Evaluation and Application of Conditional Symmetric Instability, Equivalent Potential Vorticity, and Frontogenetic Forcing in an Operational Forecast Environment

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ABSTRACT

The related fields of equivalent potential temperature ($\theta_e$), geostrophic momentum ($M_g$), equivalent potential vorticity (EPV), and frontogenetic forcing were computed, analyzed, and evaluated for two mid-Atlantic states snowstorms in which conditional symmetric instability (CSI) is believed to have been present. Observed data as well as model gridded data were used with the National Weather Service PC-based Gridded Information Display and Diagnostic System in evaluating the event, thereby providing both initial and model forecast fields. A brief physical and historical review of CSI is provided as a basis for better understanding and increased proper application of this technique into operational forecasting. The implications of CSI for the operational forecaster are then demonstrated through a diagnosis of events occurring in Maryland and Virginia on 26 February 1993 and 30 January 1995, in which precipitation banding related to frontogenetic forcing and CSI are believed to have occurred. The intent here is to demonstrate that it is possible to improve short-range forecasts of CSI-related precipitation in an operational forecast environment through improved techniques for the recognition and evaluation of CSI via model-derived forecast fields, augmented with WSR-88D radar and other new observing systems.

Vertical cross sections of forecast $M_g$, $\theta_e$, layer mean geostrophic frontogenesis, and EPV fields were constructed and revealed regions of conditional symmetric instability and frontogenetic forcing that were nearly coincident with the observed enhanced snow bands. The availability of gridded model output and the increased capability for quickly manipulating those datasets in real time have now enhanced the potential for improved forecasts of CSI-related phenomena in operational forecast settings.

1. Introduction

Conditional symmetric instability (CSI) is an atmospheric phenomenon in which a combination of gravitational and inertial forces are acting simultaneously (static instability and inertial instability). In this way a parcel that is stable with respect to vertical forces (gravitational, convective) and stable to horizontal forces (inertial) may nevertheless be unstable to displacement in a slantwise direction. The term conditional refers to an atmosphere near saturation. The result is enhanced precipitation oriented in narrow bands, sometimes embedded within a larger precipitation area. The relationship between CSI and equivalent potential vorticity (EPV) is explained in detail by Snook (1992) and Moore and Lambert (1993). Snook (1992) states that an equivalent criterion for CSI in a saturated environment is that the observed EPV be less than zero, that is, the absolute vorticity on a surface of equivalent potential temperature ($\theta_e$) be negative. The slantwise displacement that results from this instability is an adjustment process to restore atmospheric neutrality. CSI has serious implications for precipitation forecasting but has yet to enjoy wide understanding within the operational forecast community. Many operational forecasters have observed the result of CSI in the form of enhanced banded precipitation, and while some have suspected it as being the result of CSI, few have had the tools and techniques available to them for a thorough understanding and diagnosis of the processes involved. In a real-time operational forecast environment, it is imperative that these techniques be quick and easy to facilitate the forecast process and conform to strict time constraints. For the most part, this has not yet been the case concerning CSI, and until recently no model-derived forecasts of CSI or EPV have been available.

The evaluation of CSI as a regular part of operations is in about the same evolutionary time period as the analysis of $Q$ vectors (Barnes 1985) was around the middle 1980s. $Q$-vector diagnosis as a method for estimating quasigeostrophic forcing had been a topic of study and discussion in the literature since the late 1970s (see Trenberth 1978) and was available in limited diagnostic programs in the operational community by the late 1980s (see Foster 1988). It was not widely used,
however, in the operational community until these fields became available in model-derived forecasts via advanced workstations such as the prototype National Weather Service (NWS) Advanced Weather Interactive Processing System, PC-based graphics programs such as PCGRIDDDS (PC-based Gridded Information Display and Diagnostic System; Petersen 1993), or the GEM-PAK software package for workstations (desJardins et al. 1991). As is now often evident in State Forecast Discussions (SFDs), the analysis of \( \mathbf{Q} \)-vector divergence has become a routine part of the forecast decision process in many NWS forecast offices. Similarly, it is hoped that improved techniques for the diagnosis and prediction of other mesoscale phenomena such as CSI, along with improved access to datasets and gridded model output, will encourage increased application of these techniques into routine operations.

To lay a foundation, we will begin with a brief historical perspective, a discussion of the nature of CSI, implications for forecasters, and a report on the current status of CSI applications in operations. The methodologies and mechanics for calculating CSI, EPV, and frontogenetic and \( \mathbf{Q} \)-vector forcing, as well as geostrophic momentum \((M_g)\) and \(\theta_e\) surfaces via PCGRIDDDS, are described in the appendix. The paper follows with a case study of the 26 February 1993 and 30 January 1995 events to demonstrate how the recognition of CSI and improved real-time observing systems such as the Weather Surveillance Radar-1988 Doppler (WSR-88D) might have aided short-term forecasts. The article closes with a discussion of the results and the methods and procedures for more routine analysis and prediction of CSI and other phenomena within the operational forecasting environment.

