrate the importance of boundary layer circulations in reducing CIN to values supportive of CI. Carlson et al. (1980, 1983) developed the lid strength index (LSI) to illustrate the roles of differential horizontal advection and transverse vertical circulations on focusing CI. Unlike CIN, the LSI is not an integral quantity and considers average parcel and environmental properties within two separate layers.

In the current study we analyze physical factors influencing $B_{\text{min}} = \min(T_{vp} - T_{vc})$, which is the minimum
buoyancy (in temperature units) of a vertically displaced air parcel. The blue line in Fig. 1 illustrates $B_{\text{min}}$ for a 50-hPa-deep surface-based parcel. This thermodynamic parameter is related to the CIN in a manner analogous to how the minimum lifted index (LI) is related to the CAPE, and has been used successfully to determine the susceptibility to deep convection in modeling studies ranging from the examination of the effects of land–atmosphere exchange strength on deep convection (Trier et al. 2011) to mechanisms supporting long-lived propagating convection (Laing et al. 2012). Though it only approximates the effects of CIN, a practical advantage of analyzing $B_{\text{min}}$ is that it does not require there to be an LFC (and thus positive CAPE) and therefore, like the LSI and LI, it may be displayed as a continuous spatial field (section 3).

More important to the objectives of the current study is that, unlike for CIN, the calculation of $B_{\text{min}}$ requires no vertical integration. This aspect simplifies the related budget calculations (sections 4 and 5), which enable attribution of the simulated thermodynamic destabilization to different physical processes. Specifically, such analysis using $B_{\text{min}}$ requires knowledge of atmospheric forcing at only two different levels: the air parcel origination level and a level approximating where the lifted parcel attains its minimum buoyancy. This results in a conceptually simple and physically revealing framework within which to view thermodynamic destabilization. Such an analysis may be performed for many different desired parcel origination levels (i.e., model vertical grid points), and the results easily combined. This allows a complete picture of the processes governing thermodynamic destabilization to emerge, which itself can occur through layers that are several kilometers deep.

It should be noted that for conditionally unstable cases, the choice of which environmental level to analyze within the negatively buoyant region between the parcel origination level and the LFC is somewhat arbitrary. However, by choosing the level at which $B_{\text{min}}$ occurs, we focus on the location where physical processes have most restricted deep convection. For example, when a strong inhibiting inversion exists near the top of the PBL, $B_{\text{min}}$ is often situated near that location, and analysis of buoyancy rates of change and its forcing at the level of $B_{\text{min}}$ is optimal for providing insight into how processes inhibiting deep convection have evolved.

In section 2 we discuss the CI cases we simulate, which includes an MCS fed by conditionally unstable air from the daytime PBL and two others fed by conditionally unstable air located 1–2 km above the surface. In section 2 we also describe the numerical model we use to simulate these cases. Section 3 provides an overview of and compares the simulated MCS initiations with observations, illustrates the association of CI with local and regional evolutions in $B_{\text{min}}$, and compares these $B_{\text{min}}$ evolutions with evolutions of more widely used thermodynamic variables such as CIN and CAPE. Our budget methodology used to estimate the causes of $B_{\text{min}}$ changes in the thermodynamic destabilization time intervals prior to CI, and results from this analysis for the three different simulated CI cases, are presented in sections 4 and 5, respectively.

2. Case selection and numerical model

a. Cases

The three cases selected for the current study are obtained from two separate 6-day periods, 10–15 June 2002 and 3–9 July 2003, which respectively included portions of the International H2O Project (IHOP; Weckwerth et al. 2004) and the Bow-Echo and Mesoscale Convective Vortex Experiment (BAMEX; Davis et al. 2004). A total of twelve 24-h simulations (six for each period) were performed. The model’s ability to successfully simulate CI and subsequent MCS evolution and our desire to represent a range of MCS evolution and their differing thermodynamic environments were the primary considerations influencing case selection for this study.

Daytime convection over the southern Great Plains (SGP) often initiates near surface drylines, cold fronts, or mergers of the two. The 12 June 2002 initiation of a late-afternoon squall line over western Oklahoma and the eastern Texas Panhandle during IHOP (case 1) is an excellent example of this archetype. A desirable aspect of using this case is that it has been well observed using high-resolution field data (Weckwerth et al. 2008; Champollion et al. 2009) and simulated by others (Xiao and Sun 2007; Liu and Xue 2008). Wilson and Roberts (2006) found that approximately 50% of CI during IHOP, which was conducted over the SGP, comprised elevated convection (e.g., Corfidi et al. 2008) where air fueling convection originates from above the PBL. The second and third cases we present are examples of elevated CI. Case 2 commenced in Iowa during BAMEX on the evening of 4 July 2003. This case, which became a large quasi-stationary MCS, was associated with strong warm advection above a surface front. In case 3, which occurred later at night in the Texas Panhandle on 13 June 2002 of IHOP, MCS initiation occurred without strong warm advection in its immediate vicinity and included multiple convective bands (Marsham et al. 2011).

b. Numerical model and experiment design

Our simulations for the two IHOP cases and the BAMEX case use the Advanced Research core of the
Weather Research and Forecasting Model (ARW-WRF; Skamarock and Klemp 2008). Each simulation uses the same physical parameterizations, which include the Rapid Radiative Transfer Model (RRTM) longwave (Mlawer et al. 1997) and Dudhia (1989) shortwave radiation schemes. The PBL parameterization is the Mellor–Yamada–Janjić (MYJ) scheme (Janjić 2002), which predicts turbulent kinetic energy (TKE) and governs vertical mixing between model layers. Subgrid horizontal mixing is determined using a Smagorinsky-type first-order closure. Each simulation uses the Thompson et al. (2008) bulk microphysical parameterization, which predicts cloud water, cloud ice, rain, snow, and graupel mass.

