

# BASELINE CLIMATOLOGY OF SOUNDING DERIVED PARAMETERS ASSOCIATED WITH DEEP MOIST CONVECTION

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## Abstract

*A baseline climatology of several parameters commonly used to forecast deep, moist convection is developed using an extensive sample of upper-air observations. Previous climatologies often contain a limited number of cases or do not include null cases, which limit their forecast utility. Three years of evening (0000 UTC) rawinsonde data (approximately 60,000 soundings) from the lower 48 United States are evaluated. Cloud-to-ground lightning data and severe weather reports from Storm Data are used to categorize soundings as representative of conditions for no thunder, general thunder, severe, significant hail/wind, or significant tornado. Among the detailed calculations are comparisons between both convective available potential energy (CAPE) and lifted condensation level (LCL) using a most unstable parcel versus a mean lifted 100-hPa parcel. Lapse rates for several different layers are inspected to determine the utility of using static stability versus CAPE to forecast storm severity. Lastly, low-level shear is studied in an attempt to distinguish between severe and significant tornado episodes.*

*One of the major findings is a considerable difference between 0-1 km above ground level (AGL) magnitude of vector difference of wind for significant tornado episodes versus the other five categories. Statistically significant differences are also noted between LCL/mean lifted LCL (MLLCL) heights AGL for significant tornado events and the other convective categories. In addition, much less seasonal variation is found for 0-1 km shear, 0-6 km shear, and MLLCL heights AGL for significant tornado events compared with the remainder of the data set.*

## 1. Introduction

Meteorologists at the NOAA/NWS Storm Prediction Center (SPC) in Norman, Oklahoma, routinely prepare forecasts of severe thunderstorm potential for the lower 48 states. Since 1999, SPC has been issuing probabilistic forecasts of tornadoes, damaging winds, and large hail. In addition, probabilistic forecasts of significant severe weather (i.e., tornadoes with intensities of F2 or greater, wind gusts  $\geq 120 \text{ km h}^{-1}$  [65 kt] wind gusts, or hail  $\geq 5 \text{ cm}$  [2 in.] diameter) are composed. Over the past several years, the availability of gridded model output has made

access to explicit forecast parameters such as vertical wind shear and lapse rates possible. This has allowed forecasters to finally use techniques developed over half a century ago in a real-time operational setting (e.g., Showalter and Fulks 1943). Surface to 6 km above ground level (AGL) magnitude of vector difference of wind (hereafter 0-6 km shear) and 700-500 hPa lapse rates are used frequently in assessing severe potential, particularly for the prediction of supercells.

The purpose of this study is to use rawinsonde data to examine several parameters commonly used to forecast severe thunderstorms and tornadoes. The research complements work by Rasmussen (2003) and Rasmussen and Blanchard (1998), but includes a much larger dataset (an order of magnitude larger), null cases, and does not attempt to determine convective mode.

### a. Lapse rate

Recent work relating instability to tornado occurrence has focused on convective available potential energy (CAPE; Moncrieff and Miller 1976), with less research devoted to the effects of lapse rates on severe storm/tornado formation. However, past research has been compiled that studied the effects of elevated mixed layers (steep middle level [e.g., 700-500 hPa layer] lapse rates) and the associated capping inversion (or lid) on deep, moist convection and severe thunderstorm formation. Although there is no standard definition of a "steep" lapse rate, we will arbitrarily classify any lapse rate exceeding  $7^\circ\text{C km}^{-1}$  as steep. This is derived from values found useful in operational severe thunderstorm forecasting at the SPC. Carlson et al. (1983) discussed a conceptual model of how the capping inversion associated with the elevated mixed layer focuses the location and even enhances the intensity of severe local storms. The capping inversion prevents convection from developing in areas of high CAPE, allowing the boundary layer to moisten further and permit the build up of additional potential instability.

Doswell et al. (1985) discussed the importance of steep 700-500 hPa lapse rates for both the creation of strong conditional instability and for enhancing the atmospheric response to quasi-geostrophic forcing. The superposition of steep lapse rates and low-level moisture was shown to be ideal for severe storm/tornado formation.

Lanicci (1985) and Lanicci and Warner (1991a, b, c) studied the elevated mixed layer over the southern and central Great Plains and the importance of the capping inversion for severe thunderstorm climatology. Since the capping inversion is normally located between 850 and 700 hPa, steep 700-500 hPa lapse rates are typically associated with an elevated mixed layer. Therefore, this parameter is useful in tracking elevated mixed layer air and capping inversions that have originated over the higher terrain of the western U.S. or northern Mexico. The importance of mountainous terrain on the creation of steep 700-500 hPa lapse rates was shown by Cortinas and Doswell (1998). A minimum in static stability in the 700-500 hPa layer (and thus a maximum in 700-500 hPa lapse rate) was found over the Rocky Mountains during much of the year. Although the elevated mixed layer is more common over the central and southern Great Plains, Farrell and Carlson (1989) found that it played an important role during the major tornado outbreak on 31 May 1985 in Ohio and Pennsylvania.

Steep lapse rates were found to be associated with most major tornado outbreaks by Craven (2000). The 700-500 hPa lapse rate was greater than or equal to  $7^{\circ}\text{C km}^{-1}$  during 80% of tornado outbreaks from 1950 to 1998 that obtained a Destruction Potential Index (DPI) of 100 or more (Thompson and Vescio 1998). For reference, a typical moist adiabatic lapse rate is  $\sim 5.5^{\circ}\text{C km}^{-1}$ , the standard atmosphere lapse rate is  $\sim 6.5^{\circ}\text{C km}^{-1}$  from 0-6 km AGL, while a dry adiabatic lapse rate is  $9.8^{\circ}\text{C km}^{-1}$ . The DPI is calculated using the product of the tornado path area (path length multiplied by maximum path width) and the F-scale (F-scale + 1 so that F0 tornadoes can be assigned a non-zero number).

#### *b. Lifted condensation level*

Recent research indicates a relationship between tornadic supercells and relatively high boundary layer relative humidity, which can be represented by low lifted condensation levels (LCL). Rasmussen and Blanchard (1998) found that the parameter that showed the most utility for discriminating between significant tornadoes and supercells with either weak or no tornadoes was the height of the LCL. The median LCL height was  $\sim 500$  m lower for the strong or violent tornado cases. Nearly identical results were found by Edwards and Thompson (2000), with a mean difference in LCL height for significant tornadic versus weak or non-tornadic supercells of  $\sim 500$  m. During severe weather episodes in the north-central United States, Johns et al. (2000) compared the median LCL height near the location of the first intense tornado versus the median LCL height 100 statute miles into the warm sector. This work highlighted that the median LCL height in the warm sector was nearly 800 m higher than in the area where the tornadoes occurred.