2. Historical perspective

CSI (also known as moist slantwise convection) has been a topic of much discussion and study over the years as a possible mechanism for banded precipitation in extratropical cyclones. The fact that precipitation is often organized in bands has been recognized since the early part of this century (Bjerknes 1919). More recently, Bennetts and Hoskins (1979) suggested that the bands may be a manifestation of symmetric baroclinic instability. They stated that the bands take the form of rolls oriented along the thermal wind, with their motion then generating conditional gravitational (convective) instability at midlevels in preferred linear regions. The effect of latent heat release during ascent was included in this theory, leading to the concept of “conditional symmetric instability,” with the term conditional indicative of an atmosphere at or near saturation.

Emanuel (1979, 1983a, 1985) expanded on this by demonstrating that circulations associated with CSI are fundamentally mesoscale in nature and are a combination of gravitational (convective) forces acting vertically and inertial (centripetal) forces acting horizontally to give a slantwise displacement. Emanuel (1983a) demonstrated that these combined forces can be estimated from a single atmospheric sounding and by considering cross sections of pseudoangular momentum and equivalent potential temperature. He also argued that frontogenesis and vertical motion are sharply enhanced in narrow regions where equivalent potential vorticity decreases toward zero in a saturated frontal zone.

Since these and other foundation studies, research into the nature of CSI has accelerated in recent years with particular attention to case studies addressing the problem of recognition and application of CSI in operational forecasts. Howard and Tollerud (1988), Moore and Blakley (1988), Dunn (1988), Lussky (1989), Tollerud et al. (1991), Colman (1992), Snook (1992), Conger and Dunn (1993), Moore and Lambert (1993), Reuter and Yau (1993), Grumm and Forbes (1994), and McCann (1995) have all presented studies exploring the nature of CSI and techniques for assessing CSI based upon analyses of observed data, as well as its potential impact on forecast operations.

A logical next step in CSI assessment, as recognized by Lussky (1989), is to evaluate CSI by using prognostic data from operational numerical models. Some initial attempts at this have shown promise (e.g., Grumm and Forbes 1994). However, the conventional method of estimating locations of CSI in a vertical cross section is rather cumbersome, whereby one estimates the relative slopes of \(\theta_e\) surfaces to \(M_g\) surfaces. Moore and Lambert (1993) were first to show the relationship between CSI and EPV and its practical application in operational forecasting. Their method thus provides an objective, quantitative, and simple technique for quickly and effectively diagnosing CSI in an operational forecast environment. Use of EPV facilitates the evaluation of CSI in real-time operational forecasting when used in conjunction with conventional cross sections of \(\theta_e\) and \(M_g\).

We extend the work of Moore and Lambert by incorporating the technique for computing EPV into PCGRIDDDS and applying it to gridded model data, thereby providing model-derived forecast fields of EPV (see the appendix). Of related interest is recent work by McCann (1995) that describes a technique to examine EPV in plan view (i.e., quasi-horizontal) on pressure surfaces computed from forecast gridded model data. Although this technique promises a potentially more expedient assessment of CSI over large geographic regions, the choice of a single pressure level divulges no information concerning the vertical distribution of CSI and has not been employed here.

3. CSI: Implications for operational forecasters

The basic theory of CSI is illustrated schematically in Fig. 1. This shows the resultant accelerations (double arrows) when displacing a parcel horizontally (point A), vertically (point B), or in a slantwise direction (point C) as in the case of CSI. This figure, which is explained
in greater detail in the appendix, shows that a slantwise displacement from point C results in an acceleration in the same direction as the displacement. Displacement from either point A or point B, on the other hand, results in accelerations that quickly act to restore the parcel to its initial position.

J. T. Moore (1993, personal communication) has summarized the synoptic setting that is characteristic of CSI, and conventional techniques for evaluating CSI.

- CSI occurs in regions in which the vertical wind profile is increasing and veering with height indicative of baroclinicity (shear). Moderate to strong vertical shear on the order of 10–20 m s<sup>-1</sup> over a vertical span of 1–2 km in the lower to middle troposphere should be present for CSI to occur.
- CSI occurs in the presence of low static stability in the middle troposphere and a statically stable air mass within the boundary layer. This typically occurs north of a surface frontal boundary. Thus, a thermodynamic profile that is nearly saturated and close to the moist adiabatic lapse rate at lower to middle levels is necessary (i.e., a parcel is neutral or stable with respect to moist ascent). The lifted parcel must be nearly saturated within the region of symmetric instability, typically occurring between 800 and 500 hPa in the central and eastern United States. The rawinsonde observation discussed in section 5 is typical of a CSI environment.
- CSI often occurs near a warm front and ahead of a large-scale upper trough in regions of strong, moist southwesterly midtropospheric flow, where the above conditions are typically satisfied. In these regions, the atmosphere has a disposition for weak large-scale ascent, which is important for saturating the air mass. CSI is then released by parcel perturbations, of which the slantwise displacements are unstable. These regions are also characterized by relatively low values of absolute vorticity. In the mid-Atlantic region, CSI often appears in association with cold-air damming situations (Grumm and Forbes 1994).
- CSI can be evaluated by constructing a vertical cross section (oriented normal to the thickness contours) of $M_z$ and $\theta_e$. CSI theoretically exists where the slope of the isentropes ($\partial M_z/\partial \theta_e$) is steeper (more vertical) than the isopleths of momentum. In practice, however, CSI may be suspected even in areas where the surfaces are nearly parallel, indicating only weak conditional symmetric instability or conditional symmetric neutrality conditions (Howard and Tollerud 1988).