The Noah land surface model (Ek et al. 2003) is coupled to ARW-WRF. This land surface model (LSM) contains a single vegetation canopy layer and predicts soil temperature and volumetric soil moisture in four soil layers. The depths of these layers are sequentially from the top 0.1, 0.3, 0.6, and 1.0 m, and the root zone is contained in the upper 1 m (top three layers). The initial land surface conditions for the simulations are provided by the National Center for Atmospheric Research (NCAR) high-resolution land surface data assimilation system (HRLDAS). The HRLDAS (Chen et al. 2007) is run offline for an 18-month spinup period prior to each forecast, but using the same 3-km horizontal grid spacing that is used in the coupled atmospheric simulations to be discussed shortly. The initialization data for HRLDAS in these simulations are discussed in Trier et al. (2011).

The CIs for both IHOP cases 1 and 3 are contained in a single 24-h simulation beginning at 1200 UTC 12 June 2002. The 24-h simulation for the BAMEX case (case 2) begins at 1200 UTC 4 July 2003. Hereafter, all times are expressed in local daylight time (LT), which is UTC − 5 h.

Each simulation uses a single horizontal domain with 3-km horizontal grid spacing. However, there are differences between these model domains used in case 2 and cases 1 and 3, which are chosen primarily because of geographical differences between the IHOP and BAMEX study regions. The horizontal domain used in cases 1 and 3 has 800 × 750 grid points and is centered at 35.2°N, 101.6°W, while the horizontal domain used in case 2 has 800 × 650 grid points and is centered to the northeast at 40°N, 96°W. The vertical grid for cases 1 and 3 contains 42 levels that are vertically stretched to provide enhanced resolution within the PBL. There, the vertical spacing is $\Delta z < 100$ m, and near the top of the model at ~50 hPa it is ~1 km. The vertical grid for the case 2 simulation is very similar but contains 43 levels instead of 42.

The initial conditions for ARW-WRF for the cases 1 and 3 simulation are obtained from the National Centers for Environmental Prediction (NCEP) Environmental Data Assimilation System (EDAS) analyses. Lateral boundary conditions with a 3-h frequency are generated from the corresponding operational Eta Model. For case 2, more realistic results were obtained using the initial and lateral boundary conditions from the Rapid Update Cycle (RUC) data assimilation and modeling system (Benjamin et al. 2004) analyses, which have a horizontal grid spacing of 20 km compared to the 40-km horizontal grid spacing in the EDAS analyses used for the cases 1 and 3 simulation. Another difference is that the RUC analyses provide lateral boundary conditions with 1-h frequency.

3. Overview of convection initiation and its environment for the different cases

For case 1 (Fig. 2), late afternoon CI evolves into a broken squall line (e.g., Bluestein and Jain 1985) in both the observations and the simulation. Particularly well simulated is the initiation denoted by the arrows along the Texas Panhandle–western Oklahoma border region (Figs. 2a,c). We examine the thermodynamic destabilization indicated by trends in $B_{\text{min}}$ values in this general location indicated by the dashed rectangles in Fig. 3. For calculations here and throughout the remainder of the paper, we require $B_{\text{min}}$ to be located below $p = 500$ hPa. The reduction in $B_{\text{min}}$ magnitude to values that approach zero within the dashed rectangle by midafternoon nicely delineates where CI occurs ahead of a weak cold front during the next hour (Fig. 3b). This simulated convection becomes widespread in the south-central Texas Panhandle and western Oklahoma by several hours later (cf. Fig. 2d).

CI in case 2 occurs during the midevening over northern Iowa and extreme northeast Nebraska (Fig. 4a) and evolves rapidly into a large east–west-oriented MCS (Fig. 4b). The initial morphology and subsequent MCS evolution are reasonably well simulated (Figs. 4c,d), but the CI occurs ~2 h earlier and is located somewhat south of where it was observed (Fig. 4c). As in the surface-based convection in case 1, there is a decrease in the magnitude of $B_{\text{min}}$ prior to CI in the region where convection later develops (Fig. 5). However, the synoptic pattern is quite different from that of case 1 and consists of significant warm advection within a much stronger lower-tropospheric frontal zone. In this case the smallest $B_{\text{min}}$ magnitudes near the time of CI for air parcels originate near 2 km AGL (Fig. 5b).

The CI examined in case 3, which like case 2 consists of elevated convection, comprises northwest–southeast-oriented bands over the Texas Panhandle during the
early morning hours (Figs. 6a,c). An MCS forms over the northern Texas Panhandle and western Oklahoma during the next few hours (Figs. 6b,d) after merger of these and other bands. Mesoscale reductions in the magnitude of $B_{\text{min}}$ above the surface over a period of 2 h clearly facilitated CI for one of these bands (Fig. 7), which Marsham et al. (2011) suggest may have been directly initiated by propagating disturbances (including bores and gravity waves) trapped below in the stable nocturnal boundary layer. The mesoscale region of thermodynamic destabilization where the simulated CI in the northern Texas Panhandle occurs is situated within a zone of convergence and deformation (Fig. 7b) owing to the juxtaposition of (i) easterlies located behind a cold front to the north, (ii) southeasterlies occupying the region directly to the south that was influenced earlier by the MCS in case 1, and (iii) a southwesterly low-level jet (LLJ) located along the west Texas–eastern New Mexico border.

