Proceedings of 8 ICSHMO, Foz do Iguaçu, Brazil, April 24-28, 2006, INPE, p. 1417-1425.

LOCAL ATMOSPHERIC CIRCULATIONS IN THE AMAZONAS RIVER MOUTH

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1. INTRODUCTION

The estuarine systems have an historic and fundamental importance for the development of mankind. About 2/3 of the planet mega-cities, for example, are in proximities of estuaries. These systems are responsible by the supply of important nutrients for the development of life in the oceans, and in these systems are located great harbor poles, as an example we have Santana Harbor in Amapá and Belém Harbor in Pará.

The region of the Amazon River mouth is rich in mangrove swamps, the main responsible by the protection of the cost line from storms, besides being a protection for several marine and estuarine species. These environments are over the influence of meteorological phenomena that vary since a micro-scale till a synoptic scale. We can notice that the main tempest causer phenomenon is the squall lines.

According to Cohen (1989), the squall lines (SL) that are formed in the Atlantic Cost of Amazon are responsible by the formation of rain near the cost of the States of Pará and Amapá, as well as in Central Amazon. The SL is characterized for having a great conglomerate of cumulus-nimbus clouds and it is formed through the coast because of the circulation of sea breeze.

The shift of SL in east direction to west is due to a phenomenon called east waves, that according Riehl (1954), are oscillations in pressure fields and wind that are in phase in the surface, which carry the perturbations in its propagation direction. According to the author, these waves have a phase speed of 6° of longitude a day, period from 3 to 4 days and wave horizontal length from 2000 km to 3000 km. Ezpinoza (1996), using meridional wind data, for a period of 10 years, showed that these ondulatory disturbances exist through all the year with different wave length and that during the summer (DJF), autumn (MAM) and winter (JJA) they move themselves arriving in the north northeast coast of Brazil. According to the same author, on summer the wave length is from 6000 km to 7000 km, and the phase speed is from 10m/s to 14m/s and on autumn the wave length is from 5000 km to 6000 km and the phase speed if of 10 ms^{-1} to 13 ms^{-1} and on winter the wave length are shorter, varying from 3500 km to 4000 km and the phase speed from 10 ms^{-1} to 13 ms^{-1}.

As notice previously, the marine breeze is one of the main forming mechanisms of SL. Among the local wind systems, the sea, lacustrine and land breezes are typical systems of littoral regions, or near to great water bodies. This kind of circulation is basically induced by the difference of daily heating between the land and water surface. It is also influenced by the topography, by the coast curvature, by the local latitude, besides of the synoptic conditions and the climatologic circulation of large scale (Saraiva, 1996).

The circulations of breeze kind are known as local circulations. Their formation is basically due to the difference of temperature between the land and water surface. During the day the land receives about 51% of all the radiation emitted by the sun. This percentage will heat the land and water surface, but this heating is in a different way because the land and the water have different capacities to store the heat. During the day the land have to heat itself faster than the water, occurring then a difference of temperature between them.

According to Kousky and Dias (1982), the high specific heat and the high water transparence act together maintaining the surface temperature in an almost constant value. Part of the heat received by the water is also used in the process of evaporation making that the surrounding air stay relatively cold.

Due to this temperature difference between the land and water surface, there will be a circulation with its direction according to the temperature gradient. During the day the land surface is hotter than the water surface, creating then a temperature gradient that goes from the water to land. With the continent surface heating, the air near it tends to expand, creating then an inclination in the isobaric surfaces. One force will tend to compensate this volumetric growth on the continent, bringing then the air that is over the ocean in direction to the continent, creating then the breeze circulation. During the day this circulation will have a surface component pointing to land, and in higher levels component pointing to the water. But during the night period this temperature gradient will have an inverted direction, as a consequence of the conservation of the water temperature and the decrease of the temperature on the land surface. This circulation will go from the land to the water on the surface and in higher levels from the water to the land.

It is known that among other factors, the breezes are a great importance phenomenon to the characterization of the dispersion conditions of polluters, due to the effects of circulation associated to them, that cam moisture existing polluters over the ocean with other existing in the local atmosphere, that facilitates the production of photochemical oxidants (Saldanha, 2003).

The study area is located in the North region of Brazil, between the latitudes 3°S and 2°N and between the longitudes 53°W and 49°W, formed by the lower part of Amazon and Tocantins-Pará rivers and by part of the continental platform near the mouth of both (Figure1).