If any of these factors are lacking, the atmosphere’s potential for CSI is lessened. From these it can be seen that CSI may be operating in an otherwise seemingly innocuous synoptic setting and may contribute to unexpectedly large precipitation amounts. Stronger vertical ascent, like that found ahead of migratory extratropical cyclones, tends to more uniformly mix the mass and momentum fields in the troposphere, thus reducing the vertical shear and inertial instability. In addition, upper-level absolute vorticity will often be weak and even close to zero (anticyclonic wind shear) in areas where CSI is possible. Bluestein (1986) has pointed out that CSI is nearly always found on the anticyclonic shear side of an upper-level jet streak since this is where inertial instability is greatest.

4. Assessing CSI in operational forecasting

As mentioned earlier, conventional techniques for assessing CSI, through use of vertical cross sections of $M_z$ and $\theta_e$ surfaces, are rather cumbersome and subjective, and not conducive to widespread use within the operational forecast community where speed and ease of use are crucial. Moreover, it has been shown that new diagnostic programs do not gain wide acceptance and routine application by operational forecasters until they become applicable to model-derived prognostic fields (Dunn 1991; E. Thaler 1995, personal communication). It is for this reason, we believe, that CSI has remained mostly within the realm of the research and academic community, while only penetrating the “outer fringes” of operational forecasting thus far. It is, however, the very nature of CSI that has also kept it on the fringe. Knowing that the atmosphere is conducive to CSI is one thing, but knowing where the bands will then form remains one of the most difficult and challenging of forecast problems. All the forecaster can do is wait for the bands to form on radar and then react with updated forecasts. Contributing to the problem is the fact that some forecasters may simply be using CSI improperly. Forecasters may, for example, confuse CSI with what is really convective instability and/or frontal bands. Thus knowing when and how to use CSI/EPV in an
operational environment is crucial to its successful application.

Three events have recently occurred, however, that may lead to wider and increased attention to CSI in routine operational forecasting: 1) gridded model data are widely available, 2) PCGRIDDS and other interactive software packages are being installed in forecast offices for diagnosis and manipulation of the model data in real time, and 3) techniques for evaluating and displaying CSI/EPV graphically are now available.

It should be noted, however, that by evaluating only equivalent potential vorticity as a means for assessing CSI potential can be misleading in some cases. Bennetts and Sharp (1982), Colman (1992), and Moore and Lambert (1993) have shown that an atmosphere can be unstable to both CSI and convective instability. Because upright convection has a growth rate that is much faster than the inertial instability of CSI, it predominates in such situations. If the atmosphere is convectively unstable, buoyancy forces will always dominate any inertial forcing. The flow in the atmosphere may develop a CSI-induced circulation, but may evolve, through differential advective effects of mass and thermal properties, into a region of convective instability, whereby the gravitational (convective) instabilities would dominate. It is necessary, therefore, to differentiate between those situations in which potential energy is released through convective instability and those unstable to CSI. One cannot dispense with looking at the respective slopes of $M_e$ and $\theta_e$ in a cross section, since it is the vertical distribution of $\theta_e$ that differentiates between convective and inertial instability. In areas of the cross section where CSI is suggested but in which $\theta_e$ is decreasing or constant with height, convective instability will then dominate.

The PC-based software package PCGRIDDS (Petersen 1993) is used to compute and display $M_e$, $\theta_e$, and scalar frontogenesis fields. Other meteorological software packages exist, such as GEMPAK (desJardins 1991), that can also compute and display these fields, although it was more expedient to use PCGRIDDS for this analysis. (See the appendix for details on the PCGRIDDS commands used to compute $M_e$ and EPV.)

The gridded model data used for this study are from models run at the National Centers for Environmental Prediction (NCEP, formerly the National Meteorological Center). Data from the operational NCEP Regional Analysis and Forecast System (RAFS) are used to assess basic synoptic-scale parameters. In addition, gridded data are also used from an experimental, high spatial resolution (approximately 40-km grid spacing with 39 vertical levels) version of NCEP’s Eta coordinate model (Black 1994). This model will be referred to as the “mesoscale” Eta or simply “meso-Eta” for this article. NCEP continues research and development on the meso-Eta using an experimental version with enhanced vertical and spatial resolution and improvements to the model physics and numerics.

CSI is evaluated by using spatial cross sections oriented normal to the average geostrophic thermal wind (i.e., geopotential thickness contours) in the troposphere. Contoured fields of $\theta_e$ and $M_e$ are then examined. The layer between 1000 and 500 hPa is used to compute a thermal wind. For all spatial cross sections, the left portion of the cross section is located in relatively cold air, with warm air to the right. The contours of $\theta_e$ and $M_e$ depicted in the cross sections are referred to as “surfaces.” Similar to Sanders and Bosart (1985), the potential for CSI involves comparison of the relative slopes of $\theta_e$ surfaces versus those of $M_e$ surfaces. The criteria for a moist atmosphere was determined by examining cross sections of relative humidity (RH). An RH value of about 80% or greater is considered as meeting the near saturation “condition” necessary for CSI.

