STUDY OF A LONG-LIVED SYMMETRIC SQUALL LINE IN SOUTHEAST BRAZIL

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ABSTRACT
A squall line propagating at 14.0 m s⁻¹ reached the metropolitan area of Sao Paulo at about 1200 UTC (Universal Time) on 1 October 1997 causing heavy precipitation and strong wind gusts. Synoptic and mesoscale features associated with this system are analyzed through satellite imagery, weather radar, surface observations and numerical simulation. Satellite imagery indicates the squall line formed ahead of an advancing cold front and propagated perpendicularly to the cold front movement. Radar data analysis shows the system presented a symmetric shape with two distinct regions of precipitation: a convective leading edge and a trailing stratiform precipitation area. Analysis of the convective available potential energy (CAPE) and wind shear roles on the storm development points to two distinct stages: an initial one when CAPE dominated the scenario and a second one when wind shear was more relevant. The Regional Atmospheric Modeling System is used in two-dimension simulation to study the dynamics and evolutions of the squall line. Numerical results agree well with observational and radar data. Storm features such as convective cells triggering by the gust front, multicellular structure and trailing stratiform region are well represented in the simulation.

Keywords: squall line, convective system, observational analysis, weather radar, numerical model simulation

RESUMO: ESTUDO DE UMA LINHA DE INSTABILIDADE SIMÉTRICA DE LONGA DURAÇÃO NO SUDESTE DO BRASIL
Uma linha de instabilidade simétrica de longa duração que atingiu o estado de São Paulo em primeiro de outubro de 1997 foi estudada por meio da análise de dados do radar meteorológico de São Paulo, da reanálise do modelo de previsão do National Center for Environmental Prediction (NCEP) e dos dados de superfície da estação meteorológica do IAGUSP. A velocidade de deslocamento da linha foi estimada em 14,0 m.s⁻¹ com base em dados de radar. Os resultados indicaram a existência de duas regiões com diferentes estruturas de precipitação no interior do sistema. Uma região convectiva na dianteira da linha de instabilidade e uma região estratiforme na sua retaguarda. A evolução do sistema pode ser dividida em duas fases com relacao à principal forçante na manutenção da convecção; a energia convectiva potencial disponível (CAPE) e o cisalhamento do vento em baixos níveis, respectivamente. O modelo numérico Regional Atmospheric Modeling System (RAMS) foi utilizado em simulações bidimensionais para obter indicativos sobre os mecanismos envolvidos na evolução do sistema. Os resultados obtidos na simulação foram coerentes com a análise observacional e dados de radar. A ação da frente de rajada na produção de novas células convectivas, a estrutura multicelular da região convectiva e o aparecimento da região estratiforme foram bem representados na simulação numérica.

Palavras-chave: linha de instabilidade, sistema convectivo de mesoescala, radar meteorológico, simulação numérica

1. INTRODUCTION
Symmetric and anti-symmetric bands of heavy rainfall followed by stratiform precipitation (Houze et al. 1990) are often observed from equatorial regions to midlatitudes. Browning and Ludlam (1962) developed one of the first models to explain the air flow within these squall lines. More recently, Houze et al. (1989) developed a conceptual model based on weather radar measurements to explain the ascending front-to-rear flow and the descending rear inflow within the trailing stratiform region. Numerical simulations by Fovell and Ogura (1988, 1989) highlighted important features of squall lines such as the cold air pool, the front-to-rear updraft tilting and the environment vertical wind shear. Radial weather radar measurements have also been used to analyze 3-D air flow and rainfall patterns inside convective systems (Smull and Houze 1987, Houze et al. 1989). Biggerstaff and Houze (1991a and b) proposed a more complete conceptual model for midlevel dynamic and rainfall structures within squall lines based on a high resolution network of surface station observations that stressed the impact of the relative horizontal vorticity and the interaction between the convective leading edge and the trailing stratiform region. Furthermore, the microphysics and thermodynamics of squall
lines were studied by Braun and Houze (1994) to explain a secondary rainfall maximum often seen in the trailing stratiform region. Mesoscale surface measurements by Johnson and Hamilton (1988) also indicated the relationship between surface low and high pressure systems and rainfall patterns around squall lines. Many other studies of squall lines integrating observations and theory have been published (Moncrieff and Miller 1976; Rotunno et al. 1988; Wicker and Wilhelmson 1995; Parker 1996).