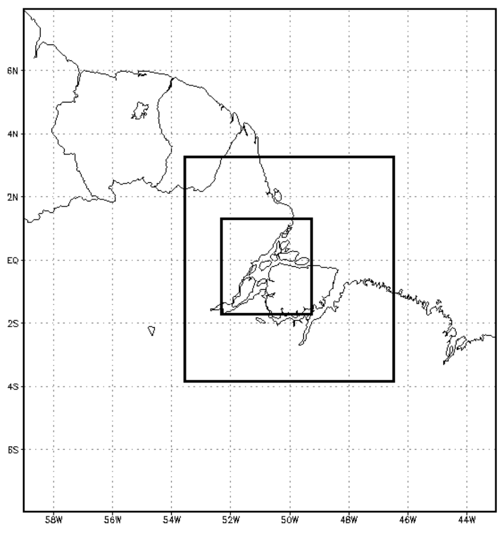


Figure 1. Image showing the study region with the three grids of the model generated in the simulation.

This work has as a general objective to study the local circulations in north region littoral of Brazil, more specifically in the region of Amazon mouth, using an atmospheric model of mesoscale, characterizing the influence of topography, vegetation and the kind of soil in the region on these circulations.

2. MATERIAL AND METHODS

This work analyzes simulations that were generated by an mesoscale atmospheric model. It is called Brazilian Regional Atmospheric Modeling System (BRAMS), in its 3.2 version. RAMS is a mesoscale model based on finite differences. It was developed by researchers of the University of the State of Colorado in the United States. Its objective is to gather several numerical simulation codes of time, using them together in the same place (Pielke et al, 1992). BRAMS is a developing project of parameters and development in general in the original numeric model of mesoscale. Nowadays several institutions are working together in the development of this model in a project called BRAMSNET, which is financed by Financiadora de Estudos e Projetos (FINEP).

BRAMS consists of an atmospheric numeric model and it is built in this way: a pre-processing data part prepares the observed meteorological data to the initialization through their borders, a processing package of atmosphere physical processes, and the post-processing that makes the output interface of the model with a variety of graphic packages.

The model has a varied initialization (through pre-processing data) and the homogeneous initialization consisting in the usage of atmospheric sounding that in the first moment is extrapolated for all grid. It was used in this work a sounding accomplished at 12 UTC in July 28, 2005 in the airport of Macapá. Figure 2 show the thermodynamic profile of the atmosphere obtained though the sounding data.

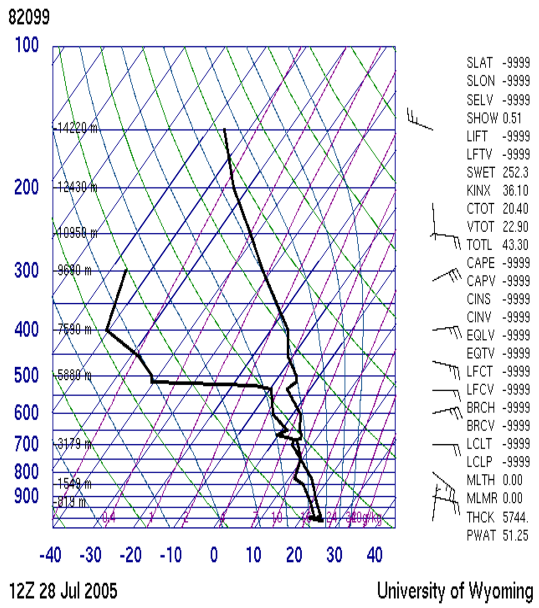


Figure 2. Vertical profile of the atmosphere obtained though the sounding realized at 12 UTC at July 28, 2005.

It was used three grids for the study, being two together grids. The first grid, that can be identified in figure 1, is centered at 0° of latitude and 51° W of longitude, with a spatial resolution of 20 km. The second and the third grids are centered at 0.2° S of latitude and 50.7° W of longitude and with 10 km and 2.5 km of spatial resolution respectively.

3. RESULTS AND DISCUSSION

In the simulation were generated hourly outputs for a period of 24 hours, starting at 12 UTC at July 28 till 12 UTC at July 29, which will be analyzed in the discussion. The direction and intensity of the wind data of the sounding were not used in the initialization, imagining that this data, being part of the large scale circulation, could inhibit the local circulations in the study region.

For the discussion it will be also used the GOES-12 satellite images for the studied period. They were picked from the CPTEC-INPE website to identify the formation of nebulosity associated to the coast breeze and estuary in Amapá.

As said previously, the breeze circulation is due to the temperature difference between the continent and the water body. We can observe at figure 3, in the first simulation time at 13 UTC, we do not have definition about any wind component that we can identify as a breeze, because it still not occurred an enough heating to be generated the circulation. The satellite image for this same time (figure 4) does not show any nebulosity to the studied area.