As shown by Moore and Lambert (1993), CSI also can be evaluated by examining fields of EPV in cross sections oriented normal to the geostrophic thermal wind. The equation of EPV follows Martin et al. (1992):

$$EPV = -\mathbf{\eta} \cdot \nabla \theta_e,$$

where $\mathbf{\eta}$ is the three-dimensional vorticity vector and $\nabla$ is the gradient (“del”) operator in $x$, $y$, and $p$ coordinates. The prime denotes the full three-dimensional form of EPV (i.e., includes both horizontal and vertical motions).

Moore and Lambert (1993) expand (1) by assuming geostrophic flow, neglecting terms with $\omega$ (i.e., vertical motion in pressure coordinates), terms involving changes with respect to $y$ (where $y$ is parallel to the thermal wind and $x$ is orthogonal to $y$), and multiplying by the gravitational acceleration ($g$) to get

$$EPV = g \left[ \frac{\partial M_e}{\partial p} \frac{\partial \theta_e}{\partial x} - \frac{\partial M_e}{\partial x} \frac{\partial \theta_e}{\partial p} \right].$$

Multiplying (2) by $g$ gives EPV in units of $10^{-6}$ m$^2$ K s$^{-1}$ kg$^{-1}$, or potential vorticity units (PVU). The notation for terms 1 and 2 follows that found in Moore and Lambert (1993). Wherever EPV is either zero or negative, and the atmosphere is nearly saturated, then the atmosphere is considered to have potential for CSI. In (2), CSI occurs whenever term 1 dominates term 2 since both terms 1 and 2 are usually negative.

CSI in a saturated atmosphere may be manifest as elongated banded precipitation structures on radar. These bands tend to be oriented nearly parallel to the thermal wind in the lower troposphere (e.g., 950–400-hPa layer). From previous studies (e.g., Emanuel 1985; Snook 1992; Colman 1992; Grumm and Forbes 1994), the horizontal scale of CSI-induced bands typically ranges from 50 to 100 km wide and from 75 to 400 km long. Bands typically last about 1–3 h in duration locally. The vertical slantwise motions associated with CSI may typically be up to 10 m s$^{-1}$, with the slope of
the CSI-induced slantwise circulation being about 1:100. The bands exhibit a quasi-stationary nature with a linear appearance on radar.

It is precisely because of the improved observing capabilities of Doppler radar and satellite that the sometimes subtle banded precipitation structure can be observed at all. Similarly, the advent of interactive software packages make the operational computation of EPV and other fields practical, as will be shown in the following case studies. This would not have been possible in the operational environment of the not-too-distant past.

5. 26 February 1993

Light precipitation on 26 February 1993 fell over a large portion of the mid-Atlantic and southeastern states. The precipitation fell mainly as snow across Maryland, the eastern panhandle of West Virginia, and the northern half of Virginia (Fig. 2). A mix of snow and rain occurred across the southern half of Virginia, with mostly rain reported farther south.

Official NWS forecasts early on the 26 February issued winter storm warnings [by definition 10 cm (4 in.) of snow expected within 12 h or 15 cm (6 in.) within 24 h] for areas across southern and central portions of Virginia and extreme southern Maryland. Snow advisories for lesser amounts were in effect across the northern portion of Virginia, the eastern panhandle of West Virginia, and most of Maryland. Within this area of light snow, locally moderate to heavy snow amounts fell in a concentrated area across portions of north-central and western Maryland, the eastern panhandle of West Virginia, and northern Virginia (see Fig. 2). The event resulted in 10–20 cm (4–8 in.) of snow in a 12-h period between 0600 and 1800 UTC 26 February (heaviest snowfall between 1000 and 1600 UTC). The meso-beta-scale nature of the event was consistent with the spatial dimensions of CSI-related multiple bands as given by Bennetts and Hoskins (1979) and more recently by Grumm and Forbes (1994).

Aided by WSR-88D radar, NWS forecasters were able to identify the banded precipitation structure and quickly respond with updated and detailed short-range forecasts for locally heavy snow. Figure 3 shows composite reflectivity from the Sterling, Virginia, WSR-88D radar, which revealed the banded nature of precipitation during the event. Specific reference to CSI potential, banded precipitation structure, and the resultant locally heavy snowfall was given in the early morning SFD at the NWS Forecast Office in Sterling, Virginia, and then applied with issuance of an updated winter storm warning for the area. Examining the structure and intensity of snowfall in such detail had rarely been possible before the installation of WSR-88D radars in NWS Forecast Offices. Conventional techniques for evaluating CSI are now presented and compared for this event. In addition, calculations of EPV are examined with the purpose of investigating ways in which this technique may have aided short-term forecasts in this and other similar situations.
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Fig. 3. Composite reflectivity from the NWS WSR-88D radar (KLWX) at Sterling, Virginia, for 1453 UTC 26 February 1993. Blue shaded colors represent reflectivity from 5 to 20 dBZ, and green shades from 20 to 35 dBZ.

a. Synoptic setting

The synoptic-scale pattern during the event featured westerly midtropospheric flow through much of the central and eastern United States (Fig. 4). A weak shortwave disturbance was propagating eastward through the mid-Mississippi and Ohio River Valleys at 1200 UTC 26 February 1993. Southwesterly flow and warm advection was observed at the 850-hPa level over the mid-Atlantic states (Fig. 5). A surface reflection of the upper-level shortwave disturbance in the form of an inverted trough was located over Kentucky (Fig. 6). Surface high pressure was centered over New England with a ridge extending southwest through Virginia and the western Carolinas. The ridge resembled a surface pattern characteristic of cold-air damming (Richwein 1980) with moist northeast to east surface flow along the eastern slopes of the Appalachians.