#### *c. Vertical wind shear*

Much work has been completed that relates deep layer shear to the potential for supercell formation (see Table 1 for a summary and definitions of each shear parameter). Weisman and Klemp (1984, 1986) and Weisman (1996)

performed extensive storm-scale modeling that indicates that "shear"<sup>1</sup> values of  $20 \text{ m s}^{-1}$  over the lowest 4-6 km AGL is sufficient to promote supercell storm formation. Davies and Johns (1993) calculated the bulk Richardson number shear for 260 strong and violent tornadoes and found that the median was  $\sim 22 \text{ m}^2 \text{ s}^{-2}$ . Using a year's worth of soundings from 1992, Rasmussen and Blanchard (1998) created a climatology of supercell/tornado parameters and found that the median of boundary layer to 6 km shear for supercells was  $19 \text{ m s}^{-1}$ . Their results also indicated that there is little difference in this deep-layer shear parameter between supercells containing significant (strong or violent) tornadoes ( $18 \text{ m s}^{-1}$ ) and those that do not. A well-defined lower threshold of  $20 \text{ m s}^{-1}$  in 0-6 km shear was found for a dataset of 260 right-moving supercells by Bunkers et al. (2000). In a study of 65 major tornado outbreaks from 1950-1998, Craven (2000) found that virtually all of the events (97%) were associated with surface to 6 km shear values  $\geq 20 \text{ m s}^{-1}$ .

These findings have identified a rather simple way to determine the kinematic potential for supercells versus non-supercell thunderstorms from a deep-layer magnitude of vector difference of wind. However, the question of tornado potential appears to be more closely related to low-level shear or storm-relative helicity. Davies-Jones et al. (1990) studied 28 tornadoes of various strengths, and found that 0-3 km storm-relative helicity (SRH) generally increases as the intensity of the tornadoes increases. The results of Rasmussen and Blanchard (1998) indicated a statistically significant difference between 0-3 km SRH for ordinary thunderstorms versus supercells that produced significant tornadoes. The SRH values were calculated using observed storm motion in each of these studies. However, considerable overlap is noted for supercells that did not produce strong and violent tornadoes, indicating a potential false alarm problem.

Marwitz (1972a) found that the mean subcloud environmental winds for supercells producing hailstorms are strong (greater than  $10 \text{ m s}^{-1}$ ) and veer by more than  $60^{\circ}$  from the mean environment winds (and also veer greater than  $50^{\circ}$  within the subcloud layer), suggesting strong low-level shear. Complementary research of non-supercell storms (Marwitz 1972b) concluded that the distinguishing characteristic of the environment that produces non-supercell storms (versus supercells) is light winds (and thus weaker low-level shear) in the subcloud layer. In a study of 21 cases of severe thunderstorms including supercells, Doswell and Lemon (1979) found that the most reliable kinematic parameter is the low-level shear (surface to  $\sim 1500$  m [5000 ft] AGL), which seems to be well-related to the low-level mean wind speed and to the region of severe convection. Johns et al. (1990) studied 0-2 km positive shear with a dataset of 242 strong and violent tornadoes. Results indicate that the majority of the tornadoes were associated with low-level (0-2 km) positive shear values in excess of  $10 \times 10^3 \text{ s}^{-1}$ . A study of severe

<sup>1</sup> Units of shear are  $\text{s}^{-1}$ ; many of the parameters presented as "shear" or "vertical wind shear" in previous research are not listed with shear units (typically are magnitude of vector difference between surface wind and wind at top of layer).

**Table 1.** Summary of measures of wind shear from previous research

Author	Depth	Name/Definition	Units
Marwitz (1972a,b)	Surface to 4 km MSL	Subcloud layer shear	s <sup>-1</sup>
Doswell and Lemon (1979)	0-1524 m AGL	Layer average vector shear	s <sup>-1</sup>
Weisman and Klemp (1984)	0-5 km AGL shear	Magnitude of vector difference	m s <sup>-1</sup>
Weisman and Klemp (1986)	0-6 km AGL shear	Magnitude of vector difference	m s <sup>-1</sup>
Davies and Johns (1993) Johns and Hart (1993)	0-2 km AGL positive mean shear	Hodograph length divided by depth of layer, setting the shear magnitude to zero for those hodograph segments where the ground relative winds back "significantly" with height	s <sup>-1</sup>
Davies and Johns (1993)	0-6 km AGL BRN shear	Magnitude of vector difference between 0-500 m AGL mean wind and the 0-6 km AGL mean wind	(m s <sup>-1</sup> ) <sup>2</sup>
Rasmussen and Blanchard 1998)	Boundary layer to 6 km AGL shear	Magnitude of vector difference between 0-500 m AGL mean wind and 6 km AGL wind	m s <sup>-1</sup>
Craven (2000) Bunkers et al. (2000)	0-6 km AGL shear	Magnitude of vector difference	m s <sup>-1</sup>

Note: "magnitude of vector difference" refers to the difference between the surface wind and the wind at the top of the layer in question.

weather outbreaks involving bow echoes versus those with supercells was completed by Johns and Hart (1993). In their small sample of cases, they found that tornado outbreaks were associated with 0-3 km SRH in excess of 400 m<sup>2</sup> s<sup>-2</sup>, while the 0-3 km SRH during the bow echo cases was less than 120 m<sup>2</sup> s<sup>-2</sup>. The 0-2 km positive shear in the supercell outbreaks was also nearly double that found in the bow echo events. Finally, Edwards and Thompson (2000) used 51 proximity soundings generated by the RUC-2 model to study several forecast supercell parameters. They found a statistically significant difference between the mean 0-1 km SRH for supercells with significant tornadoes (~150 m<sup>2</sup> s<sup>-2</sup>) versus supercells with either weak or no tornadoes observed (~100 m<sup>2</sup> s<sup>-2</sup>).