Several squall lines have been studied in distinct regions of Brazil. These studies were based on limited surface, upper air, radar and satellite measurements to characterize their internal structure, their time evolution (Gandú 1984; Pereira Filho 1989), and their associated synoptic environment. The life cycle of squall lines were also studied through satellite measurements. Some other analyses of convective systems over South America include reports by Silva Dias (1989 and 1997).

In the recent past, due to a lack of observations, numerical simulations have been used to study squall lines over the Amazon region (Ferreira 1988; Ambrozzi 1989). Recently, a large set of observations were produced by the Large Scale Biosphere Atmosphere Experiment in Amazonia (LBA). The LBA studies have provided a greater insight into the dynamics, thermodynamics and microphysics of squall lines over equatorial regions (Pereira Filho et al. 2000).

This manuscript analyses the 1 October 1997 squall line (hereinafter SL) that crossed the State of Sao Paulo, southeastern Brazil, and caused heavy precipitation, strong wind gusts and floods in the Upper Tiete River basin. In the mature stage, the squall line was formed by a 20-km wide and 250-km long leading convective band, with rainfall rates exceeding 100 mm h\(^{-1}\) followed by a wider trailing stratiform region with rainfall rates of about 30 mm h\(^{-1}\). This squall line was similar to the ones reported by Smull and Houze (1985; 1987b), Jorgensen et al. (1997), Zhang et al. (1989), Trier et al. (1996), among others.

The goal of this paper is to give a detailed analysis of the synoptic environment; radar precipitation pattern, satellite and surface signature associated with the SL. An estimate of the storm propagation speed based on surface data is proposed and compared to radar data estimates. The roles of convective available potential energy (CAPE) and wind shear on storm development are discussed through the Bulk Richardson number (Ri) analysis. Finally we show the results of a two-dimensional numerical model simulation of the SL.

The numerical simulation was carried out to explore the dynamic mechanisms associated with the SL life cycle, especially the interaction between the gust front and warmer ambient wind ahead, the role of trailing stratiform precipitation and cool-air pool on the storm propagation. These mechanisms could not be examined directly through the observational data available. The symmetric nature of the SL, as discussed in the radar analysis section, allowed the use of 2-D domain simulation.

It is not the intention of this article to analyze more realistic 3-D numerical simulations, nor the impact of topography on convection associated with the SL. Such study would require an extensive observational dataset for model validation purpose, which was not available. Although simplistic, the 2-D simulation can still provide interesting insights into the circulation within and around the SL.

Section 2 describes the methodology used in this study. The storm’s environment characterization, including the Ri technique, is explained in 2a. The procedure to estimate propagation speed is explained in section 2b. Numerical model configuration and initialization method are described in section 2c. In sections 3a and b the synoptic environment and Ri analysis results are discussed. Sections 3c and d detail the surface signature and radar precipitation pattern associated with the storm. Model results are presented in section 4. Conclusions are given in section 5.

2. METHODOLOGY

2.1. Environment characterization

The National Center for Environmental Prediction (NCEP) Reanalysis Dataset (hereinafter Reanalysis) were used to study the synoptic scale environment associated with the SL. We examined the low-level wind flow and moisture convergence, high-level wind divergence and convective instability at 0000 UTC (Universal Time) for 01 October 1997 to draw a picture of the atmosphere conditions at the time SL formed. NASA Geostationary Operational Environmental Satellite West (GOES-8) infrared imagery at 0300 and 1200 UTC were also examined to support the Reanalysis.

A Bulk Richardson number analysis was also provided to determine the role played by convective available potential energy (CAPE) and the wind shear on the SL development. According to Weisman and Klemp (1982), wind shear and CAPE were two decisive variables in determining which type of storm may be produced at a particular time. Those variables could be combined into a parameter called bulk Richardson number (Ri):

\[
Ri = \frac{CAPE}{S^2}
\]