Note that $B_{\text{min}} \to 0$, by itself, is not a sufficient condition for CI, as evident by the sizeable areas of $B_{\text{min}} > -0.5$ that remain free of convection for the next hour in Fig. 7b. This is related to many factors including the deleterious effects on CI not accounted for in parcel theory including dry air entrainment and adverse vertical pressure gradients. We do find that CI is most likely both
where magnitudes of $B_{\text{min}}$ are small (e.g., $B_{\text{min}} > -1.5$) and its trends toward zero are regionally maximized; the latter often being an indication of favorable mesoscale forcing, to be discussed in subsequent sections.

Figure 8 shows thermodynamic quantities at the time of CI (left) and their hourly rates of change for the 2–4-h destabilization periods prior to CI (right) for 50-hPa-averaged air parcels originating from the different model vertical grid points. These variables, which include CAPE (Figs. 8a,b), CIN (Figs. 8c,d), and $B_{\text{min}}$ (Figs. 8e,f), are horizontally averaged over the CI regions depicted by the dashed rectangles in Figs. 3, 5, and 7. For CAPE, there is a clear distinction between the elevated convection cases 2 and 3 and the surface-based convection case 1, with both elevated cases having considerably smaller maximum CAPE, which in contrast to the surface-based case is found for air parcels originating at 1–2 km AGL (Fig. 8a). The favorable trends for the elevated cases are restricted to parcel source levels located significantly above the surface (Fig. 8b) in contrast to the surface-based case, where favorable CAPE trends occur throughout the PBL. The peak in positive hourly CAPE changes in the surface-based case near 1.2 km AGL reflects the increase of the PBL depth during the 4-h period during which the hourly rate of CAPE change is calculated.

The distinction between the surface-based case and the elevated cases is also pronounced for CIN and $B_{\text{min}}$, with the magnitudes of these quantities prohibitively large for CI at levels beneath ~0.5 km AGL for the elevated cases (Figs. 8c,e). Though the vertical profiles of $B_{\text{min}}$ and their rates of change for parcels from the same origination levels, can differ substantially over shallow layers when compared with their CIN counterparts, the overall similarity between these $B_{\text{min}}$ and CIN-related profiles through deeper layers lends confidence to using $B_{\text{min}}$ as a proxy for CIN.

4. Calculation of an approximate forcing budget for $B_{\text{min}}$ changes

In the previous section large simulated $B_{\text{min}}$ changes were documented prior to CI. In the current section we describe the methodology used to quantify the physical processes responsible for these changes, which are discussed in the remainder of the paper.

For each case we determine the time interval during which the majority of thermodynamic destabilization occurs using model soundings from near the CI. These time intervals, $\tau$, are those used to calculate area- and time-averaged $B_{\text{min}}$ changes in Fig. 8f, which are 4 h for case 1, and 2 h for each of the elevated cases (cases 2 and 3). All available model output times that fall inclusively within these time intervals are used to estimate the forcing for the $B_{\text{min}}$ change over these intervals. The 60-min model output frequency for IHOP simulations and 30-min frequency for BAMEX simulations results in five times for both cases 1 and 2, and three times for case 3.

The temporal variation of the vertical displacements needed to reach the altitude of $B_{\text{min}}$ is a factor that
complicates the calculation of time-averaged forcing for $B_{\text{min}}$ changes. Figure 9 shows the vertical displacements typically decreasing with time, which is due to increasing air parcel relative humidities during the destabilization process. In the absence of strong static stability (e.g., lapse rates less than the moist adiabatic) above the parcel origination level, $B_{\text{min}}$ is located near the lifted condensation level (LCL) of the air parcel (e.g., Fig. 1), which drops closer to the parcel origination level when relative humidities increase. To simplify the forcing budget for buoyancy change over the destabilization time interval, we assume that the pressure at different parcel origination levels $p = p_o$ and the pressure of their $B_{\text{min}}$ levels $p = p_B$ remain constant during the time interval and select, for $p_B$, the average depicted by the solid cyan curves in Fig. 9.

In parcel theory, local buoyancy changes result only from changes to air parcel properties at their origination level and environmental changes at levels to which they are vertically displaced. The local change in buoyancy (in temperature units) at this level for the pseudoadiabatically lifted parcel is

$$\frac{\partial B}{\partial t} \bigg|_{p_B} = \frac{\partial T_{vp}(p_B)}{\partial t} - \frac{\partial T_{vp}(p_B)}{\partial t}.$$

In (3) the parcel virtual temperature at the parcel’s lifted pressure $p = p_B$ is $T_{vp}(p_B) = F(T_o, q_o, p_o, p_B)$, where $F$
is a function that depends on the pseudoadiabatic lifting process, and \(T_0\) and \(q_o\) are the temperature and mixing ratio, respectively, at the parcel origination pressure \(p_o\).

The environmental virtual temperature at \(p < p_B\) is given by

\[
T_{ve} = T_e \left( 1 + \omega \frac{\partial T}{\partial p} \right) + \omega \frac{\partial q}{\partial p} S_q, 
\]

respectively, where we make the simplifying assumption that the temperature and moisture changes resulting from the surface temperature (\(S_H\)) and moisture (\(S_q\)) fluxes are distributed evenly throughout the PBL. This is valid when the vertical fluxes of temperature \(\bar{o}T\) and moisture \(\bar{o}q\) decrease linearly through the depth of the PBL, which is a good assumption for temperature in well-mixed boundary layers, but is often less clearly justified for moisture (e.g., Stull 1988).

Figure 10 illustrates the parcel virtual temperature changes, \(\delta T_{vp}(p_B)\), resulting from the parcel perturbations \(\delta T_0\) (Fig. 10a) and \(\delta q_o\) (Fig. 10b) being added to the parcel temperature \(T_0\) and water vapor mixing ratio \(q_o\), and the modified parcel being lifted pseudoadiabatically.