## 2. Data

### a. Proximity criteria

0000 UTC rawinsonde soundings from 1997-1999 for the lower 48 states were collected. A total of 60,090 soundings are included. Proximity is defined as being within 185 km (100 nm) of the sounding release location, and during the period from 2100 UTC to 0300 UTC (6-hour period centered on the 0000 UTC sounding). The 185-km threshold lies within the range of the 80-km (Darkow 1969; Schaefer and Livingston 1988; Brooks et al. 1994) and 400-km criteria utilized by Rasmussen and Blanchard (1998). For a detailed discussion on the difficulty of defining and selecting a proximity sounding, see Brooks et al. (1994).

### b. Events

Lightning data from Global Atmospheric, Inc., (Orville 1991) and convective severe weather reports (NCDC 1997, 1998, 1999; Hart and Janish 1999) were utilized to subdivide the dataset into six categories (Table 2). Of the more than 60,000 possible events, 32,141 (53%) had non-zero CAPE based on the most unstable parcel in the lowest 300 hPa (MUCAPE). Of the 45,508 no-thunder events, 27,949 (61%) had no MUCAPE and 17,559 (39%) had non-zero MUCAPE. The categories are exclusive, and each event was assigned using the most severe report (i.e., a F2 tornado event was assigned only to significant tornadoes, even if 1-in. hail also occurred).

The lightning strike threshold of two or more cloud-to-ground (CG) strikes is consistent with the criteria established by Reap (1986) and R. Orville (2001, personal communication), similar to the  $\geq 3$ -CG strike threshold used by Hamill and Church (2000), but much less than the  $\geq 10$ -CG strike criteria used by Rasmussen and Blanchard (1998).

### c. Quality control

No attempt was made to modify the soundings. It was anticipated that the effects of unrepresentative, contaminated, or erroneous data would be damped out in the statistical analysis. A simple objective quality control procedure for the severe, significant hail/wind, and significant tornado soundings removed all soundings with MUCAPE less than 150 J kg<sup>-1</sup> (Brooks et al. 1994). General thunder soundings were removed if no MUCAPE was present. All

**Table 2.** Definitions and number of proximity soundings for the six convective categories

Quantity	Category	Definition
27949	no thunder (no CAPE)	0-1 CG strikes (and zero MUCAPE)
17559	no thunder (CAPE)	0-1 CG strikes (and non-zero MUCAPE)
11339	general thunder	≥ 2 CG strikes
2644	severe	0.75-1.99" hail and/or 50-64 kt gust and/or wind damage and/or F0 or F1 tornado
512	significant hail/wind	≥ 2.00" hail and/or ≥ 65 kt gust
87	significant tornado	F2-F5 tornado

**Table 3.** Parameters computed from soundings

Parameter	Units
MUCAPE (most unstable parcel CAPE in lowest 300 hPa)	J kg <sup>-1</sup>
MUCIN (most unstable parcel Convective Inhibition [CIN] in lowest 300 hPa)	J kg <sup>-1</sup>
MLCAPE (100-hPa mean layer CAPE)	J kg <sup>-1</sup>
MLCIN (100-hPa mean layer CIN)	J kg <sup>-1</sup>
0-3 km AGL Lapse Rate	°C km <sup>-1</sup>
3-6 km AGL Lapse Rate	°C km <sup>-1</sup>
700-500 hPa Lapse Rate	°C km <sup>-1</sup>
850-700 hPa Lapse Rate	°C km <sup>-1</sup>
DCAPE (Downdraft CAPE)	J kg <sup>-1</sup>
LCL height (lifted condensation level)	m AGL
MLLCL (100-hPa mean layer LCL height)	m AGL
0-1 km shear (magnitude of vector difference)	m s <sup>-1</sup>
0-6 km shear (magnitude of vector difference)	m s <sup>-1</sup>

CAPE values were calculated using the virtual temperature correction (Doswell and Rasmussen 1994).

Subjective quality control was minimal because of the size of the dataset. Lapse rates in excess of 11°C km<sup>-1</sup> in the 0-3 km AGL layer were removed, as were those above 10.2°C km<sup>-1</sup> in the 0-6 km AGL layer, 850-700 hPa layer, and 700-500 hPa layer. 0-1 km (0-6 km) shear values greater than 50 m s<sup>-1</sup> (100 m s<sup>-1</sup>) were also excluded. In addition, all soundings with MUCAPE and/or 100-hPa mean layer CAPE (MLCAPE) greater than 5000 J kg<sup>-1</sup> were manually inspected, and suspect soundings were excluded.

#### d. Parameters

A list of the parameters computed from the sounding dataset is shown in Table 3. These parameters cover

three main groups: a) instability/lapse rates, b) LCL heights, and c) vertical wind shear.

### 3. Results

Box-and-whisker plots (Tukey 1977) are used extensively to compare data in each category. On a single graphic, these plots show information about range, variance, and median values. The plot shows the 10th (bottom whisker), 25th (bottom of box), 50th (horizontal line within box), 75th (top of box), and 90th percentiles (top whisker) of the particular data. The 25th percentile, or bottom of the box, indicates that 75 percent of the data is larger than the particular value. For example, Fig. 1 reveals that 75% (bottom of box, or 25th percentile) of all significant tornado events have a MLCAPE value of slightly more than 500 J kg<sup>-1</sup>. Comparing box and whisker plots in different categories yields information about the similarity of the data. For example, in Fig. 6, the 75th percentile of significant hail/wind events is less than the 25th percentile of the significant tornado events (the boxes don't overlap). This lack of overlap suggests a statistically significant difference between the data.

#### a. Instability/lapse rates

Recent research suggests that the most accurate estimate of convective cloud base from 0000 UTC rawinsonde data, and thus the most accurate representation of parcel path, utilizes a mean layer parcel, say from the mean temperature and dewpoint in the lowest 100 hPa (Craven et al. 2002). Thus, MLCAPE was chosen to compare potential instability for the soundings in this database. Although the median value of MLCAPE tends to increase with increasing intensity of deep convection, there was considerable overlap in the distributions (Fig. 1). When instability was present, 75 percent of the no thunder events had MLCAPE values less than 250 J kg<sup>-1</sup>, while more than 50 percent of thunder soundings had more MLCAPE. Likewise, 75 percent of the thunder soundings had less than 1100 J kg<sup>-1</sup>, while more than 50 percent of significant hail/wind and significant tornado events had more MLCAPE.