The numerator is defined as:

\[
CAPE = g \int_\text{LFC}^{\text{EL}} \frac{\theta(z) - \theta_a(z)}{\theta_s(z)} \mathrm{d}z
\]

where \(\theta_a(z)\) and \(\theta(z)\) are the environmental air potential temperature and the adiabatically ascending surface parcel potential temperature, respectively. Integration is performed from the Level of Free Convection (LFC) to the Equilibrium Level (EL). And g is gravity acceleration. The denominator of Eq. 1 is given by:

\[
S^2 = 0.5 \left( \bar{u}_{6000} - \bar{u}_{500} \right)^2
\]

where, \(\bar{u}_{6000}\) and \(\bar{u}_{500}\) are the density-weighted mean wind speeds over the lowest 6 km and 500 m, respectively. \(S^2\) is not only a measure of the wind shear in the lower troposphere, but can also be thought as an estimate of the inflow kinetic energy made available to the storm by the vertical wind shear.
If one assumes that the storm moves with the lowest 6 km density-weighted mean wind speed and that the lowest 500m density-weighted mean wind is a good representative of storm influx (Weisman and Klemp, 1982), then (3) can be rewritten as:

\[ S^2 = 0.5 \cdot (c - \mu_{500})^2 = 0.5 \cdot U^2 \]  

where \( c \) is the velocity of the storm and \( U \) is the low-level influx at the storm relative framework.

Numerical simulations by Weisman and Klemp (1982) have shown that the wind shear interaction with either the storm updraft or storm cold pool could determine the degree of organization and longevity of the convection and hence the probability of severe weather. According to the article, when buoyancy was the very strong and wind shear was weak the resultant cell lifetime was limited. Storms produced by such cells were therefore relatively short-lived. On the other hand, when wind shear was stronger, storms tended to last longer and usually presented a more organized structure leading to the formation of mesoscale convective systems, such as squall lines.

Average CAPE, \( S^2 \), and \( Ri \) fields, extracted from the Reanalysis, were examined to explore the mean state of the atmosphere during the storm event. Averages were performed between 0000 and 1200 UTC for 1 October 1997. Although the Reanalysis data does not have the appropriate temporal and spatial resolutions to capture the impact of a mesoscale system in the environmental circulation, the authors believe that the examination of mean \( Ri \) field might provide a insightful characterization of the regional average atmospheric instability condition experienced by the SL.

### 2.2. Radar and surface signature analysis

Two weather radars made measurements of the 1 October 1997 squall line, namely, the Bauru weather radar (BWR) and the Sao Paulo weather radar (SWR) located in central and eastern Sao Paulo State, respectively (Fig. 1). Both radars are S-band systems but only the BWR has Doppler capability. BWR was used to analyze the initial and developing stages of the squall line while SWR the mature and decaying stages.

Surface measurements of air pressure, temperature, wind and precipitation at the Instituto Astronomico e Geofisico of the University of Sao Paulo weather surface station (IAG in Fig. 1) were used to analyze the surface signature of the squall line during its mature. This dataset was also used to estimate the SL propagation speed. Seitter (1986) shows that surface cold outflows of mesoscale convective systems behave as density currents. The driving force of the density current is the pressure gradient force resulting from the increased hydrostatic pressure in the cold air. The cold outflow speed is related to the velocity at which the convective system propagates. The leading edge of such outflows, also known as gust fronts, works as a lifting mechanism on the warm moist-laden air ahead of the storm creating new cells, which will then feed the convective system (Parker, 1996; Fovell and Ogura, 1988 and 1989). According to Seitter (1986) the speed of a cold pool \( V_p \) can be estimated by:

\[ V_p = k \left( \frac{\Delta p}{\rho} \right)^{0.5} + 0.62 \bar{U} \]

where, \( \Delta p \) is the surface pressure difference between the density current and the environment (Pa); \( \rho \) is the air density ahead of the cold pool (kg m\(^{-3}\)); \( k \) is the internal Froude number equal to 0.79; and \( \bar{U} \) is the ambient wind parallel to the current motion in the warm air ahead, positive in the direction of the current motion (m s\(^{-1}\)). Simpson and Britter (1980) showed that

![Figure 1: a) Map of South America and the state of Sao Paulo in the detail. Country boundaries as well as Brazilian state boundaries are shown. b) Locations of the Bauru weather radar (BWR), the Sao Paulo weather radar (SWR) and, the IAG/USP surface weather station (IAG). AB line represents the direction of cross-sections shown in Fig. 9.](image-url)
the component of the ambient wind parallel to the current motion could not be simply added to \( V_p \). Instead they found out through laboratory experiments that the ambient wind value must be reduced by a factor of 0.62. To determine the value of \( k \) Seitter compared 20 observed gust fronts events and found though a best-fit line \( k \) to be equal to 0.79 with a 0.84 correlation coefficient. The speed of the SL’s cold pool obtained from Eq. 5 is compared to the one derived from radar data.