Using (4)–(6), the contributions to the parcel virtual temperature changes, \(\delta T_{vp}(p_B)\), at the lifted pressure (Fig. 10) are

\[
\text{HTADV}_P = \Delta t \left[ - \mathbf{v} \cdot \nabla_{p_o} T \right] \frac{\partial F}{\partial T_{0,q_o,p_o}}, \\
\text{VTADV}_P = \Delta t \left[ \frac{RT}{c_p p} - \frac{\partial T}{\partial p} \right] \omega \left( \frac{\partial F}{\partial T_{0,q_o,p_o}} \right), \\
\text{HMADV}_P = \Delta t \left[ - \mathbf{v} \cdot \nabla_{p_o} q \right] \frac{\partial F}{\partial q_{o,T_{0,p_o}}}, \\
\text{VMADV}_P = \Delta t \left[ \frac{\partial q}{\partial p} \right] \omega \left( \frac{\partial F}{\partial q_{o,T_{0,p_o}}} \right), \\
\text{SFLUX}_P = \Delta t \left[ S_H \right] \frac{\partial F}{\partial T_{0,q_o,p_o}} + \Delta t \left[ S_q \right] \frac{\partial F}{\partial q_{o,T_{0,p_o}}}. 
\]
and SFLUX\(_p\) in (7) are related to effects at parcel source levels of horizontal moisture advection, vertical moisture advection, and surface fluxes, respectively.

Environmental virtual temperature changes, \(\delta T_{ve|p_b}\), at the same level to which the air parcel is lifted are approximated by

\[
\delta T_{ve|p_b} = (1 + \varepsilon q_e) \left( -\mathbf{v} \cdot \nabla_{p_b} T_e + \omega \left( \frac{RT_e}{c_{p_b}} - \frac{\partial T_e}{\partial p} \right) + S_H \right) \Delta t,
\]

(8)

where we have neglected the change of moisture in the definition of \(T_{ve}\). The first two terms on the right side of (8) have a similar interpretation to their counterparts for the parcel temperature change in (7) and their contributions to \(\delta T_{ve}\) are henceforth termed \(-HTADV_E\) and \(-VTADV_E\), respectively. The third term results from the surface heat flux and will provide a contribution to \(\delta T_{ve}\), termed \(-SFLUX_E\), only when \(p_B\) is located within the PBL. Several factors can contribute to \(p_B\) being located within the PBL including boundary layers that are not perfectly well mixed (i.e., are somewhat stable) and/or the prescribed \(p_B\) being below the level at which \(B_{min}\) actually occurs. The reason for the minus signs in front of these contributors to \(\delta T_{ve}\) is that it is convenient to view the effects of environmental changes from the parcel’s perspective. For example, positive (i.e., warm)
temperature advection in the environment at the parcel lifted level will decrease the buoyancy of the parcel (i.e., make it more negative).

The total virtual temperature changes to the parcel and the environment at the level to which the parcel is lifted, $p = p_B$, are thus

$$
\delta F = \delta T_{vp} \bigg|_{p_B} = HTADV_p + VTADV_p \\
+ HMADV_p + VMADV_p + SFLUX_p \quad \text{and} \quad (9)
$$

respectively. The net time-integrated forcing ($\delta T_{vp} - \delta T_{ve}$) for $T_{vp} - T_{ve}$ changes at the parcel lifted level can be compared to actual buoyancy changes during the 2–4-h thermodynamic destabilization intervals, $\Delta \text{BUOY}$, to determine the accuracy of the budget (section 5). In upcoming discussions, the terms on the right sides of (9) and (10) that do not involve the effects of surface fluxes are collectively referred to as the grid-resolving forcing (GRES). In these derivations, for simplicity we have neglected other effects that are not grid resolved including radiative cooling.

For calculation of the budget it is important to note that the partial derivative terms in (4) and (7) are sensitive to the magnitudes of parcel perturbations $\delta T_o$ and $\delta q_o$. Specifically, when these perturbations are large, the initial parcel can become supersaturated, which would strongly impact $\delta T_{vp} \big|_{p_B}$. This sensitivity necessitates the use of a small time increment of $\Delta t = 1 \text{ min}$, when computing these finite perturbations using the instantaneous values of the forcing terms on the right side of (5) and (6) at each model output time. The resulting small-amplitude parcel temperature $\delta T_o$ and moisture $\delta q_o$ perturbations allow use of the linear approximation where the tendencies arising from different components of the parcel forcing in (9), when individually computed, collectively add to produce the tendency resulting from the total parcel forcing.

For each parcel origination level $p = p_o$, the terms on the right sides of (9) and (10), which are evaluated at the parcel lifted levels $p = p_B$, are computed for all horizontal grid points within the dashed rectangles of Figs. 3, 5, and 7. Results are horizontally averaged for each of the available model output times spanning the $\tau = 2–4$-h thermodynamic destabilization intervals, and are then time-averaged over $\tau$. Since $\tau \gg \Delta t$, these forcing results are then scaled by the factor $\tau/\Delta t$ for comparison to the buoyancy changes (section 5) occurring between the two end times that define $\tau$. A flow diagram that summarizes the primary steps in the foregoing $B_{\min}$ forcing budget calculations is found in the appendix.

5. Budget results

The evolution in the thermodynamic vertical structure during the destabilization period prior to CI is exemplified by model soundings (Fig. 11) taken at the locations of the dots within the dashed rectangles in Figs. 3, 5, and 7. Here, there are particularly large differences between the daytime surface-based CI case 1 (Fig. 11a)
and the two elevated CI cases (Figs. 11b,c). Strong surface warming and PBL deepening occur over a 4-h period in case 1 (Fig. 11a), while dramatic moistening occurs over a shorter 2-h period beginning at about 1 km AGL in both elevated cases 2 (Fig. 11b) and 3 (Fig. 11c). In the budget analyses that follow, we restrict our attention to the buoyancy changes and their time-averaged forcing for air parcels that originate within these different layers.