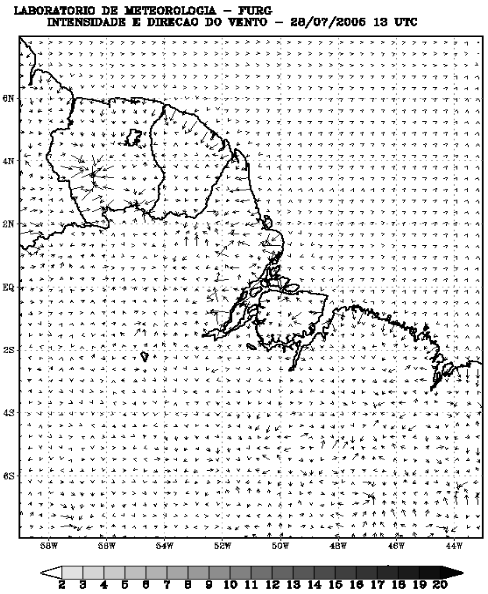


Figure 3: Prognostic of the direction and wind intensity fields at 13 UTC at July 28, 2005.

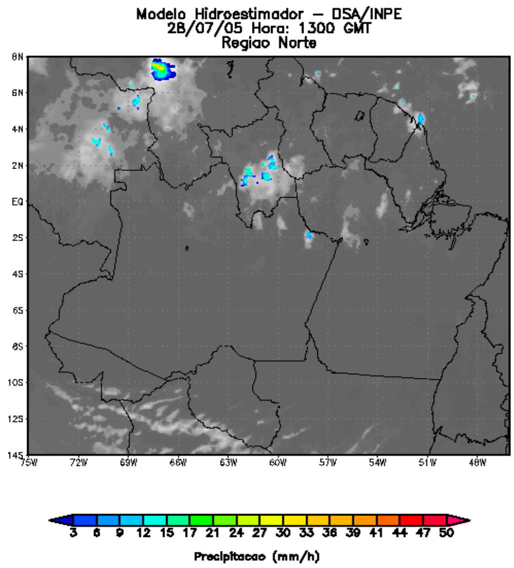
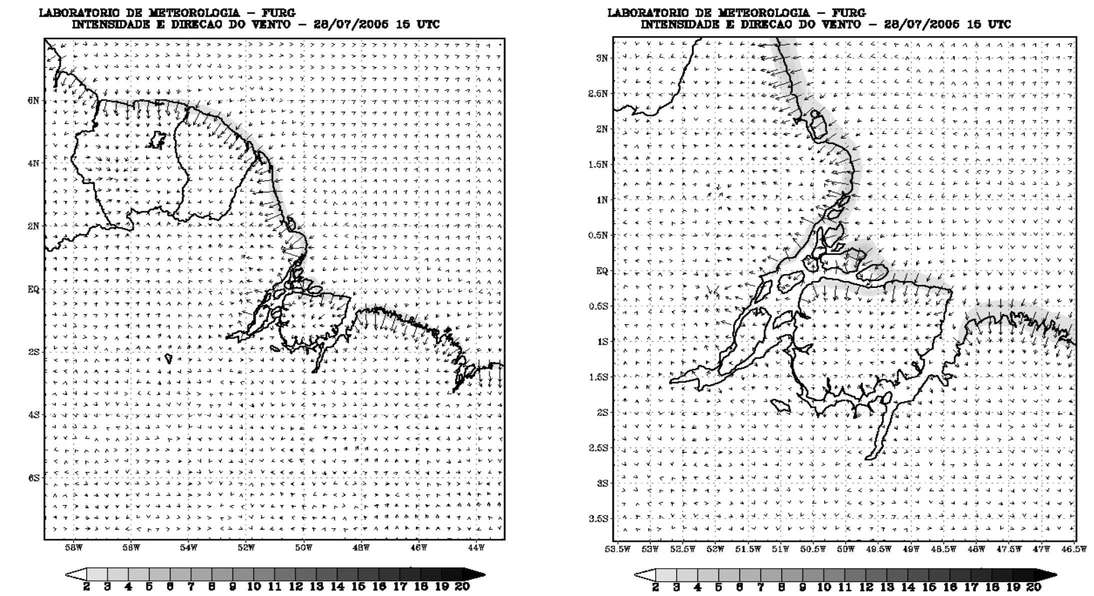


Figure 4: Satellite image at 13 UTC at July 28, 2005.

The greatest incidence of radiation over the land surface mainly occurs during the middle of the day when we have an almost right angle between the incident radiation and the Earth's surface. In this time we have a great heating of the Earth. With a greater heating, the temperature difference between the two surfaces starts to generate a wind component with ocean-continent direction. In the simulation its beginning is at 15 UTC, as can be observed in the figure 5a. In the figure 5b is showed the second grid used in the simulation where we can also observe that the sea breeze circulation starts to develop.