### 2.3. Numerical simulations

Numerical simulations were performed to further investigate the life cycle of the SL on a 2-dimensional domain. The Colorado State University Regional Atmospheric Modeling System (RAMS) was used in this study (Tripoli and Cotton 1982; Pielke et al. 1992) on an X-Z domain setup (Table 1). Through the analysis of radar precipitation pattern presented in section 3d, the SL was found to be symmetric in relation to the position of the leading convective cells and the trailing stratiform region (Houze et al. 1990). Such symmetric nature allowed the use of 2-D simulations in this study. The horizontal dimension of the domain extends 600 km with a 1-km resolution. The vertical structure was described by 35 model levels with a telescopic resolution ranging from 50 m, in the lowest layer, up to 1000 m in the upper part of the domain stretching a total of 17.5 km high.

The initial temperature and moisture condition for the simulation were based on the Sao Paulo airport sounding at 1200 UTC for 1 October 1997 (Fig. 2). The sounding site is about 5 km north of the IAG station (Fig. 1). Analysis of temperature and humidity profiles showed that convection had probably been initiated when sounding was done. Nevertheless, the profiles were used to initialize the simulation since no other soundings were available to the authors and because the sounding still showed strong enough instability to sustain deep convection. The simulation was carried out without topography and included a complete 3-phase microphysics scheme. Convection was initiated in the simulation with a small perturbation in the potential temperature (+5 K) and the vapor mixing ratio (+20% g/kg) fields over an area 10 km wide and 1.2 km deep placed right above the surface. This technique was discussed by Fovell and Ogura (1988). A 24-hour simulation was done starting at the Sao Paulo airport sounding time.

The relative vorticity (RV) equation was used to explore two-dimension aspects of the SL life cycle based on model results and the observation data. If both Coriolis and friction effects are neglected, the RV equation on the XZ plane can be expressed as:

\[
\frac{D \xi}{Dt} = \frac{\partial \xi}{\partial t} + \vec{V} \cdot \nabla \xi = \frac{\partial B}{\partial x} \tag{6}
\]

\[
\xi = \nabla \times \vec{V} = \left( \frac{\partial u}{\partial z} - \frac{\partial w}{\partial x} \right) \hat{j} \tag{7}
\]

where, \( \xi \) is the relative horizontal vorticity, \( u \) and \( w \) are the horizontal and vertical wind components, respectively. \( B \) is the buoyancy forcing term:

\[
B = -g \frac{\rho - \rho_0}{\rho_0} \tag{8}
\]

where, \( g \) is the gravity acceleration, \( \rho^* \) and \( \rho_0 \) are the actual and the reference density, respectively. According to equations 6 to 8 relative vorticity can only be produced by horizontal gradients of buoyancy and redistributed by advection on a XZ plane. The role of RV in the simulated storm’s evolution was also explored.

### 3. RESULTS

#### 3.1. Synoptic-scale analysis

The GOES-8 IR imagery shows a NW-SE area of cloudiness at 0300 UTC for 1 October 1997 (Fig. 3a) with colder tops (brighter tons) associated with a surface cold front extending from Paraguay to southern Brazil. To the north of the cold front there was a region of colder cloud tops located in the northwestern corner of Sao Paulo. Fig. 3a shows what was believed to be the first indication of the October 1st squall line.
Fig. 4a shows the 850-hPa wind field reanalysis at 0000 UTC for 01 October 1997. An area with strong winds stretching from east Bolivia through Paraguay and western Mato Grosso do Sul could be seen with speeds greater than 16 m s\(^{-1}\). High-humidity content air was being advected from the Amazon Basin region into southern and southeastern Brazil. The 850-hPa humidity divergence field (Fig. 4b) shows humidity convergence greater than 2.0 \(10^{-4}\) g kg\(^{-1}\) s\(^{-1}\) to the northwest of Sao Paulo. The more intense humidity convergence (4.0 \(10^{-4}\) g kg\(^{-1}\) s\(^{-1}\)) area to the south was associated with the surface cold between Southern Brazil and Paraguay.

Two areas of intense upper level mass divergence were found in the region. One was located over the border between Argentina and Brazil, above the surface cold front, and a second one farther north over western Sao Paulo (Fig. 4c). Fig. 4d shows that convective potential instability was high over most of western Sao Paulo and eastern Mato Grosso do Sul, where CAPE was higher than 1000 J kg\(^{-1}\).