For each case there is good agreement between the time-averaged forcing and $\Delta\text{BUOY}$ (Fig. 12), which the forcing is expected to diagnose, as evinced by relatively small residuals (red curves) whose magnitudes range from only 0% to 20% of $\Delta\text{BUOY}$ within these layers. In case 1 (Fig. 12a), the relevant total forcing includes both GRES + SFLUX, while in the elevated cases (Figs. 12b,c) we use only GRES, which we justify by the time of
day and near-surface stabilization occurring in the soundings (Figs. 11b,c).

The buoyancy changes in \( D_{BUOY} \), which are for air parcels that were lifted to the average level that their \( B_{min} \) occurred during the destabilization period (section 4), are also a good approximation to the actual \( B_{min} \) changes, \( D_{BMIN} \) (Fig. 12). A factor that likely contributed to the good agreement between \( D_{BUOY} \) and \( D_{BMIN} \) in Fig. 12 is the fairly continuous variation during the destabilization periods of vertical displacements required to reach \( B_{min} \) from parcel origination levels (Fig. 9). Tests were conducted lifting parcels at all model output times to their \( B_{min} \) levels occurring at only the earliest and then only their

![Figure 9](image9.png)

**Fig. 9.** Area averages of the vertical displacements needed for air parcels at the origination levels indicated by the symbols to reach their level of maximum negative buoyancy for cases (a) 1, (b) 2, and (c) 3. The horizontal locations of the area averages for cases 1, 2, and 3 are indicated by the dashed rectangles in Figs. 3, 5, and 7, respectively.

![Figure 10](image10.png)

**Fig. 10.** Schematic diagram illustrating the relationships between the pseudoadiabatic process embodied in the function \( F \), and the increments \( \delta F = \delta T_{vp}(p_B), \delta T_{so}, \delta q_{so} \), defined in the text. The thin blue lines indicate the initial virtual temperature and moisture vertical structure and the thick blue lines indicate the resulting unmodified pseudoadiabatic ascent of a surface air parcel. The dashed (solid) red lines indicate the unsaturated (saturated) portion of the pseudoadiabatic ascent resulting from modifications in (a) temperature and (b) moisture, associated with the parcel forcing increments \( \delta T_{so} \) and \( \delta q_{so} \), respectively. The distances between the circles indicate the magnitudes of these changes to the parcels \( \delta F = \delta T_{vp}(p_B) \) at the parcel lifted pressure \( p_B \).
latest times (instead of their average $B_{\text{min}}$ levels for all times), and agreements between $\Delta B_{\text{BUOY}}$ and $\Delta B_{\text{MIN}}$ were not nearly as good. The readers should note that in Fig. 12 and similar displays of budget results presented later, the actual buoyancy changes and buoyancy changes predicted by the forcing are evaluated at the parcel lifted levels, which are typically 0.7–1.5 km above the parcel origination levels indicated along the y axis (cf. Fig. 9).

Encouraged by the fidelity of the budget calculations discussed above, we now explore physical processes responsible for the destabilization that contributes to CI in the individual cases using more detailed budget calculations and regional-scale vertical cross sections.

a. Daytime surface-based MCS initiation (case 1)

A line-averaged vertical cross section through the destabilizing region for case 1 (Fig. 13), whose location is shown in Fig. 3b, illustrates a 3-h time-averaged mesoscale PBL vertical circulation with ascent near the eventual CI region and subsidence maximized 200–250 km to its southeast. The circulation is associated with horizontal gradients in potential temperature and depth of the PBL at the end of the time-averaging period (Fig. 13a) that contribute to the regional differences in $B_{\text{min}}$ and CI (Fig. 3b). The maximum ascent and eventual CI occur where the PBL is warmest and deepest (Fig. 13a), and are collocated with an enhanced moisture gradient (Fig. 13b) situated slightly ahead of the cold front to its northwest. This latter finding is consistent with observations near the time of CI in this case (Weckwerth et al. 2008; Champollion et al. 2009) and the solenoidal nature of the thermally direct circulation in the current simulation is also similar, though larger in horizontal scale, than those found in previous simulations of drylines (e.g., Ziegler et al. 1997). Here, the scale and intensity differences from previous cases are likely due in part to our temporal and spatial averaging. The circulation in the current case is also consistent with effects of regional-scale land surface contrasts found in simulations over a surrounding 12-day period (Trier et al. 2008). The ascending branch of the circulation on this
particular day may have also been influenced by the cold front located to the northwest. The net forcing for buoyancy changes from the combined changes to air parcels and the environment into which they are lifted (Fig. 14a) is dominated by effects of surface fluxes and terms involving vertical motion ($\text{VERTVEL} = \text{VTADV}_P + \text{VMADV}_P + \text{VTADV}_E$). These results are consistent with the sounding evolution (Fig. 11a), which shows both PBL warming from surface heat fluxes and a weakening of the stable layer above the PBL related in part to vertical motions in the CI zone (Fig. 13a).