The overall synoptic pattern appeared to satisfy those conditions conducive to the development of CSI as previously outlined by J. T. Moore (1993, personal communication) and presented in section 3. This was evident in the 1200 UTC 26 February 1993 sounding for Dulles International Airport (Fig. 7a), which shows a nearly saturated profile with a moist adiabatic lapse rate above 600 hPa, as well as veering and increasing winds with height indicative of baroclinicity. The vertical cross section of wind is better illustrated by the WSR-88D velocity azimuth display (Fig. 7b), which more clearly showed the veering and increasing winds with height before and during the event. Another important synoptic precondition for CSI was the strong west to southwesterly flow aloft ahead of an upper-level trough (Fig. 4). This resulted in weak large-scale vertical ascent over the region. No well-defined surface warm frontal boundary was evident, but the 850-hPa analysis (Fig. 5) shows a moderately strong baroclinic zone lying west to east across Maryland and Virginia within a warm advection region.
b. The relationship of CSI and frontogenetic forcing in the 26 February 1993 case

Sanders and Bosart (1985) recognized that frontogenetic ascent plays an important role in saturating the atmosphere, a necessary condition for CSI. Emanuel (1985), Sanders and Bosart (1985), and Sanders (1986) demonstrated that the role of low symmetric stability was then to enhance the effects of large-scale frontogenetical forcing by increasing the intensity of vertical motion while decreasing the horizontal scale of that motion. Similarly, Gyakum (1987) suggested that the frontogenetic effects of a thermally indirect circulation might act synergistically with the slantwise instability to produce locally strong bands of precipitation. Thus, he concluded, CSI appears to enhance the intensity of the frontal circulation while reducing the scale of horizontal motion.

Additionally, Grumm and Forbes (1994) show that precipitation banding associated with CSI may occur in conjunction with jet streak entrance regions and low-level cold-air damming common to the eastern slopes of the Appalachian Mountains in the mid-Atlantic region of the United States. The inertial instability and ageostrophic transverse vertical circulations associated with these jet streaks enhances lower-tropospheric cold-air advection, thereby reinforcing low-level damming and frontogenetic forcing. Frontogenetic forcing will increase the geostrophic vertical wind shear as the thermal wind responds to the enhanced thermal gradient. This, in turn, will make the $M_g$ surfaces flatter, thus making the region more prone to CSI.

In the 26 February 1993 case it is believed precipitation amounts across central Maryland and northern Virginia were likely enhanced due to the combined effects of frontogenetic forcing and CSI. In Fig. 8, the 12-h meso-Eta forecast of 700–600-hPa layer-averaged geostrophic frontogenesis depicts the best area for frontogenetic forcing is found extending in a band across northern and central Maryland, nearly coincident with the area of heaviest snowfall. Enhanced upward vertical motion on a frontal scale would be expected on the south (warm) side of the elongated east–west axis of geostrophic frontogenesis over central Maryland, which would aid in saturating the atmosphere.

This area of frontogenesis was likely enhanced by a southwest–northeast-oriented jet streak as revealed in the meso-Eta Model forecast. The 12-h forecast showed a small jet streak with a core of winds of 33–36 ms$^{-1}$ (65–70 kt) between about 700 and 500 hPa moving from extreme southwestern Virginia at 0600 UTC (not shown) to over northern Virginia and central Maryland at 1200 UTC 26 February (Fig. 9). As described by Grumm and Forbes (1994), ageostrophic circulations, upper-level divergence, and midlevel convergence associated with the jet streak may have acted to enhance low-level cold-air damming and frontogenetic forcing. The vertical cross section in Fig. 10 depicts the ageostrophic transverse vertical circulation associated with this jet streak. The cross section was constructed along the north–south plane shown in Fig. 11a. It reveals the ascending branch of a thermally indirect circulation as well as the northerly low-level ageostrophic wind com-
ponent. Snowfall was maximized in an east–west band between the region of maximum frontogenetic forcing and the ascending branch.

c. Evaluating CSI and EPV

Specific techniques for evaluating CSI and EPV are now presented. The PCGRIDDS graphics routines were used to construct a vertical cross section of $M_g$ and $\theta_e$ (Fig. 11a) from a point east of the coastal Carolinas (point A) to northern Pennsylvania (point B) oriented normal to the 950–400-hPa thickness contours. The resultant cross section depicted in Fig. 11b, derived from the meso-Eta Model initialized at 0000 UTC 26 February 1993, is valid for 1200 UTC 26 February. The cross section reveals CSI in the region of northern Vir-
Virginia and Maryland between 600 and 800 hPa (shaded area between points C and D). This region closely corresponds to an area of convectively stable conditions over northern Maryland as evident by the vertical distribution of $\theta_e$ below 600 hPa. This suggests convective instability is an unlikely mechanism in the enhanced precipitation banding. The area of symmetric instability is also nearly coincident with the region of frontogenetic forcing previously illustrated in Fig. 8, as well as model-derived fields of vertical motion and $Q$-vector convergence (not shown). These illustrate a concentrated vertical motion field within a region of possible CSI. In this situation, one might expect the development of enhanced banded precipitation.