As simulation time passes, the temperature gradient starts to increase. Then it increases the breeze intensity as can be observed in figure 6, referring to 18 UTC at July 28, 2005. The direction of the wind component brings the humid air from the ocean in direction to the continent. The instable atmosphere due to the increase of the latent heating flow and to the humidity given to the lowest layer of the atmosphere makes the part of the air that is in the surface rises and condense in higher levels. Then it forms clouds near the coast, as can be observed in the satellite image for the same time (figure 7).

According to Atikinson (1981), in a stable atmosphere the higher layers act as a muffled mechanism to the circulation of sea breeze, although an instable atmosphere facilitates the horizontal and vertical circulation.



(a) (b)

Figure 5: (a) Prognostic of the wind direction and intensity fields at 10 meters for the 15 UTC, at July 28, 2005 (grid 1) (b) Prognostic of the wind direction and intensity fields at 10 meters for the 15 UTC, at July 28, 2005. (grid 2).

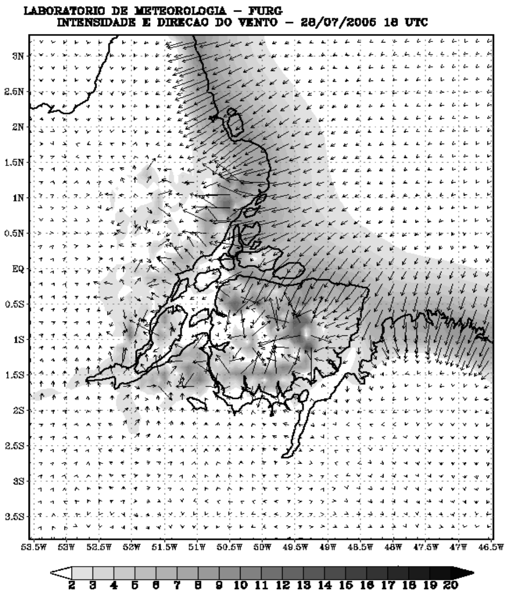


Figure 6: Prognostic of the wind direction and intensity fields at 10 meters for the 18 UTC at July 28,2005 (grid 2)

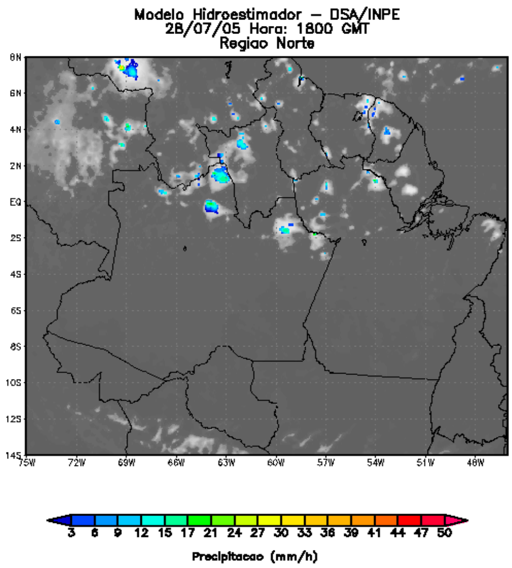


Figure 7: Satellite image of the 18 UTC, at July 28, 2005.

The breeze circulation, by its definition, has a main current on the surface, according to the temperature gradient, and it has a return current in higher levels forming then a circulation cell. Since the coastal line orientation in the studied region has the northeast-southeast direction, we cannot clearly identify the breeze in the vertical section that was done only for a wind component. So we analyzed the vertical sections done for the zonal (u) and meridional (v) components, both for the 20 UTC at July 28. The two sections for the u and v components were made in the 1.5°N latitude and with vertical coordinate z varying since 10 meters high to 4100 high.

Atkinson (1981) making reference to other authors, showed through observational studies, that were registered sea breeze intensity values which vary from 2.3ms^{-1} till values near 11ms^{-1}, and return current with intensity varying from 1 to 7ms^{-1}. The section made for the u component (figure 8a) shows the greater intensity of the sea breeze scoring 12ms^{-1}, It occurred at 150 meters high. The return current could also be observed in the section, and it has a speed 6ms^{-1} at approximately 3km high. The vertical section made for the v component of the wind (figure 8b) shows the sea breeze penetrating in the continent with intensity of 4ms^{-1} at approximately 200 meters high and the return current is visible at approximately 4.5 km high with intensity of 2 ms^{-1}. The analysis of these figures clearly shows the circulation cell formation of the sea breeze described in the introduction.