The combined forcing for removing parcel negative buoyancy is overwhelmingly due to changes in the parcel itself for parcels originating in the lower part of the

![Simulated and Derived Buoyancy Changes](image)

**Fig. 12.** Budgets diagnosing time- and area-averaged GRES and total forcings (GRES + SFLUX) for $\Delta$BUOY at parcel lifted levels that are the average of those of the maximum negative buoyancy for the thermodynamic destabilization period, for air parcels at the origination levels indicated by the symbols for cases (a) 1, (b) 2, and (c) 3, where the area-averaging regions are indicated by the dashed rectangles in (a) Fig. 3, (b) Fig. 5, and (c) Fig. 7. The actual area-averaged maximum negative buoyancy changes that $\Delta$BUOY approximates are indicated by $\Delta$BMIN (see text for further explanation). The budget error is indicated by the red curves.
PBL (Fig. 14b), but is substantially influenced by changes to the environment that the parcels rise into for parcels originating in the upper part of the PBL (Fig. 14c). The total forcing for changes to the parcel is closely approximated by the linear combination of individual forcing terms in which the surface flux effects dominate (Fig. 14b). The positive effect of the fluxes on decreasing the negative buoyancy of the parcel is somewhat offset by its warming effects on the environment (Fig. 14c). In contrast, the net positive effect of vertical motions on decreasing negative buoyancy (Fig. 14a) is almost entirely due to its effect on the environment (Fig. 14c) into which the parcels rise because here the associated cooling is greater than at parcel origination levels in the near-neutral PBL below.

b. Postfrontal elevated MCS initiation (case 2)

CI during case 2 occurs during the evening within a frontal zone with strong warm advection (Fig. 5b). The north–south-oriented line-averaged vertical cross section (indicated in Fig. 5b) illustrates a strong surface temperature gradient associated with the front (Figs. 15a,b) and vapor mixing ratios that are larger and extend deeper above the surface than they do outside of the frontal zone. Strengthening southerly flow approaches the surface front slightly above the surface and is maximized at ~2 km AGL for about 100 km behind its leading edge (Fig. 15b). These time-averaged southerlies contribute to the prominent warm-advection zone near the top of the frontal surface that extends northward into the region where elevated CI occurs (Fig. 15d).

In contrast to case 1 (Fig. 13a), there are two centers of environmental upward motion in case 2 (Fig. 15a). An upward motion center at the leading edge of the surface front (Fig. 15a) coincides with strong negative time-averaged horizontal moisture advection (Fig. 15d), which results in the air being too dry for CI at this location despite substantial lifting. The second time-averaged upward motion center (Fig. 15a) that originates along the upper surface of the front at the northern terminus of the time-averaged southerlies (Fig. 15b) is where CI occurs. The terminus region of a southerly LLJ is a location where conditions favor MCSs, as previously documented in numerous studies (e.g., Maddox 1983; Kane et al. 1987; Trier and Parsons 1993).

The budget analysis of net forcing for buoyancy changes in the vicinity of subsequent CI illustrates the dominance of terms involving vertical motion (Fig. 16a), with horizontal moisture and temperature advections having minor contributions in the bottom and top of the destabilizing layer, respectively. Overall, parcel changes (Fig. 16b) dominate those of the environment (Fig. 16c) in this case, because of warm-advection (HTADVp).
contributions to the parcel throughout the destabilizing layer and strong contributions from vertical moisture advection (VMADVₚ) in the upper half of this layer.

c. Prefrontal elevated MCS initiation (case 3)

Case 3 is another example of MCS initiation near the terminus of south-southwest flow located above the surface (Fig. 7b), but it differs from case 2 in several ways. This case begins near midnight and, as a time- and line-averaged cross section through the initiation region (Fig. 7b) indicates, the initiation occurs southwest (upwind) of the surface front (Fig. 17b) and outside of the immediate vicinity of significant warm advection (Fig. 17c). The surface moisture has a $>14 \text{ g kg}^{-1}$ maximum in the vicinity of the surface front, but the depth of the water vapor mixing ratio exceeding $10 \text{ g kg}^{-1}$ is greatest in the vicinity of CI (Fig. 17a). This greater depth of moisture near subsequent CI results from both upward vertical displacements of moisture associated with the time-averaged maximum vertical velocities in this region (Fig. 17a) and horizontal moisture advection (Fig. 17c). These horizontal and vertical advections of moisture have a positive effect on reducing parcel negative buoyancy through the terms HMADVₚ and VMADVₚ in the CI region (Fig. 17a) and horizontal moisture advection (Fig. 17c).

Fig. 15. Vertical cross sections of (a) vertical velocity (colors, scale below), (b) meridional winds (colors, scale below), and (c) quasigeostrophic vertical velocity (colors, scale below) each time averaged from 1730 to 1930 LT 4 Jul 2003 with potential temperature (dotted contours with 1-K intervals) and 10 and $15 \text{ g kg}^{-1}$ water vapor mixing ratio contours (thick solid black contours) near the time of CI at 1930 LT 4 Jul 2003. (d) Vertical cross section of horizontal temperature advection (colors, scale below) and horizontal moisture advection (black contours with 0.8 (g kg$^{-1}$ h$^{-1}$ contour intervals, negative values dashed), each time averaged from 1730 to 1930 LT 4 Jul 2003. The location of the cross sections is indicated by the large solid rectangle in Fig. 5b, with the line averaging performed for 250 km over the smaller (crosswise) dimension of the rectangle.
effects on decreasing negative buoyancy from temperature changes from vertical motions (VTADV_e) are not compensated by the negligible warm advection (HTADV_e). This is a major difference from case 2 (Fig. 16c) and is reflected in the much more substantial cooling of the destabilizing layer in case 3 (Fig. 11c) than in case 2 (Fig. 11b).

**FIG. 16.** As in Fig. 14 but for case 2 and with the surface-flux-related contributions to the buoyancy changes not presented (see text for explanation). The area averaging for each of these panels is performed within the black dashed rectangle shown in Fig. 5b.