A vertical cross section of EPV (Fig. 12) was constructed in the same plane as in Fig. 11a using the same meso-Eta 12-h forecast initialized at 0000 UTC 26 February. Figure 12 reveals regions of negative EPV (shaded areas) that denote potential regions of CSI. However, the negative area in the lower-right portion of Fig. 12 is also an area where the atmosphere is convectively unstable ($\theta_e$ decreasing with height) and, as a result, convective processes are likely to dominate inertial forcing. A second negative EPV region around 800 hPa and between points C and D, where the atmosphere was saturated but convectively stable, was therefore susceptible for the release of CSI. Thus, this cross section provides an objective means of quickly identifying those regions in which CSI is likely. It should be used in conjunction with conventional $M_v$ and $\theta_e$ cross sections. A command function to compute EPV was written for PCGRIDDS and is contained in the appendix.

As it turned out, multiple banded structures in the precipitation field (Fig. 3) detected by the WSR-88D...
radar at Sterling, Virginia, were observed close to this region. The precipitation bands observed on this day were nearly parallel to the mean thermal wind (950–400-hPa thickness contours). A band would form and persist for several hours before dissipating. The band typically was observed to exhibit virtually no north–south movement, but appeared to stretch along an axis parallel to the 950–400-hPa thermal wind. Once a band dissipated, a new one would form approximately 40–50 km south of the previous band. This formation and dissipation of precipitation bands occurred roughly between 1000 and 1800 UTC 26 February.

6. 30 January 1995

A case from 30 January 1995 is also presented to further reinforce and emphasize the role of CSI in banded precipitation events. This again was an event involving concentrated bands of heavy snowfall over a relatively small area of southeast Virginia. The details of the synoptic setting will not be presented here but CSI is once again believed to have been a primary precipitation mechanism. Snowfall totals for 30 January are shown in Fig. 13a, in which the narrow banded nature of the precipitation is quite evident. In Figs. 13b and 13c the area of potential CSI and negative EPV is more clearly revealed than in the previous 26 February case.

The overall synoptic setting was weak and rather innocuous and did not appear to be conducive to heavy snowfall. However, vertical cross sections of $M_g$, $\theta_v$, and EPV via PCGRIDDS revealed a potential for CSI in the area of snowfall banding. Figure 13b shows a conventional vertical cross section of $M_g$ and $\theta_v$ between points A and B depicted on the snowfall totals map (Fig. 13a). This figure represents a meso-Eta-derived 18-h forecast, initialized at 0000 UTC 30 January and valid at 1800 UTC, nearest the time of maximum snowfall. It shows a strong potential for CSI between points C and D between about 900 and 650 hPa. The area of greatest CSI potential in Fig. 13b is shaded. However, the 300-K $\theta_v$ contour is nearly vertical between 800 and 900 hPa, suggesting that the area may also be convectively unstable. This demonstrates that using an EPV
Fig. 10. Vertical cross section of ageostrophic transverse vertical circulations constructed from point A to point B as in Fig. 11a. Cross section depicts a 12-h forecast from the meso-Eta Model valid at 1200 UTC 26 February 1993. Arrows are proportional to the speed. Horizontal wind component in m s$^{-1}$; vertical component in $\mu$bar s$^{-1}$.

The vertical cross section of $M_g$ and $\theta_e$ in Fig. 12b was available and recognized by Sterling and Wakefield NWS offices during the snowfall event, and became the basis for extending and later canceling advisories for the area. A portion of the evening SFD from one office follows: "... a cross section of momentum and theta-e surfaces via PCGRIDDS revealed a strong potential for CSI between about Norfolk and Quantico [southeast Virginia]. The forecast derived from the NGM grids showed the potential maximized at 18Z [1800 UTC] ... weakening by 00Z [0000 UTC] ... and no longer evident by 06Z [0600 UTC]. From that one might conclude that the enhanced bands would weaken by 00Z and beyond ... which they have." Concurrently, WSR-88D radar showed a steadily weakening and shrinking of the snow bands and the advisory for the area was subsequently canceled.

This event demonstrates the utility of model-derived CSI forecasts coupled with real-time radar, satellite, and vertical profile sounding data in developing effective short-term forecasts and updates.

analysis alone may lead to false assumptions on exactly which process is at work, and that continued use of the conventional $M_g$ and $\theta_e$ cross sections is also important. The lower-right portion of Fig. 13b also shows an area in which convective instability is more apparent by $\theta_e$ contours that fold back over themselves ($\theta_e$ decreasing with height). Figure 13c is a vertical cross section of EPV along the same points as Fig. 13b. This cross section is somewhat easier to interpret in that it quickly reveals potential CSI wherever there are negative values of EPV. Once again a region of potential CSI is indicated between points C and D at lower and middle levels ($\sim 4$ PVU); the negative values of EPV at the lower-right portion of Fig. 13c reveals a region of convective instability. However, had one examined the EPV cross section alone, it would have been impossible to differentiate between those areas of inertial and convective instability or whether both processes may have been at work. For this reason, it is again important to emphasize that the EPV technique must be used in conjunction with conventional $M_g$ and $\theta_e$ cross sections.
Fig. 11. Meso-Eta Model 12-h forecast valid at 1200 UTC 26 February 1993 of (a) 950–400-hPa thickness (solid, dm). Solid bar denotes location of cross section, and (b) vertical cross section of \( M_f \) (heavy solid, \( \text{m s}^{-1} \)) and \( \theta \) (dashed lines, K) constructed from point A to point B shown in (a). Cross section depicts 12-h forecast from the Mesoeta Model valid at 1200 UTC 26 February 1993. Shading indicates regions of weak symmetric stability.
7. Discussion: Methodologies and procedures for routine application of CSI in operations