Frizzola and Fisher (1963), in observational studies realized for the area of the New York City, showed that the return current can be difficult to be encountered in the presence of large scale winds.

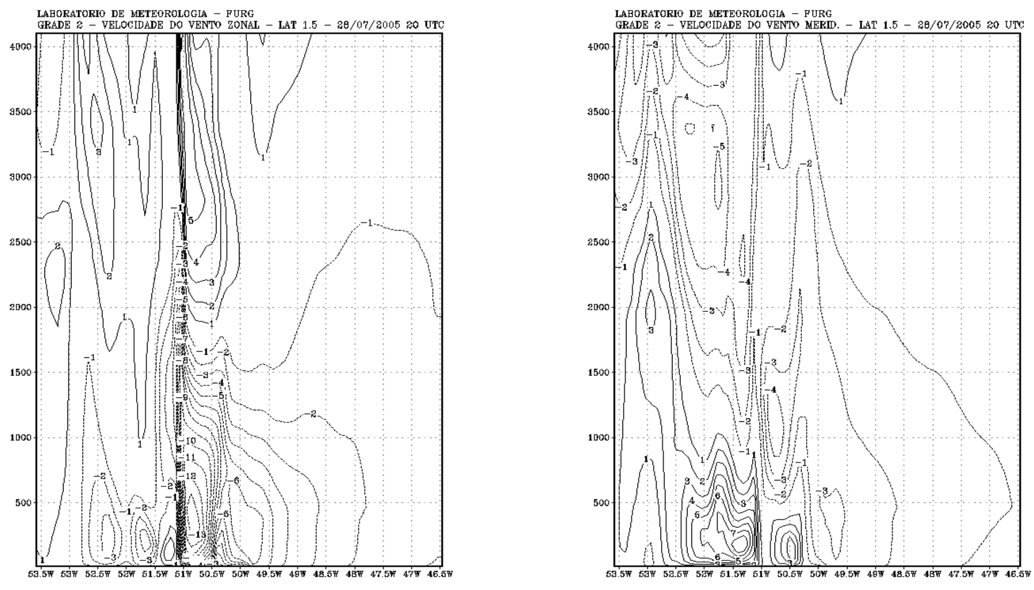
As seen previously, the greater breeze intensity was observed at 20 UTC at July 28. Figure 9 shows the penetration of the sea breeze for this same time. Ferreira and Cohen (2001) using sounding data for the region of Alcântara base in the state of Maranhão, observed that the greater wind intensity occurred near the 20 UTC, suggesting that this time is of greater sea breeze intensity in the region.

In this time is also visible the breeze front formation, that according to Pearson (1973), is the place where we have the maximum temperature variation and a clear change in the wind intensity. It can be observed in figure 10, that shows the temperature and u component intensity fields for a period of 15h in the centered point in 1.5°N and 50.5°W, starting at 12 UTC at July 28 till the 2 UTC at July 29. As we can observe in the figure, we had a significant reduction in the temperature and a great increase in the wind intensity at 20 UTC at July 28, with speeds near to 7ms^{-1} at 10 meters high. Pearson also shows that high values of vertical wind component are encountered in the area in the breeze front. It can be observed in the vertical section made to vertical wind component at 20UTC, where we can encounter values of 1.2ms^{-1} at approximately 2 km high (figure 11). The rising movement of the air generated by the breeze circulation is one of the main factors that influences in the formation of the nebulosity near to the littoral, as was observed in the satellite image at 20UTC (figure 12)

The rising movement, which contributes to the nebulosity formation, has its main cause in the convergence in low levels. The contrasts between the water and the land and also in the vegetation of the region generate this convergence. It can be observed in the vertical section made for the wind vector in the 0.3°S latitude at 20 UTC at July 28 (figure 13). Hong et al. (1995), using a bi-dimensional model, showed that the vegetation has a fundamental role in the generation of mesoscale circulation. But not only these previously described contrasts are responsible by the convergence in the surface, but also the topography has a great influence. The same figure shows that in approximately 52°W we have the convergence near the surface. It is directly related with the rising in the topography at about 300 meters high for this point (figure 14). This rising will generate another kind of local circulation, which was not described yet, called valley-mountain breeze. The breeze created by this rising will converge in surface together with the sea breeze, it can then generate a rising movement that analyzed previously.