**FIG. 17.** Vertical cross sections of (a) vertical velocity time averaged from 2300 LT 12 Jun to 0100 LT 13 Jun 2002 (colors, scale at right) with the water vapor mixing ratio at 0100 LT 13 Jun 2002 (dotted contours, 2 g kg$^{-1}$ contour interval) near the time of CI, (b) horizontal winds parallel to the cross-sectional time averaged from 2300 LT 12 Jun to 0100 LT 13 Jun 2002 (colors, scale at right) with potential temperature at 0100 LT 13 Jun 2002 (dotted contours with 1-K intervals) near the time of CI, and (c) horizontal temperature advection (colors, scale at right) and horizontal moisture advection (red contours with 0.8 g kg$^{-1}$ h$^{-1}$ contour intervals, negative values dashed) each time averaged from 2300 LT 12 Jun to 0100 LT 13 Jun 2002. The location of the cross sections is indicated by the large solid rectangle in Fig. 7b, with the line averaging performed for 250 km over the smaller (crosswise) dimension of the rectangle.
Comparison of the budget calculations among the cases has illuminated several aspects of the thermodynamic destabilization process. A variety of different forcing mechanisms can dominate the reduction of negative buoyancy, and which mechanisms are most relevant can vary with altitude within the same destabilizing layer. For the case of daytime surface-based CI, destabilization in the PBL results from a combination of surface heating and moistening and vertical motion. In the two elevated convection cases vertical motion effects dominate horizontal advections, including horizontal temperature advections, in reducing negative buoyancy.

For temperature advection to have a large direct impact on thermodynamic destabilization, differential advection, \( \partial / \partial z (-v \cdot \nabla_p T) < 0 \), must occur. However, the vertical motion, which dominates the destabilization, may itself be influenced by warm advection when the temperature advection has a significant geostrophic component. For case 2, quasigeostrophic \( \omega \) was calculated by solving its \( \mathbf{Q} \)-vector form [Holton (1992), Eq. (6.35)] subject to a lower boundary condition of \( w = 0 \). The corresponding vertical velocity, \( w = -\omega / \rho g \), has multiple updraft centers (Fig. 15c) with amplitude about \( 1/2 \) that of the total \( w \) (Fig. 15a). Though the quasigeostrophic \( \omega \) centers (Fig. 15c) are maximized about 0.5–1.0 km above the warm-temperature advection centers (Fig. 15d), their very similar horizontal locations suggest an important possible linkage since from quasigeostrophic theory \( \nabla^2 w \sim -\mathbf{Q} \cdot (\nabla \phi / \rho g) \) and, thus, \( w \sim -\nabla \phi \cdot \mathbf{V} \).

A convenient and physically illuminating aspect of our thermodynamic destabilization analysis methodology is the ability to isolate the effects of changes in air parcel properties at their origination level from changes to the environment near the level of parcel minimum buoyancy. These results are summarized for the destabilizing layers of the three cases in Fig. 19. Here, the squares indicate the parcel source levels and the triangles indicate the lifted parcel levels, which approximate the level of the minimum buoyancy of the lifted parcel. The tilts of the lines connecting these symbols indicate whether changes to the parcel (leftward with height) or the environment above (rightward with height) dominate the destabilization process. For the daytime surface-based CI case (Fig. 19a), the tilt changes reflect a shift in the causes dominating negative buoyancy reduction from surface fluxes increasing moist static energy for lower PBL parcels to vertical motions cooling the environment above upper PBL parcel origination levels.

The differences in the tilts between the elevated cases at all parcel origination levels (Figs. 19b,c) are attributed to differences in warm-advection strength. For case 2 (Fig. 19b), parcels are warmed substantially by advection at their origination levels. However, the deep layer through which the warm advection occurs partially cancels the cooling effect of environmental vertical motion above the parcel origination levels, thereby increasing the negative buoyancy the parcels experience above their source level. In case 3 (Fig. 19c) with negligible warm

**Fig. 18.** As in Fig. 14 but for case 3, and with the surface-flux-related contributions to the buoyancy changes not presented (see text for explanation). The area averaging for each of these panels is performed within the black dashed rectangle shown in Fig. 7b.
advection, parcels undergo less substantial warming both at their source levels and above, and environmental vertical motion dominates the reduction in negative buoyancy.

6. Summary and conclusions

The primary motivation for this paper is to introduce a framework within which the importance of different thermodynamic processes to convection initiation (CI) in numerical models can be quantified in a way amenable to simple interpretation. Use of the method is demonstrated with convection-permitting simulations of several different MCSs, where the character of CI varies from afternoon surface-based initiation to nocturnal initiations, in which the air responsible for CI is elevated, originating a few kilometers above the surface.

Surface fluxes warming and deepening the PBL and cooling resulting from upward motion above the PBL contribute synergistically in the daytime case of simulated surface-based CI, which occurs near a surface cold front and dryline. For the nocturnal cases of elevated convection, the effects of vertical motions dominate the thermodynamic destabilization leading to CI. However, in one elevated MCS case that occurs within a strong frontal zone, the vertical motions appear at least partly tied to horizontal temperature advections.

Though the three simulated MCS cases are substantially different, they were selected primarily for demonstration purposes, and are unable to encompass the wide range of possible MCS initiation environments. Common MCS initiation environments that a similar analysis approach could be applied to in future work include, among others, topographically influenced flows (e.g., Wulfmeyer et al. 2011; Davis and Lee 2012) and mesoscale convective vortices (MCVs, e.g., Raymond and Jiang 1990; Trier and Davis 2007; Schumacher and Johnson 2009).