The importance and implications of CSI to operational forecasters is made clear by Colman (1992). In areas where CSI is operating, large amounts of precipitation, often exceeding expectations, can occur while at the same time exhibiting very tight gradients. How then does the forecaster identify such situations and apply information about CSI in forecasts? While it is realistic today to expect forecasters to anticipate well in advance the occurrence of CSI-banded precipitation structures, forecasting the precise orientation and location of CSI-related precipitation bands is still in the future. Therefore, by using the technique of CSI/EPV discussed here, forecasters can better anticipate CSI occurrence and, once the bands form, more readily monitor the bands' development and evolution, and quickly respond with highly detailed and specialized short-range forecasts and statements. Lussky (1989) and Snook (1992) have suggested a strategy centered first on identifying the potential for CSI, and then in monitoring its evolution through utilization of new advanced observing technologies. Grumm and Forbes (1994) have suggested forecasters first concentrate on identifying synoptic patterns conducive to CSI banding, anticipate that possibility while noting the most probable location and orientation of the bands, then monitor radar trends and issue short-term forecasts to address the possibility of locally heavy precipitation. They state that it is especially important to monitor the changes in orientation of thickness contours, since this will affect the location, orientation, and intensity of the bands with time.

That was, in essence, what forecasters attempted to do on the morning of 26 February 1993. The potential for CSI banding was alluded to in the early morning SFD issued by the NWSFO in Sterling at 1225 UTC 26 February 1993 that stated [the KLWX-Sterling Doppler [88D radar is] showing good banding structure for [a] prolonged moderate snow across northern tier counties of Maryland. With 3 inches [of snow] already reported from Hagerstown into Frederick [MD] ... we have issued a winter storm warning for Maryland zone 7 [central Maryland] for today ... [with] accumulations raised to 4 to 6 inches. High resolution [meso-Eta] model data confirming swath of higher snow amounts across Baltimore into eastern shore [of Maryland]. Hence ... have raised [forecast] snow totals [there from an inch or two] to 2 to 4 inch range. . . .

Upon recognizing the enhanced bands on the WSR-88D radar as anticipated from the forecast model fields showing the potential for CSI enhancement, snow advisories already in effect for a large portion of the area were quickly upgraded to a winter storm warning for a narrow corridor across the northern tier counties of Maryland.

Nevertheless, forecasters that morning were for the most part reacting to events rather than anticipating them in advance. At the time of the event, the meso-Eta gridded model output were not routinely available to NWS field forecasters and where it was available, it was considered an experimental model with little history to go on. In this case, the meso-Eta Model grids (in PCGRIDDS format) were available for downloading to NWS forecasters at Sterling around 1130 UTC; by this time heavy snow bands were already evident on the Sterling (KLWX) radar. Had forecast fields of $M_g$, $\theta_e$, EPV, frontogenetic forcing, and vertical wind profiles been available, it is realistic to assume that those same forecasters may have
better anticipated the events and provided longer lead times of warnings. As Snook (1992) has stated, armed with those tools forecasts of short-term changes in band orientation and intensity might then be possible.

For the two events (26 February 1993 and 30 January 1995), the proper assessment of CSI based upon high-resolution gridded model (i.e., meso-Eta) forecast data and the proper CSI/EPV techniques allowed operational forecasters to better anticipate possible precipitation enhancement from CSI-induced snow bands. Once the bands formed, forecasters reacted to the enhancement that accompanies the CSI-related banding. These case studies illustrate how the wide array of new technology and information from gridded model data, mesoscale models, advanced workstations, satellite imagery, Doppler radar, and wind profilers can be employed effectively when utilized selectively for the particular situation the forecaster is faced with on any given day.

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**APPENDIX**

**Concept of CSI and Computation of EPV**

*a. Basic CSI concept*

The concept of CSI is best illustrated in the simple schematic diagram presented in Fig. 1 as described by Sanders and Bosart (1985). This schematic represents a vertical cross section normal to a geostrophically balanced base-state flow (the flow is into the page),
with the speed of the flow increasing with height $z$. Absolute geostrophic momentum surfaces of the base-
(or environmental-) state flow are represented in Fig. 1 by $\bar{M}_g$ (solid lines) and equivalent potential temper-
ature by $\bar{\theta}_e$ (dashed lines). In illustrating the unstable
nature of slantwise displacement, it is important to un-
derstand that parcels are assumed to conserve their
initial $M_g$ and $\theta_e$ values for any hypothetical displace-
ment discussed here. Sanders and Bosart (1985) have
shown that a parcel is stable if displaced in any purely