Low-level lapse rates had a different signal than MLCAPE (Fig. 2). The 0-3 km AGL layer displayed little difference between thunder, severe, and significant hail/wind events, with medians near 7.5°C km<sup>-1</sup>. However, low-level lapse rates were much smaller for significant tornado events, with 75 percent of those events occurring with values less than 7.5°C km<sup>-1</sup>. It is interesting to note that the significant tornado distribution looks much like that of the no thunder (CAPE) distribution. More evidence for the reasons behind this will be presented later. However, it is likely related to a moist boundary layer, which reduces the degree of mixing and results in a shallower boundary layer (if strong moisture flux convergence is not occurring). In addition, capping inversions associated with the elevated mixed layer in the Plains are often associated with tornado events. The presence of the capping inversion in the 850-700 hPa layer would result in smaller 0-3 km AGL lapse rates.

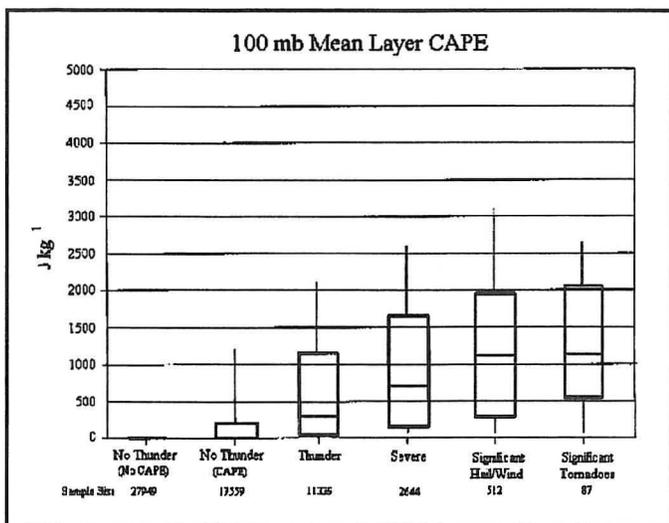


Fig. 1. Box and whisker plot of 100-hPa mean layer CAPE ( $J\ kg^{-1}$ ). 10th, 25th, 50th, 75th, and 90th percentiles are shown.

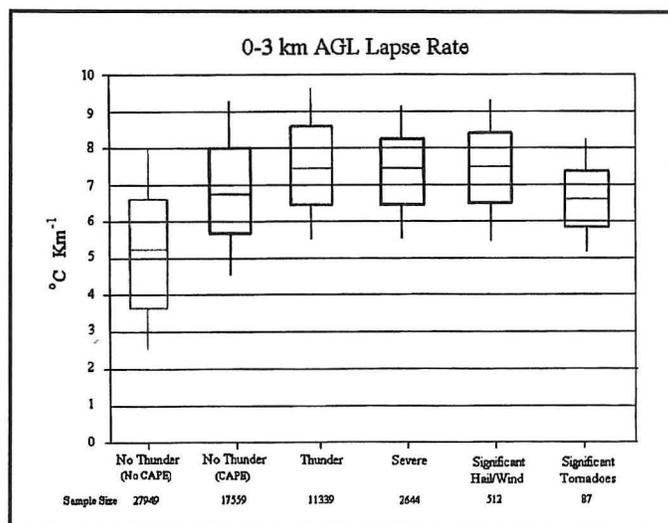


Fig. 2. As in Fig. 1, except for 0-3 km AGL lapse rate ( $^{\circ}C\ km^{-1}$ ).

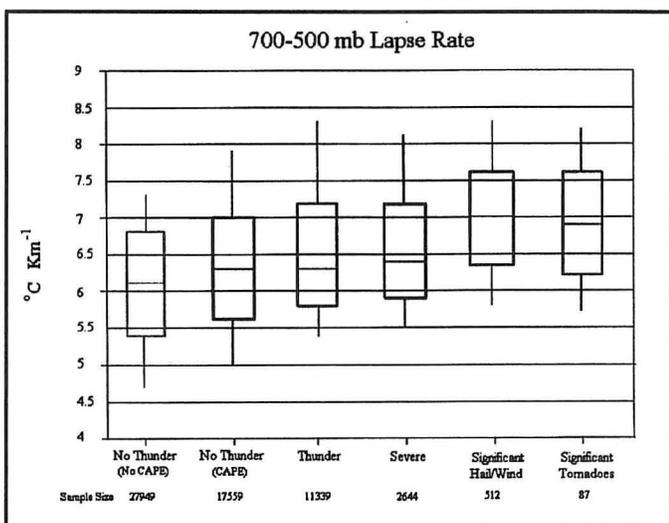


Fig. 3. As in Fig. 1, except for 700-500 hPa lapse rate ( $^{\circ}C\ km^{-1}$ ).

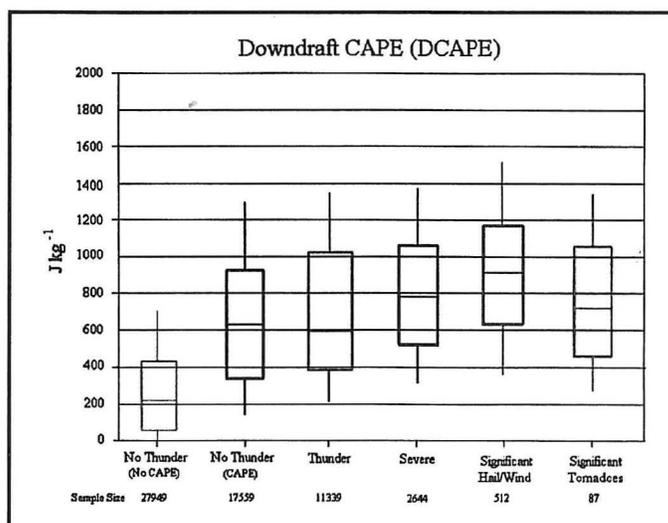


Fig. 4. As in Fig. 1, except for downdraft CAPE (DCAPE;  $J\ kg^{-1}$ ).

Midlevel lapse rates suggest that significant severe weather episodes tend to have steeper values than the rest of the data set (Fig. 3). While most of the categories show similar medians below  $6.5^{\circ}C\ km^{-1}$ , the significant hail/wind and significant tornado events tend to occur when 700-500 hPa lapse rates are above  $6.5^{\circ}C\ km^{-1}$ . However, considerable overlap does exist.