The method we use for assessing thermodynamic destabilization is based on the well-known and widely applied parcel theory of convection. Central to the method is the quantity $B_{\text{min}}$, which is the buoyancy minimum for an air parcel lifted from a specified height. CI occurs for the three simulated cases over mesoscale regions having large increases in $B_{\text{min}}$ (i.e., reductions in negative buoyancy) and within $\sim 1$ h of $B_{\text{min}}$ reaching values of about $-1^\circ$C. In these three cases, both $B_{\text{min}}$ and its temporal rate of change have area-averaged vertical profiles (where the vertical coordinate is the parcel origination level) that are qualitatively similar to counterparts using the CIN (2), which is the vertically integrated negative buoyancy.

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**Fig. 19.** Area-averaged buoyancy changes at air parcel lifted levels (approximating the level of parcel maximum negative buoyancy) related to forcing for changes to the parcel at its origination level (indicated by the squares) and to changes to the environment at the parcel lifted level (indicated by the triangles) for (a) case 1 from 1100 to 1500 LT 12 Jun 2002, (b) case 2 from 1730 to 1930 LT 4 Jul 2003, and (c) case 3 from 2300 LT 12 Jun to 0100 LT 13 Jun 2002. The different colors of the lines highlight the different parcel origination levels. The area-averaging regions for (a), (b), and (c) are denoted by the black dashed rectangles in Figs. 3b, 5b, and 7b, respectively.
Thermodynamic analysis using \( B_{\text{min}} \) is intended to supplement, rather than replace, the evaluation of the integral quantity CIN to anticipate CI. Since \( B_{\text{min}} \) values are indicative of parcel and environmental temperature differences at a single vertical level, they are exposed to sampling errors that CIN is less sensitive to. While this issue is not problematic for the current simulations (cf. Figs. 8c,e and 8d,f), it could be significant when analyzing high-resolution observational soundings. Though such errors may be mitigated somewhat by the customary vertical averaging through nominal parcel depths (e.g., 50 hPa), as has been done in this study, and perhaps surrounding the environmental level, \( B_{\text{min}} \) can at best only approximate the CIN.

While instantaneous \( B_{\text{min}} \), by itself, provides less complete information than CIN for determining the likelihood of CI, the conceptual simplicity of \( B_{\text{min}} \) may be exploited for a more straightforward diagnosis of thermodynamic destabilization. The primary advantage of using \( B_{\text{min}} \) lies in the relative ease in calculating a forcing budget for local changes in \( B_{\text{min}} \) compared to CIN, which requires consideration of the impacts of thermodynamic forcing on the environment at multiple levels that the modified air parcel is rising into. The interpretation of the \( B_{\text{min}} \) forcing is also simpler since the character of this forcing can change through the depth of CIN layers.

Another attractive aspect of \( B_{\text{min}} \) is that unlike CIN, it does not require there to be CAPE and is, thus, continuous. Assessing thermodynamic destabilization rates, which are often important to the forecasting of CI, is more meaningful using \( B_{\text{min}} \) than CIN in circumstances where the environment is initially conditionally stable and CIN is thus undefined. Future investigations with a larger sample of cases are needed to determine the utility of \( B_{\text{min}} \) trends in both models and observations for improving forecasts of CI.

For the simulated cases, CIN beneath the LFC is the most important obstacle to overcome prior to CI. However, in other cases insufficient CAPE may be a more significant limiting factor. The method described herein is easily modified to analyze physical factors influencing changes in the magnitude of conditional instability. There, the same parcel origination levels could be retained, and the altitude of the \( B_{\text{min}} \) level could simply be replaced by the level of the maximum parcel buoyancy \( B_{\text{max}} \) as the choice for the environmental level to be analyzed.

In the current paper we have analyzed mesoscale thermodynamic influences on CI. However, regional changes in thermodynamic conditions can influence organized convection throughout its life cycle. In Part II of this paper (Trier et al. 2014), we use approaches similar to those described here to assess mechanisms for thermodynamic variations in composite environments of simulated mature MCSs.

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**APPENDIX**

**Sequence of Steps in the Calculation of an Approximate \( B_{\text{min}} \) Budget**

1. Determine thermodynamic destabilization time interval, \( \tau \), from model soundings in proximity to CI.
2. Calculate minimum parcel buoyancy \( B_{\text{min}} \) beneath 500 hPa and the pressure at which it occurs for all output times during \( \tau \).
3. Calculate buoyancy (BUOY) at lifted parcel pressure level \( P_{\text{a}} \), which is the average pressure level of \( B_{\text{min}} \) during \( \tau \), and calculate \( \Delta\text{BUOY} \) (approximating \( \Delta B_{\text{min}} \)) at endpoints of \( \tau \).
4. Calculate \( \delta T e \) (environment component of forcing for \( \Delta\text{BUOY} \)) at pressure level \( P_{\text{a}} \) for all output times during \( \tau \).
5. Calculate \( \delta T p \) (parcel component of forcing for \( \Delta\text{BUOY} \)) by comparing unmodified lifted parcel \( T_e \) at \( P = P_{\text{a}} \) with that of the lifted parcel modified by \( \delta T e \) and \( \delta p \) for all output times during \( \tau \).
6. Scale \( \delta T p \) terms by a factor of \( \tau/\Delta t \) to obtain parcel forcing consistent with the time interval over which \( \Delta\text{BUOY} \) occurs.
7. Time-average \( \delta T e \), scaled \( \delta T e \) and associated components over \( \tau \). Area-average \( \Delta\text{BUOY} \), and time-averages of \( \delta T e \), scaled \( \delta T p \) and resulting net forcing (\( \delta T p \cdot \delta T e \)) for \( \Delta\text{BUOY} \).

**FIG. A1.** Flow diagram illustrating calculations for the budget approximating changes in the minimum parcel buoyancy \( \Delta B_{\text{min}} \) during the 2–4-h thermodynamic destabilization period \( \tau \) (see section 4 of text for details).
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