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**Fig. 13. (Continued)**
horizontal or vertical manner (points A and B, respectively) but unstable to a slantwise displacement (point C). Additionally, it is important to note that in Fig. 1 both $\overline{M}$, and $\overline{\theta} \epsilon$ increase upward (positive $z$ direction) and to the right (positive $x$ direction). From Sanders and Bosart (1985), parcel accelerations in the $x$ direction:

$$\frac{du}{dt} = f(M - \overline{M}) = fM', \quad (A1)$$

where $f$ is the Coriolis parameter (assumed constant) and $M'$ is the parcel’s angular momentum excess (or deficit) with respect to its environment; parcel accelerations in the $z$ direction,

$$\frac{dw}{dt} = g[(\theta_e - \overline{\theta_e})/\overline{\theta_e}] = g\theta_e'/\overline{\theta_e}, \quad (A2)$$

where $g$ is the acceleration of gravity, $\theta_e'$ is the parcel’s potential temperature excess (or deficit) with respect to its environment, and $\overline{\theta_e}$ is a constant reference equivalent potential temperature (Emanuel 1983b).

From this it can be seen that in the case of a leftward horizontal displacement from point A,

$$M > \overline{M}, \quad \text{therefore } M' > 0 \text{ and } \frac{du}{dt} > 0;$$

similarly,

$$\theta_e > \overline{\theta_e}, \quad \text{therefore } \theta_e' > 0 \text{ and } \frac{dw}{dt} > 0.$$

Therefore, the leftward displacement from point A results in an excess of parcel momentum over background geostrophic momentum with the response ($\frac{du}{dt} > 0$) acting to restore the parcel to its initial position toward the right. Additionally, although the leftward displacement results in a positive buoyancy ($\frac{dw}{dt} > 0$), it is insufficient to counteract the horizontal acceleration and is not maintained since the parcel is within a stable stratification.

For the case of an upward displacement from point B,

$$M < \overline{M}, \quad \text{therefore } M' < 0 \text{ and } \frac{du}{dt} < 0;$$

similarly,

$$\theta_e < \overline{\theta_e}, \quad \text{therefore } \theta_e' < 0 \text{ and } \frac{dw}{dt} < 0.$$

Therefore, the upward displacement from point B results in a negative buoyancy ($\frac{dw}{dt} < 0$). Moreover, the horizontal acceleration ($\frac{du}{dt} < 0$) is quickly damped in an inertially stable environment.

Now, for the case of a slantwise displacement from point C,

$$M < \overline{M}, \quad \text{therefore } M' < 0 \text{ and } \frac{du}{dt} < 0,$$

and,

$$\theta_e > \overline{\theta_e}, \quad \text{therefore } \theta_e' < 0 \text{ and } \frac{dw}{dt} > 0.$$

Therefore, the slantwise displacement from point C results in an acceleration in the same direction as the displacement. This schematic also shows that for symmetric instability to exist, the slope of the $\theta_e$ surfaces must be steeper (more vertical) than the $M_e$ surfaces. The preceding has been condensed from Sanders and Bosart (1985), to which the reader is referred for a more complete derivation and explanation. Holton (1992) and Emanuel (1983b) are also appropriate references with regard to this discussion.

Unlike frontogenetical forcing, which is usually evident in single banding of precipitation, CSI is manifest in multiple bands oriented along or parallel to the thermal wind (thickness contours). However, as stated earlier, CSI may enhance circulations associated with frontogenesis while decreasing the scale of vertical motion, and in this way result in narrow strong bands of precipitation, with a broader single band, often convective in nature.

### b. Commands to compute EPV using PCGRIDDDS

Vertical cross sections of angular geostrophic momentum ($M_e$), along with equivalent potential temperature ($\theta_e$), can be computed and displayed using versions of PCGRIDDDS dated after March 1993. The PCGRIDDDS commands used to compute $M_e$ surfaces is defined as

```
SSUM NORM GEOS SMLT FFFF DIST.
```

PCGRIDDDS commands can be saved in a simple ASCII text file for later execution as a “command” file. The filename used for a PCGRIDDDS command file must be limited to 1–4 characters with an extension of “.CMD” (e.g., MSFC.CMD).

Computation of EPV in a vertical cross section was not possible with PCGRIDDDS versions dated before January 1995. An updated PCGRIDDDS version released in late 1995 allows users to effectively extend the PCGRIDDDS command set through use of “aliasing.” The updated PCGRIDDDS version allows a user to create and maintain an ASCII text file called “ALIAS.DAT” that contains the definitions of new commands using strings of previously defined commands. In essence, the command set of PCGRIDDDS is extendable and allows an almost unlimited number of commands to be defined by users. Below are the commands that are entered on separate lines (as shown below) into the ALIAS.DAT file to calculate EPV [see (2)] (boldface denotes the newly defined command names):

```
MSFC CLR9 SSUM NORM GEOS SMLT DIST FFFF
TRMA SDVD SDIF MSFC LV+1 MSFC LV-1 SDIF PRES LV+1 PRES LV-1
TRMB TANG GRAD THTE
TRMC TANG GRAD MSFC
TRMD SDVD SDIF THTE LV+1 THTE LV-1 SDIF PRES LV+1 PRES LV-1
TRM1 SMUL TRMA TRMB
TRM2 SMUL TRMC TRMD.
EPV1 DNEG SMLC 9.8 SDIF SMLT TRMA TRMB SMLT TRMC TRMD.
```

The quantity $EPV_1$ is entered while in cross-section mode. Areas of negative $EPV$ will be depicted with
REFERENCES