The SL reached the coast of Sao Paulo at 1200 UTC 1 October 1997 (Fig. 3b). The SL propagated perpendicularly to the advancing movement of the cold front as it moved across the state. Cloud free areas to the north and to the south of the surface cold front indicated the presence of high pressure systems that were advecting warmer and moister air southward and colder and drier air northward, respectively. In between these two areas, the cold front was less evident and apparently did not move significantly, probably due to a blocking pattern imposed by the Atlantic subtropical high to the north. This synoptic scale subsidence mechanism could be responsible for the decay of the SL as it moved eastward across the state.

3.2. Bulk Richardson number analysis

Mean CAPE, \(S^2\), and Ri are shown in Fig. 5. Averages were performed between 0000 UTC and 1200 UTC for October 1, 1997. CAPE was about 1800 J kg\(^{-1}\) where the SL was formed and decreased towards the coast of the southeastern Brazil (Fig. 5a). Values around 1500 J kg\(^{-1}\) were found in the central part of the state of Sao Paulo where the storm reached its mature stage and dropped to less than 600 J kg\(^{-1}\) in the eastern part of the state. Consequently, as the SL moved across the state, it experienced a significant CAPE decrease. The storm mean inflow kinetic energy field (\(S^2\)) is shown in Fig. 5b. Unlike CAPE, it increased eastward from 60 J kg\(^{-1}\) in the west to over 140 J kg\(^{-1}\) in the Mantiqueira region and 130 J kg\(^{-1}\) along the coast. Therefore, SL experienced higher inflow kinetic energy as it moved across the state.

For comparison, the CAPE and storm inflow kinetic energy calculated from the sounding in Fig. 2 were equal to 400.0 and 144.0 J kg\(^{-1}\), respectively. The Ri field is shown in Fig. 5c. Higher values were found over western Sao Paulo and lower ones over the eastern part of the state. Therefore, the earlier stages of the SL were supported by a highly unstable atmosphere over the western part of the state of Sao Paulo. Once these initial storms developed and started moving eastward, a more organized structure began to develop due to low-level
inflow of moist air and a more prominent interaction between wind shear and updrafts. As the storm system reached its mature stage, the interaction between the inflow and the cold pool was sufficient to keep the storm going, even though it had entered a low-CAPE environment. This process continued as the SL advanced eastward until it dissipated over the Atlantic Ocean.

3.3. Surface signatures of the squall line.

Fig. 6a shows the time evolution of the rainfall rate measured at IAG (Fig. 1) between 1100 UTC and 1300 UTC. The rainfall started at 1220 UTC and reached a maximum rate of 68.0 mm h⁻¹ at 1240 UTC. This first maximum was possibly associated with the convective cells in the leading edge of the SL. Afterwards, the rainfall rate decreased rapidly to less than 10 mm h⁻¹ at 1310 UTC. The secondary rainfall rate maximum of 15 mm h⁻¹ at 1340 UTC was associated with the stratiform precipitation area of the SL. Surface winds shown in Fig. 6b indicates the SL gust front moved across the weather station site at around 1200 UTC, with winds gusting at over 9.4 ms⁻¹, and shifting direction from NNE to W. The gust front anteceded the rainfall by 20 minutes. The wind turned south once rainfall began and remained this way throughout the stratiform region passage.

The time evolution of surface pressure and air temperature between 1100 UTC and 1600 UTC are shown in Fig. 6c and d, respectively. After the gust front, pressure increased 2 hPa in 10 minutes while air temperature decreased (Fig. 6d). These two features were associated with the advancing of the storm’s cold-air pool. According to Johnson and Hamilton (1988), the cold-air pool is created by evaporative cooling and mid-troposphere air dragged downward by intense precipitation. The high-pressure area associated with the cold pool is called meso-high. Surface air temperature and pressure decreased after 1200 UTC. The air pressure decreased 3.5 hPa in about 2 hours. This indicated the approaching of the meso-low, which
was believed to be a result of mid troposphere subsidence warming (Johnson and Hamilton, 1988). A secondary surface pressure increase of 1 hPa in 10 minutes at 1500 UTC reflected the meso-low moving away from the surface weather station site. Afterwards, the surface air pressure returned to the existing pre-frontal synoptic conditions.

An estimate of the SL speed of propagation can be obtained through Eq. 5 if it were assumed that the surface wind was a good approximation of the wind speed in the warm air ahead of the storm. In this case we would have $\Delta p = 2.86 \text{ hPa}$; $\rho = 1.09 \text{ g m}^{-3}$; $k = 0.79$ e $\overline{U} = -0.4 \text{ m s}^{-1}$, which applied to Eq. 5 yielded 12.6 m s$^{-1}$.