We cannot forget the importance of the river in the generation and conservation of the breeze in the region because it occupies a considerable area in the study site. The figure 15, valid for the 16 UTC at July 28, shows a breeze generated by the temperature difference between the river and the continent. The breeze generated by the river also helps the increase of the humidity flow in direction to the continent. Since the difference between the water temperature of the river and the continent is lower than the ocean, the river breeze can delay more to develop, having its greater intensity delayed in relation to the sea breeze. In this study case this delay was of only one hour in relation to the beginning of the sea breeze, that occurred at 15 UTC at July 28.

During the night, the continent gets colder faster than the ocean, creating then a gradient with contrary direction in relation to the gradient during the day. With this situation, we have a diminution in the breeze intensity, and sometimes it can change its direction. In the final hours of the simulation, we already observed that the breeze intensity starts to diminish, and in the end of the simulation, at 12 UTC, July 29 (figure 16), we only have a diminution in the wind intensity on the continent. The same occurs with the river breeze.



(a) **(b)**

Figure 8: (a) A vertical section made on longitude 1.5°N in the u component, valid at the 20 UTC at July 28, 2005 (grid 2) (b) A vertical section made on latitude 1.5N° in the v component, valid at the 20 UTC at July 28, 2005 (grid 2)

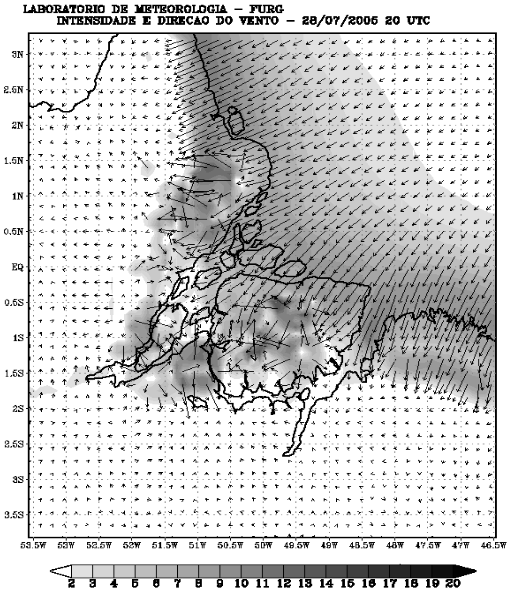


Figure 9: Prognostic for the direction and wind intensity fields at 10 meters at the 20 UTC at July 28, 2005 (grid 2)

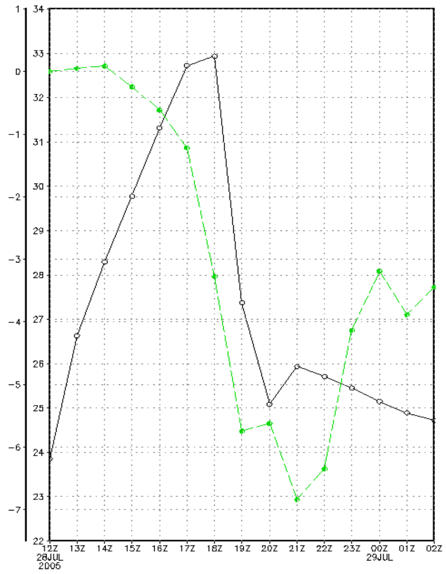


Figure 10: Temperature fields, continue line, and intensity of the u component of the wind, traced line, during the time in the centered point in 1.5°N and 50.5°W.

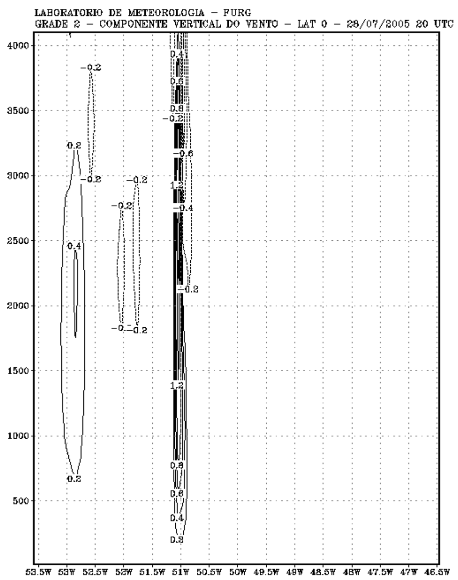


Figure 11: Vertical section made over the latitude 0° in the wind vertical component, valid for the 20 UTC in July 28, 2005 (grid 2)