Downdraft CAPE (DCAPE; Gilmore and Wicker 1998) in this study was calculated by taking the minimum wet-bulb temperature in the 700-500 hPa layer pseudo-adiabatically to the surface without entrainment. The area between this line and the ambient temperature is the DCAPE. Thus, DCAPE is maximized by a combination of steep lapse rates below 700 hPa and a very dry layer between 700 and 500 hPa. There was a tendency for DCAPE values to increase during the progression from thunder to significant hail/wind, with median values increasing from  $600\ J\ kg^{-1}$  to over  $900\ J\ kg^{-1}$  (Fig. 4). Although there was considerable overlap, the significant tornado events tend to occur with lower values than sig-

nificant hail/wind events. A possible explanation is that higher DCAPE values permit stronger rear flank downdrafts, which could result in an outflow-dominated supercell storm that undercuts the mesocyclone, thereby inhibiting strong tornadogenesis. Much like low-level lapse rates, the distributions of DCAPE between significant tornado events and no thunder (CAPE) events are quite similar. Since DCAPE values are proportional to low-level lapse rates (the steeper the low-level lapse rate, the higher the DCAPE value), these results are consistent with the results from the 0-3 km AGL lapse rate dataset.

*b. Cloud bases/LCL heights*

The 100-hPa mean layer LCL (MLLCL) height AGL shows little difference between most events, with median values above 1200 m AGL (Fig. 5). However, cloud bases tended to be lower during significant tornado events, with 75% of the cases containing MLLCL heights less

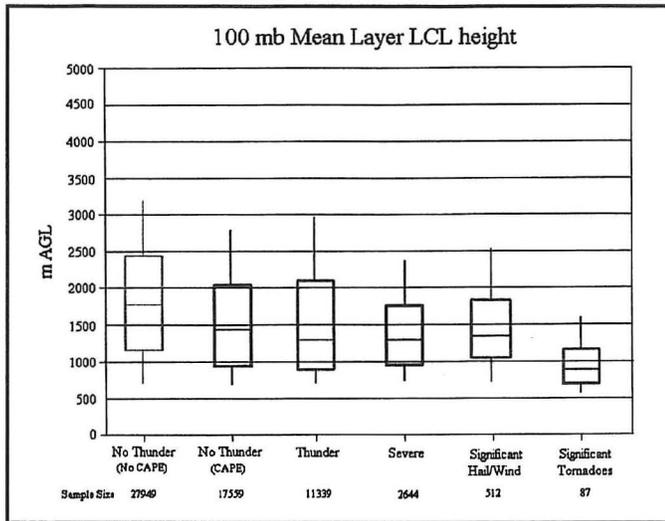


Fig. 5. As in Fig. 1, except for 100-hPa mean layer LCL (MLLCL) height (m AGL).

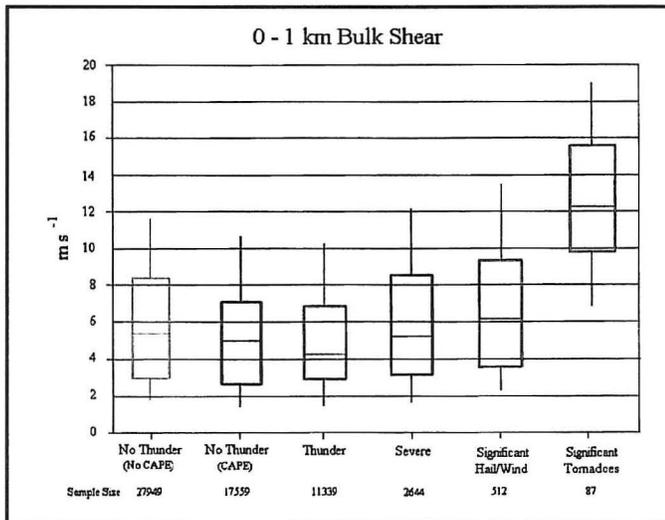


Fig. 6. As in Fig. 1, except for 0-1 km shear ( $\text{m s}^{-1}$ ).

than 1200 m AGL. The median values of the cloud bases in this category were about 500 m less than the rest of the dataset, which is consistent with earlier research (Rasmussen and Blanchard 1998; Edwards and Thompson 2000; Johns et al. 2000; Markowski et al. 2000). The lower cloud bases likely indicate that less sub-cloud evaporation will take place, decreasing the chance that the storm will be dominated by cold outflow that would undercut the mesocyclone. This is probably also related to smaller surface-to-3 km AGL lapse rates (Fig. 2). The drier the boundary layer is, the deeper the mixed layer can become and the stronger the surface-to-3 km AGL lapse rate will likely be (again, if strong moisture flux convergence is not occurring). Thus, a moist boundary layer with associated low MLLCL heights would also be associated with somewhat smaller low-level lapse rates. In addition, capping inversions are often found between 850 and 700 hPa during strong tornado events. This capping inversion, often the result of an elevated mixed layer from upstream higher terrain (e.g., the Rocky

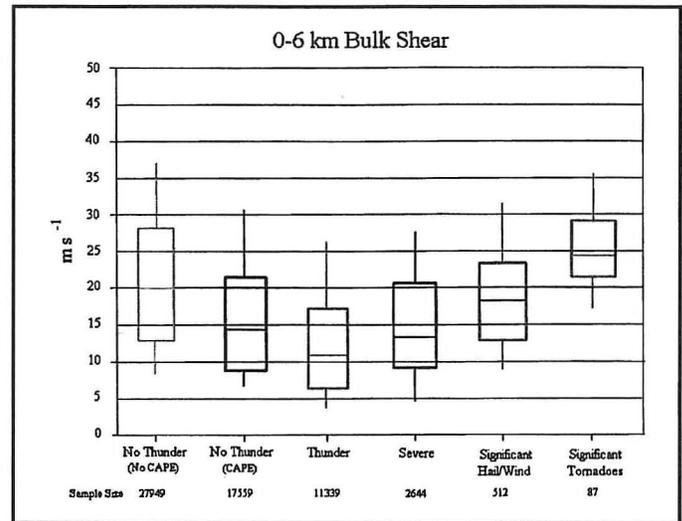


Fig. 7. As in Fig. 1, except for 0-6 km shear ( $\text{m s}^{-1}$ ).

Mountains), would result in smaller surface-to-3 km AGL lapse rates.