Figure 5: NCEP Reanalysis mean a) convective available potential energy [J kg$^{-1}$]; b) inflow kinetic energy [J kg$^{-1}$]; and c) bulk Richardson number. Means were taken between 0000 and 1200 UTC for 1 October 1987.

Figure 6: Time evolution of a) rainfall rate [mm h$^{-1}$]; b) wind speed [m s$^{-1}$] and direction; c) surface pressure [hPa] and d) surface air temperature [$^{\circ}$C] between for 01 October 1997. Measurements from the IAG/USP surface weather station indicated in Fig. 1.
3.4. Weather radar analysis.

The first echoes from the SL were measured by the BWR at about 0300 UTC. By 0800 UTC, the SL was overhead the BWR. Fig. 7a and b show 0.3° plan position indicators (PPI) for the reflectivity at 0800 UTC and radial velocity fields at 0830 UTC for 01 October 1997, respectively. The reflectivity field showed a NE-SW band of high rainfall rates at the leading edge of the SL. An area of weaker echoes was seen to the SW of the SL. The radial velocity field showed a bipolar structure that indicated the eastern section of the SL had already passed by the radar at that time while a trailing section was yet to come.

Reflectivity and radial velocity vertical cross-section are also shown in Fig. 7. The reflectivity cross-section at 0800 UTC showed a multi-cellular storm structure with separate precipitation cores stretching up 5 km into the storm. The radial velocity cross section showed a distinct wind configuration in the rear part of the SL where wind was moving towards the BWR with speeds over 18 m s⁻¹. The downward rear-to-front jet descended as it approached the leading edge of the storm. According Smull (1987) and Biggerstaff (1991) such jet was a major factor in sustaining convection in long-lived squall lines by increasing the low-level convergence zone underneath the storm’s convective leading edge.

Figure 7: Bauru weather radar a) 0800 UTC reflectivity 0.3° PPI, and vertical cross-section [dBZ]; b) 0830 UTC radial velocity 0.3° PPI, and vertical cross-section [m s⁻¹] for 01 October 1997. Cross-section direction is indicated by the thick black AB line in panel b. White cross in the center of PPI indicates the radar position.
The SL was also measured by the SWR. Fig. 8 shows a sequence of constant altitude PPIs or CAPPI from 1200 UTC to 1400 UTC. One can notice that the SL moved eastward with a small northward component. A line of convective rainfall rates over 75 mm h\(^{-1}\) was trailed by a larger area of stratiform rainfall rates between 20 and 30 mm h\(^{-1}\). A transition zone was apparent between the convective and the stratiform regions where rainfall rates were lower than 10 mm h\(^{-1}\). The trailing area enlarged as the SL moved eastward but decayed once over the Atlantic Ocean at 1400 UTC. The precipitation patterns were consistent with Houze (1985) who classified this type of SL as symmetric. The mean speed of the SL estimated from Fig. 8 was 14 m s\(^{-1}\). This number was slightly higher than the speed calculated with the density current approximation (12.6 m s\(^{-1}\)).

Fig. 9 shows vertical cross-sections of rainfall rates along the line AB in Fig. 1. The intense convective leading edge and the stratiform rear band are seen in the sequence of cross-sections. The convective leading edge was 20.0 to 30.0 km wide and 14 km deep (Figs. 9.b and c) with rainfall rates higher than 50 mm h\(^{-1}\) and at some locations greater than 125 mm h\(^{-1}\). As the squall line moved eastward, new cells developed ahead of it. These new cells intensified while advected rearward to form a region of intense rainfall rates.

Older cells continued to move rearward to form a region of stratiform precipitation 60.0 to 90.0 km wide with rainfall rates between 10 and 40 mm h\(^{-1}\). An intense vertical gradient of rainfall rates can be seen between 4 km and 5 km high in the stratiform region. This bright band was probably formed by ice particles and snow as they melt after crossing the 0°C isotherm to form rain droplets and or drops.

The transition zone indicated as a low-precipitation region behind the convective leading band (Fig. 9c) was created by mass compensation mechanism, which reduced the growth of ice particles and increased the evaporation of rain droplets. This mass compensation mechanism was predominately caused by the updrafts in the leading edge of the SL and may have also been reinforced by evaporative cooling according to Braun and Houze (1994).