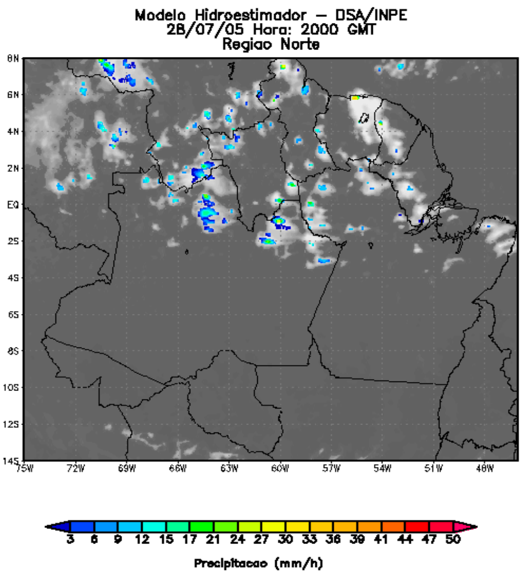


Figure 12: Satellite image of the 20 UTC, at July 28, 2005.

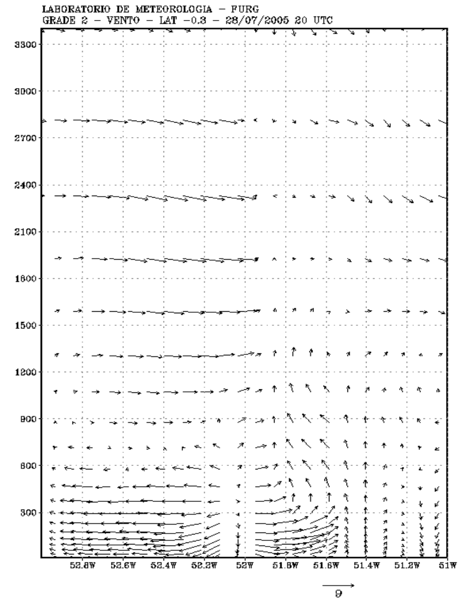


Figure 13: vertical section made over the latitude 0.3°S in the wind vector, valid for the 20 UTC in July 28, 2005 (grid 2)

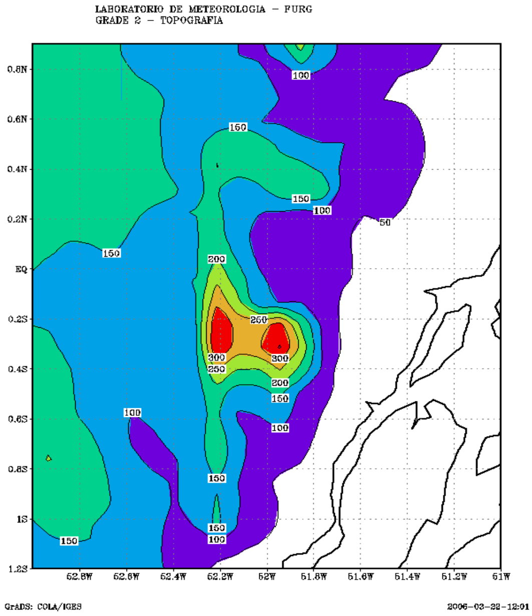


Figure 14: topography field (grid 2).

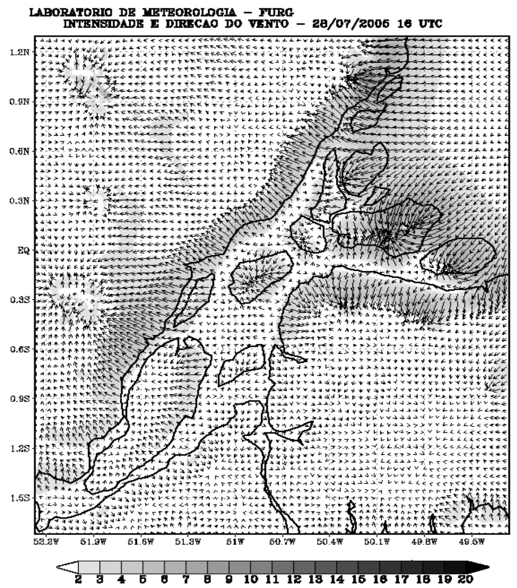


Figure 15: A prognostic of direction and wind intensity fields at 10 meters for the 16 UTC in July 28, 2005 (grid 3)