### c. Vertical wind shear

The most striking results of this study involved the low-level shear (Fig. 6). There was little difference in 0-1 km shear for the first five categories. However, a very substantial difference is evident between significant tornado events and the rest of the dataset. Nearly 75 percent of significant tornado events occurred with values in excess of  $10 \text{ m s}^{-1}$ . In contrast, more than 75 percent of the significant hail/wind events had less low-level shear. Thus, much like the lower threshold that has been established for deep-layer shear and supercell development (i.e.,  $20 \text{ m s}^{-1}$ ; Weisman and Klemp 1982, Davies and Johns 1993; Rasmussen and Blanchard 1998; Bunkers et al. 2000; Craven 2000), it appears that  $10 \text{ m s}^{-1}$  ( $20 \text{ kt}$ ) may be used as a lower threshold for significant tornado events. Stronger low-level shear appears to be associated with a higher frequency of strong and violent tornado events. These results are consistent with Edwards and Thompson (2000), who found a substantial difference between the mean 0-1 km SRH for supercells with significant tornadoes versus supercells with either weak or no tornadoes observed. While SRH requires an estimated or observed storm motion, using a 0-1 km shear vector does not.

The no thunder (no CAPE) soundings represent almost half of the data set, and are dominated by cold season situations where strong horizontal temperature gradients result in large thermal winds (Fig. 7). Since vertical wind shear is proportional to the strength of the thermal wind, the 0-6 km AGL shear values can be quite high. However, this fact is somewhat irrelevant since the lack of instability precludes development of deep convection. In addition, notice that the distributions of no thunder (CAPE) and severe are very similar. Recall that there was a substantial difference between the MLCAPe for these categories (Fig. 1), with the 25th and 75th percentiles barely overlapping around  $250 \text{ J kg}^{-1}$ . Thus, it is possible that updrafts in the no thunder (CAPE) environment may

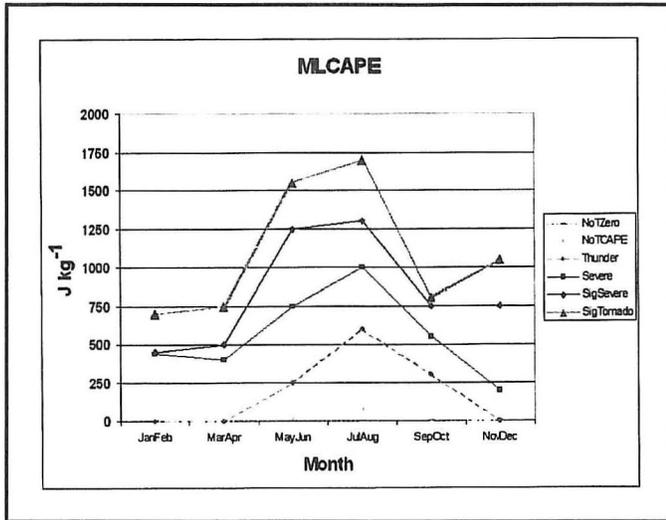


Fig. 8. Seasonal variation in 100-hPa mean layer CAPE (MLCAPE;  $J kg^{-1}$ ) for each category: no thunder (no CAPE; short dashed/rectangle), no thunder (CAPE; long-short dashed/star), thunder (dashed/square), severe (solid/square), significant hail/wind events (solid/diamond), and significant tornadoes (solid/triangle).

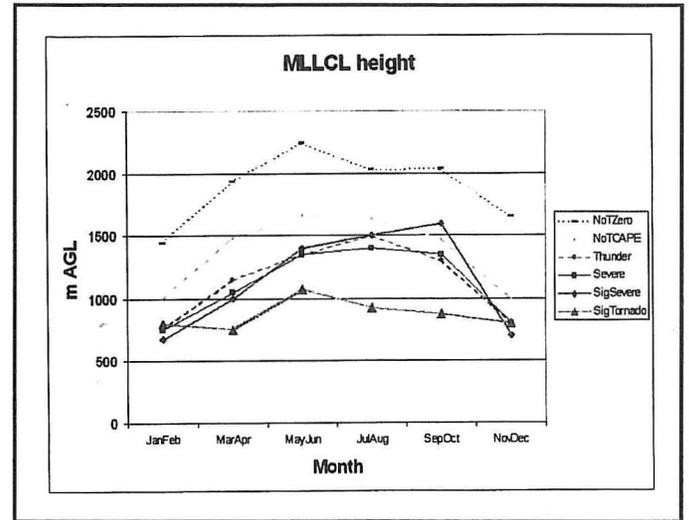


Fig. 9. As in Fig. 8, except for 100-hPa MLLCL height (m AGL).

have trouble sustaining themselves given relatively high vertical wind shear and low potential instability.

Otherwise, deep-layer shear appears to increase with increasing severity of deep convection. The 0-6 km shear increases during the progression from thunder events to significant tornado events. Although there is substantial overlap between severe events and significant hail/wind events, there was no overlap between the upper quartile of severe events and the lower quartile of significant tornado cases. The expected lower threshold for supercells of 18 to 20  $m s^{-1}$  is evident in the significant tornado events. About 75 percent of severe events occur with at least 10  $m s^{-1}$  of deep-layer shear, while almost half of the thunder events had less shear. Thus, there appears to be some value in 10  $m s^{-1}$  as a possible lower threshold for severe versus thunder forecasts, although considerable overlap exists in the data set.

d. Seasonal variations

Due to small sample size, dividing the significant hail/wind and significant tornado groups further into seasonal groups and using box and whisker plots was problematic. Given the degree of overlap within groups using the entire three-year dataset, showing only the median values to indicate seasonal variation is potentially misleading, especially if one would like to determine useful forecast thresholds. Thus, the authors advise that caution be used when comparing categories using the median values alone. The purpose of these figures is to indicate the seasonal variation within each group and show the apparent lack of seasonal variation in a few of the parameters for the significant tornado events.