Figure 8: Sao Paulo weather radar 1.5-km precipitation rate CAPPI at a) 1200, b) 1230 UTC, c) 1300 UTC, and d) 1400 UTC [mm h\(^{-1}\)] for 01 October 1997. Cray scale indicates intensities. Black shaded areas represent the Mantiqueira Mountain Range.
3.5. Numerical simulation

Results from 2-D numerical simulation of the SL are shown in this section. Only horizontal wind, liquid mixing ratio and relative vorticity fields are shown. A few air parcel trajectories were computed to identify major features within and around the SL. The simulated SL moved from left to right along the domain.

Fig. 10 shows the horizontal component of the wind. Intense convergence below and divergence aloft were seen at the leading edge of the SL. Ahead of the storm, moisture-rich air confined to a layer 2000-m thick above the surface moved toward the storm. This concentrated flow clashed with the storm’s cold outflow to form an area of intense low-level convergence, which marked the convective leading edge of the SL. Analysis of vertical wind field show that at least three convective cells formed at the low-level convergence zone. These cells were advected rearward as the SL moved forward, which resembled the multicellular structure seen in the radar data analysis.

Low-level wind in the rear part of the SL became faster as it approached the convective leading edge. The rear-to-front jet slightly tilted downward as it approached the convergence zone. The descending rear-to-front jet stretched in altitude from about 4000 m to about 1000 m in the first 100 km behind the leading edge of the SL. This rear-to-front jet became significantly weaker over the 12 hours of simulation. This feature agrees with the BWR radial velocity data shown in Fig. 7b.

An equivalent potential temperature field analysis at 12 hours of simulation (not shown) indicated that most of the highly humid air (over 345 K) was located ahead of the storm and was restricted to the first 0.5 km above ground. At the rear of the storm, air was potentially cooler, especially between 1.0 km and 4.0 km over the first 100 km behind the leading edge, where values were lower than 338 K.
Figure 10: Horizontal wind field after a) 3, b) 6, c) 9 and d) 12 hours of simulation [m s⁻¹].

Figure 11: Same as in Fig. 10 for liquid water mixing ratio [g kg⁻¹].
Fig. 11 shows liquid water mixing ratio fields between 3 and 12 hours of simulation. At the beginning of the simulation, water was concentrated mostly around the convective leading edge. As the SL progressed eastward, the liquid water content developed a trailing region. Two distinct areas of moderate to high liquid water content formed after 2 hours of simulation (Fig. 11d).

If liquid water mixing ratio is assumed proportional to rainfall, as the SL moved, two areas of rainfall maxima were generated; one at the convective leading edge and another in the rear part of the SL around 100 km from the first one. The first rainfall maximum was associated with the main updraft area, while the second was linked to the stratiform trailing area of the storm. These results agreed well with radar observations. A narrow band without rainfall at the surface developed between the two main precipitation areas (Figs. 11c and d) where the transition zone was observed in the radar reflectivity analysis.

The simulated spatial and temporal patterns of the liquid water mixing ratio resembled well the SL reflectivity pattern. Precipitation features were well represented by the 2-D model simulation. Cotton and Anthes (1989) suggested that 2-D numerical simulations tend to overestimate and underestimate precipitation due to the strength of low level moisture convergence and the entrainment of dry air in the storm cloud layer, respectively.

Relative horizontal vorticity fields are shown in Fig. 12. Light shading areas represent regions with positive (clockwise) relative vorticity. Between 1500 and 5000 m along the storm trajectory a vorticity dipole pattern was formed around the SL updraft with positive values ahead and negative values behind it. According to Eq. 6 such dipole were created by strong horizontal buoyancy gradient associated with the unstable air in the updraft. The dipole played an important role in keeping a straight updraft, which in turn, helped maintain the leading edge convection.

An area with intense negative vorticity was found between surface and 1500 m ahead of the storm. This area corresponds to the SL gust front. The cool pool behind the gust front was moving away from the storm while the warm and moist air in the undisturbed environment ahead of the storm was moving towards it. According to Eq. 6 this buoyancy difference generates negative RV. The advection of negative RV by the updraft towards the rear of the storm may be linked to the generation of the descending rear-to-front jet seen in the horizontal wind field, since both features became stronger at the same time as the storm intensified. Fig. 12 shows that the negative RV at the gust front became stronger over time (Figs. 12a to d).

Finally, air-parcel trajectories associated with the air flow within the SL are shown in Fig. 13. The starting points of each trajectory were chosen arbitrarily at the front and at the back of the SL to illustrate main circulation patterns. Trajectories A, B and C represent air parcels with high buoyancy initially located ahead of the storm, while trajectories D, E and F represent air parcels initially located at the back of the convective system.