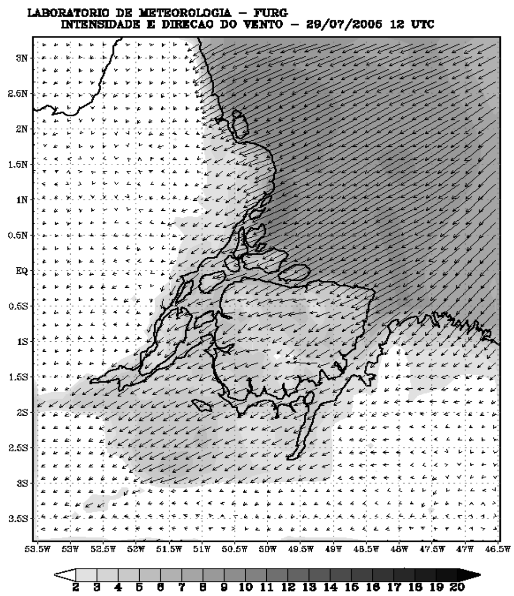


Figure 16: A prognostic of direction and wind intensity fields at 10 meters for the 12 UTC in July 29, 2005 (grid 2)

4. CONCLUSIONS

According the analysis of the generated figures from the realized simulation for 24 h starting at 12 UTC in July 28, 2005, we can conclude that we had a formation of sea breeze and of the river breeze on the study region. The breeze formed itself mainly due to the thermic contrast that exists between the continent and the ocean and in the case of the generation of the river breeze, the contrast existed between the river and the continent.

The sea breeze started at 15 UTC, at local 11 o'clock at July 28, and the river breeze delayed one hour in relation to the sea breeze start, starting at 16 UTC, at local 12 o'clock. The delay in the river breeze can occur due to the thermic difference between the river and the continent water be lower than the difference between the ocean and the continent, existing then a delay in the intensification of the temperature gradient that is enough to generate the breeze.

The greater intensity of the sea breeze was observed at 20 UTC, at July 28, with the intensity of $12ms^{-1}$ to the u component and of approximately $4ms^{-1}$ to the v component. It can also be characterized the return current of the circulation formed on the region. This current occurred near 3 km high to the u component and 4.5 km to the v component. The breeze front was also observed at 20 UTC at July 28, when were analyzed the temperature and intensity of the wind u component parameters.

The kind of the vegetation of the study area also has a great influence on the generation and maintenance of the breeze. As we have a forest region, it has great humidity, what delays the soil and the environment heating on the continent, causing then a delay in the formation of the breeze circulation.

The nebulosity formed during the simulation, which generated an accumulation of precipitation of 65mm at July 28, in the city of Macapá, had its main cause in the rising movement caused by the convergence in low levels, that left an instable atmosphere and consequently rose the humid air part that was near the surface to higher levels and then condensed it.

The preliminary results presented in this work show the importance of Amapá estuary in the generation of local atmospheric circulation in the region.

5. REFERENCES

Atkinson, B. W., 1981: Mesoescale Atmospheric Circulation. Pp 125-209.

Cohen, J. C. P., 1989: Um Estudo Observacional de Linhas de Instabilidade na Amazônia. Dissertação de Mestrado em Meteorologia. INPE - 4865 - TDL/376, São José dos Campos, São Paulo.

Espinoza, E. S., 1996: Ondas de Leste no Atlantico Tropical. In: VIII Congreso Brasileiro de Meteorologia - Climanalise Especial 10 Anos., Campos de Jordao.

Ferreira, S. A., Cohen, J. C. P., 2001: Estrutura Vertical da Atmosfera Associada às Circulações de Mesoescala Durante a Operação Norte III do Programa REVIZEE. Revista Virtual de Iniciação Científica da UFPA, vol 1, no 2., Belém, Brasil.

Frizzola, J. ª, Fisher, E. L., 1963: A Series of Sea Breeze Observations in the New York City Area. Journal of Applied Meteorology, vol. 2, pp 722-739.

Hong, X., Leach, M. J., Raman, S., 1995: Role of Vegetatin in Generation of Mesoscale Circulation. Atmospheric Environment, Vol 29, No 16, pp 2163-2176.

Kousky, V. E., Dias, M., 1982: Meteorologia Sinótica: Parte I. INPE, 118 pp.

Pearson, R. A., 1973: Properties of the Sea Breeze Front as Shown by a Numerical Model. Journal of the Atmospheric Sciences: Vol. 30, No. 6, pp. 1050–1060.

Pielke, R. A., 1984: Mesoscale Meteorological Modeling. pp 455-471.

Riehl, H., 1954: Waves in the easterlies and the polar front in the Tropics. Chicago Univ., Dept. of Meteorology, 79p(misc. rept. 17)

Saldanha, R. L., 2003: Estudo das Circulações Atmosféricas Locais Sobre o Estuário da Lagoa dos Patos. Dissertação de Mestrado em Oceanografia Física, Química e Geológica. FURG, Rio Grande, RS.

Saraiva, J. M. B., 1996: Previsão do Tempo na Região Sul: Efeitos Locais e Influência da Liberação de Calor Latente. Tese de Doutorado. IAG - USP.