The data were also partitioned into six 2-month groups to account for seasonal variability. The median values of MLCAPE indicated the expected result of high-

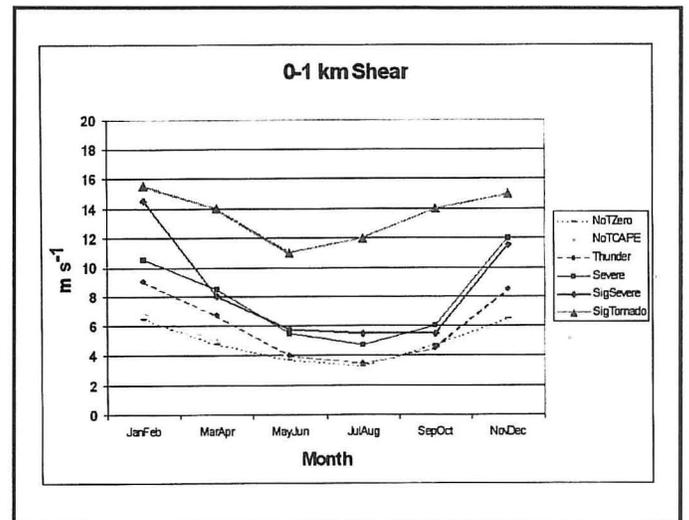


Fig. 10. As in Fig. 8, except for 0-1 km shear ( $m s^{-1}$ ).

er values during the warm season and lower values during the cold season (Fig. 8). The higher the category, the higher the median value of MLCAPE tends to be.

The median cloud base heights indicated the opposite annual trend (Fig. 9). MLLCL heights AGL were higher during the warm season due to deeper mixing. However, minimal seasonal variation was observed in the significant tornado events, where median values tend to remain below 1000 m AGL. The other categories increase 400 to 600 m from the cold season to the warm season. The discrimination between categories is lost during the cold season, when all median values are near 800 m AGL from November to February. However, the difference between significant tornadoes and other events increases to 300 to 500 m during the warm season.

Similar to the MLLCL heights, there was little seasonal variability in either 0-1 km shear or 0-6 km shear in the data set for significant tornado cases (Figs. 10, 11). While the other five categories displayed substantial decrease during the warm season, both low-level and deep-layer shear values remain well above the 10  $m s^{-1}$

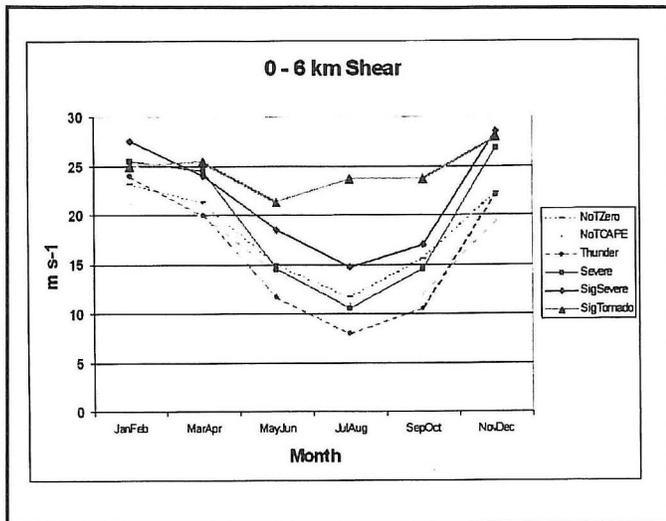


Fig. 11. As in Fig. 8, except for 0-6 km shear ( $\text{m s}^{-1}$ ).

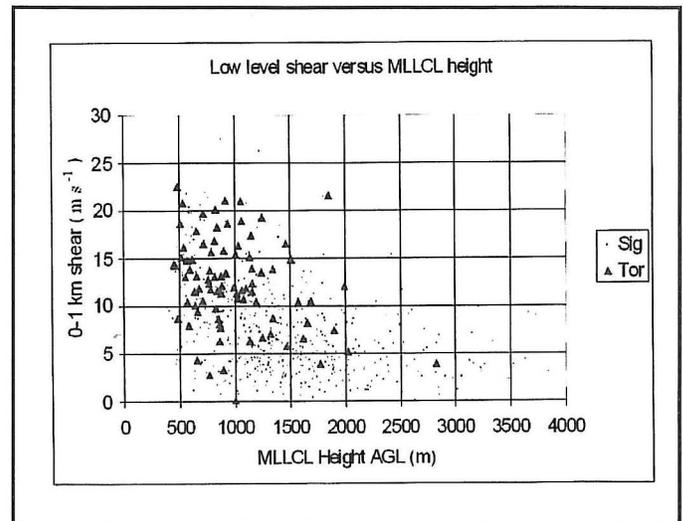


Fig. 12. Scatter plot of 0-1 km shear ( $\text{m s}^{-1}$ ) versus MLLCL height (m AGL) for significant tornadoes (triangles) and significant hail/wind events (dots).

and  $20 \text{ m s}^{-1}$  thresholds (respectively) for significant tornadoes and supercells throughout the year.

#### 4. Parameter Combinations

##### a. 0-1 km shear vs. MLLCL height

Examining low-level shear and MLLCL height yields a strong signal between significant tornadoes and significant hail/wind events (Fig. 12). Strong/violent tornadoes tend to occur with relatively high 0-1 km shear (e.g.,  $> 10 \text{ m s}^{-1}$ ) and relatively low MLLCL height (e.g.,  $< 1200 \text{ m AGL}$ ). Storms that produce hail  $\geq 2 \text{ in.}$  diameter and/or wind gusts  $\geq 65 \text{ kt}$  but no strong/violent tornadoes tend to have weaker low-level shear and higher cloud bases.

##### b. Significant severe parameter

In general, individual parameters did not discriminate well between thunder and severe events. However, when considering both instability and shear (Davies and Johns 1993; Johns et al. 1993) simultaneously, the results showed a noticeable improvement. Calculating the product of MLCAPE and 0-6 km shear (defined as significant severe parameter in  $\text{m}^3 \text{ s}^{-3}$ ) yielded better discrimination between thunder events and the three severe categories, especially between thunder events and the significant hail/wind and tornado events (Fig. 13). Possible lower thresholds of  $10,000 \text{ m}^3 \text{ s}^{-3}$  (severe),  $20,000 \text{ m}^3 \text{ s}^{-3}$  (significant hail/wind), and  $30,000 \text{ m}^3 \text{ s}^{-3}$  (significant tornadoes) may be used given the distributions of this instability/shear parameter (see Appendix).