Figure 12: Same as in Fig. 10 for relative vorticity [s⁻¹] centered at the leading edge of the squall line.
The first set of trajectories shows parcels being lifted initially at the gust front and carried by the updraft to over 8,000 m of altitude. Eventually, parcel A turned east after being lifted and moved forward ahead of the storm. Parcels B and C turned to the back of the storm after being lifted. Air parcels following trajectory A formed the storm anvil cloud, while those following trajectories B and C fed the trailing stratiform precipitation region.

The second set of trajectories shows air moving from the rear part of the storm to the convective region. All three trajectories descended as they moved forward. Trajectory F eventually reached the surface and diverged rearward, while trajectory E kept moving down and forward, passed the intense convective zone and finally reached the gust front. Air parcel paths E and F illustrated relevant features of long-lived trailing stratiform precipitation squall lines, namely the descending rear-to-front jet that contributed to the storm longevity by bringing dry cool air towards the gust front and the divergence zone located behind the intense convective convergence at the leading edge of the system (Biggerstaff, 1991).

4. CONCLUSION

On 1 October 1997 an intense symmetric squall line crossed the State of Sao Paulo, southeastern Brazil, causing heavy precipitation, strong wind gusts and flashfloods. The squall line was formed by a 20-km wide and 250-km long leading convective band, with rainfall rates exceeding 100 mm h\(^{-1}\), followed by a wider trailing stratiform region with rainfall rates of about 30 mm h\(^{-1}\) during its mature stage.

Low-level moisture convergence, high CAPE and mass divergence aloft associated with a moving surface cold front as indicated by NCEP reanalysis over western Sao Paulo were the main triggering mechanisms of the squall line.

Surface air pressure, temperature, humidity, rainfall rate and wind measurements at IAG in eastern Sao Paulo during the mature phase of the squall line indicated well known patterns, namely the meso-low, meso-high, and the cold pool similar to those of mid-latitude squall lines. The squall line speed of propagation estimated from surface measurements and calculated from radar data were 12.6 m s\(^{-1}\) and 14.0 m s\(^{-1}\), respectively. Although a rough approximation, the speed calculated from the surface measurements was a good indicator of the storm propagation velocity and could be used in cases where weather radars were not available and probably could be used in nowcasting decision-taking situations.

Precipitation patterns obtained from the weather radars indicated three distinct regions: 1) a convective band at the leading edge of squall line; 2) an almost rainfall-free region or transition zone; and 3) a larger area of trailing stratiform precipitation. At the leading edge rainfall rates within cells were over 100 mm h\(^{-1}\) while at the stratiform region the rates were between 10 and 40 mm h\(^{-1}\). As new convective cells formed along the expanding gust front under unstable air, older cells were advected rearward and formed a region of stratiform precipitation.

The early development and maintenance of the squall line was associated with high Ri. Buoyancy was the major thermodynamic forcing. As the system moved eastward, kinetic energy inflow became more important as CAPE decreased rapidly. Results suggested that a balance between these two forcing might have been reached at some point during the system life cycle, what would have promoted the SL’s longevity.

Numerical simulation results were consistent with the observations. Simulated liquid water mixing ratio patterns were similar to the ones obtained from radar estimates. The three rainfall regions, convective leading edge, trailing stratiform region, and transition zone were correctly simulated. Air parcel trajectories were also in agreement with other works in the literature. The downward rear-to-front jet was also simulated by the model, which agreed with the weather radar analysis and Biggerstaff (1991). The trajectories reproduced the backward tilting of the updrafts at the leading edge, which indicated an increase in relative vorticity being advected rearward as the storm progresses. This result was coherent with the Ri analysis.

The squall line examined in this study constituted a case study over South America where high-resolution surface data are still somewhat limited. Nonetheless, it exhibited major mesoscale mechanisms associated with long-lived squall lines. Its symmetric nature allowed the use of a 2-D model simulation to complement the analysis. Both simulations and observations were in agreement. Often, squall lines over the eastern Sao Paulo area are slowed by the early afternoon sea breeze that also tends to reinforce the source of moisture as well as lower level wind shear, therefore causing severe floods in the Metropolitan Area of Sao Paulo with great social and economic impacts. Furthermore, in some instances, after the passage of the squall line, precipitation produced by the associated cold front can increase the risks for floods.

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