##### c. Strong Tornado Parameter

Five of the individual parameters studied showed some promise in discriminating between significant tornado events and other categories. The following combination of parameters was examined to see if a parameter

could be assembled that would assist in diagnosing the potential for strong/violent tornadoes (Strong Tornado Parameter [STP,  $\text{m s}^{-2}$ ]):

$$STP = \frac{(MLCAPE) * (0 - 1 \text{ km shear}) * (0 - 6 \text{ km shear})}{MLLCL * DCAPE} \quad (1)$$

This is similar to the Significant Tornado Parameter (Thompson et al. 2003), but uses 0-1 km shear rather than 0-1 km SRH. Thus, an observed or estimated storm motion is not required. In addition, the Strong Tornado Parameter includes DCAPE.

Essentially, a combination of high instability and strong vertical wind shear, along with low cloud bases and low probability of strong/cold downdrafts will increase the probability of significant tornadogenesis. The results show that well more than 50% of the strong/violent tornadoes occurred with STP values  $> 0.25 \text{ m s}^{-2}$  while 75% or more of the events in the other categories occurred at  $< 0.25 \text{ m s}^{-2}$  (Fig. 14; Appendix).

#### 5. Summary

Inspection of a large data base of soundings from the lower 48 continental United States from 1997-1999 yielded the following results:

- MLCAPE discriminates somewhat between no thunder and thunder soundings, but there is considerable overlap between thunder and the three severe categories;
- Out of about a dozen parameter combinations, the Significant Severe Parameter (product of MLCAPE and 0-6 km shear) appeared to show some discrimination between thunder and severe events;
- 0-1 km shear and MLLCL height both discriminate well between significant tornado events and other severe events. A combination of the two parameters shows even more skill in distinguishing between categories;

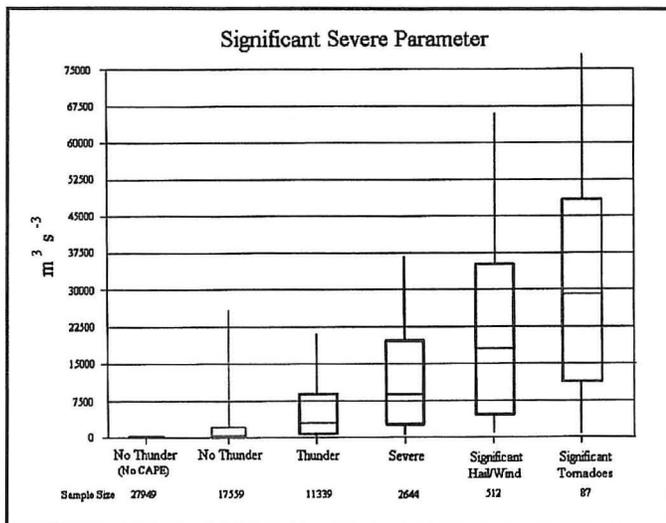


Fig. 13. As in Fig. 1, except for significant severe parameter ( $\text{m}^3 \text{s}^{-3}$ ).

- d. There is minimal seasonal variation in 0-1 km shear and MLLCL height for significant tornadoes. Considerable seasonal variation is noted in the other five categories. In addition, these parameters are better at discriminating events during the warm season.

## 6. Future Work

The dataset of significant severe events was relatively small in the present study. Additional 0000 UTC soundings from 1957 to 1996 are being compiled and examined to test the results of the three-year study against a much larger dataset. Subsets of the data to determine regional variability are also under inspection.

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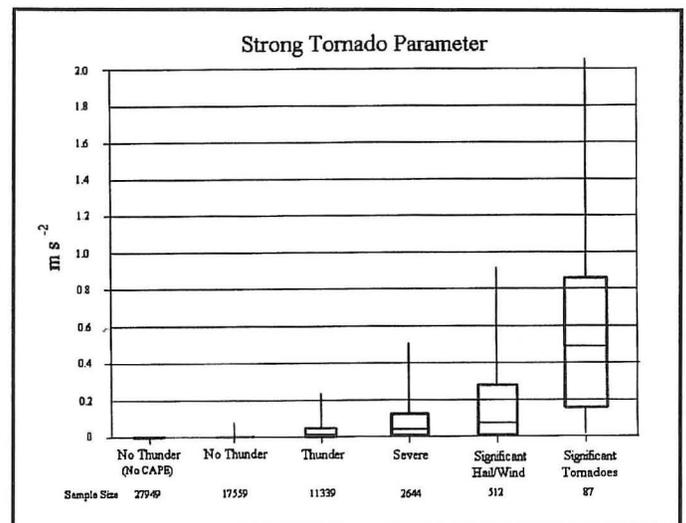


Fig. 14. As in Fig. 1, except for strong tornado parameter ( $\text{m s}^{-2}$ ).

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**Appendix: Objective Severe Weather Detection****Given Thunder, Frequency of Occurrence (%) Cape \* Shear**

SSP	Thunder	AISVR	Svr	Sig	SigTorn	%Total obs	Total
10000	62	38	29	8	1	13	4147
20000	53	47	32	12	3	6	1981
30000	46	54	35	15	4	3	1053
50000	41	59	32	21	6	1	309

**Statistics for detection of any severe event given a sounding with following thresholds**

SSP	POD	FAR	BIAS	CSI	HSS
10000	0.48	0.76	1.97	0.194	0.273
20000	0.30	0.70	0.97	0.172	0.255
30000	0.17	0.66	0.51	0.130	0.201
50000	0.06	0.64	0.15	0.028	0.084

**Given Thunder, Frequency of Occurrence (%) Strong Tornado Parameter**

STP	Thunder	AISVR	Svr	Sig	SigTorn	%Total obs	Total
0.25	56	44	31	9	4	2	1301
0.50	52	48	32	9	7	1	596
0.75	50	50	33	9	8	0.5	314
1.00	52	48	30	10	8	0.3	196

**Statistics for detection of Strong/Violent Tornadoes given a sounding with following thresholds**

STP	POD	FAR	BIAS	CSI	HSS
0.25	0.60	0.96	35.7	0.020	0.072
0.50	0.45	0.93	6.85	0.061	0.112
0.75	0.29	0.92	3.61	0.066	0.123
1.00	0.17	0.92	2.25	0.056	0.104

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Legend: SSP - Significant Severe Parameter ( $m^3 s^{-3}$ )  
 STP - Strong Tornado Parameter ( $m s^{-2}$ )  
 POD - Probability of Detection  
 FAR - False Alarm Ratio  
 CSI - Critical Success Index  
 HSS - Heidke Skill Score