13B.6 FACTORS CONTROLLING THE EAST PACIFIC ITCZ DURING EPIC2001

David J. Raymond,* Christopher S. Bretherton, and Graciela Raga
Universidad Nacional Autónoma de México

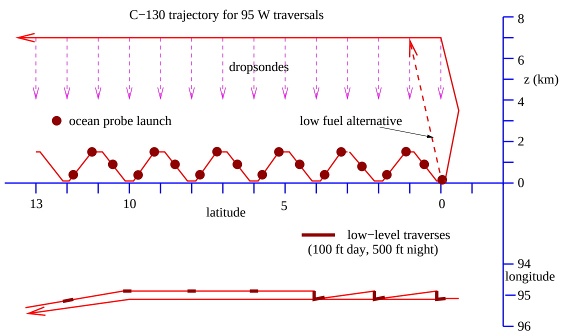

Figure 1: Pattern executed by the NCAR C-130 during the EPIC2001 project

1 INTRODUCTION

The strong, cross-equatorial sea surface temperature gradient in the eastern Pacific causes a seasonal-mean southerly flow across the equator at low levels which feeds the east Pacific ITCZ, typically located near 8° N during the northern summer and fall. The EPIC2001 project investigated this flow and the resulting ITCZ convection between 1 Sept and 10 Oct 2001. Observations were concentrated along 95° W, the longitude of the eastern-most TAO mooring line.

2 MEASUREMENTS

Figure 1 shows the pattern executed by the NCAR C-130 aircraft on 8 occasions in the above period. The porpoising pattern executed between 12° N and the equator allows conditions near the surface and 1600 m to be inferred with reasonable spatial continuity. The return traverse yields in situ measurements near 6300 m.

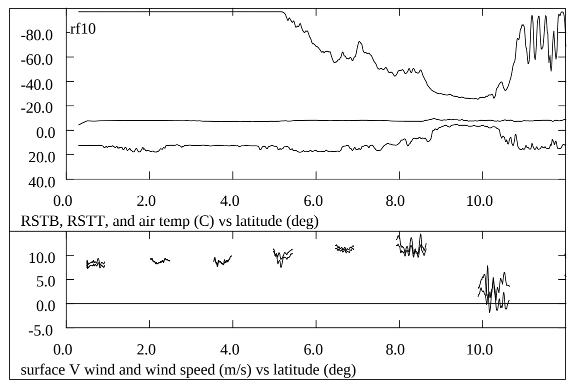

Figure 2: Flight 10 (23 Sept 2001) for the C-130. The upper panel shows the in situ temperature and upward and downward looking radiometers, which indicate the presence of cloud above and below the aircraft on its return flight near 6300 m. The lower panel shows the southerly wind component and the total wind near the surface on the outbound leg.

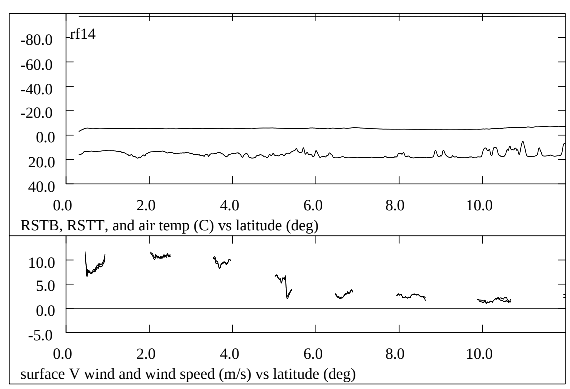

Figure 3: As in figure 2 except for flight 14 (2 Oct 2001).

In addition dropsondes were dropped every degree of latitude on the return traverse.

The cross-equatorial flow and the convection

*Corresponding author address: David J. Raymond, Centro de Ciencias de la Atmósfera, UNAM, México (on leave from Physics Department, New Mexico Tech, Socorro, NM, USA); e-mail: raymond@kestrel.nmt.edu.

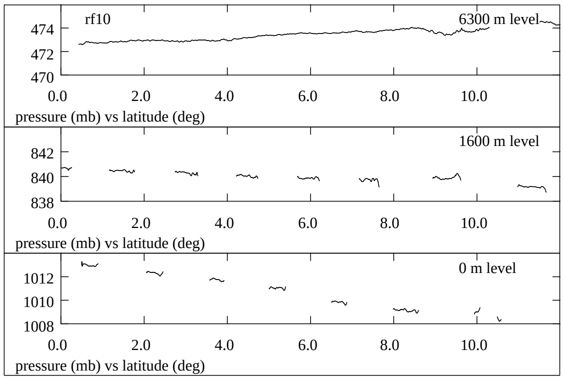

Figure 4: Pressure on three constant-level surfaces observed during C-130 flight 10, corrected for the semi-diurnal oscillation.

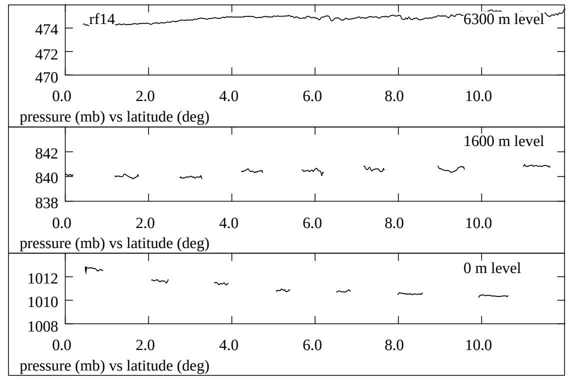

Figure 5: As in figure 4, except flight 14.

north of the equator are not steady features, but experience significant fluctuations on daily to weekly time scales. Figures 2 and 3 show the extreme limits of these fluctuations observed during EPIC2001, with the first figure corresponding to a case with strong convection and a correspondingly strong southerly flow. The second shows a case with weak flow and and no significant deep convection.

Using the radar altimeter on the C-130, the static pressure, and the hydrostatic equation, we were able to estimate the pressure on the fixed levels 0, 1600, and 6300 m. These levels are near the flight levels flown by the C-130, so the hydrostatic extrapolation of the pressure to fixed levels was small. There is a significant semi-diurnal pressure oscillation in the tropics, which gets confounded with spatial variations in pressure. This cycle was taken out of the C-130 data using the pattern of the oscillation observed by the TAO mooring pressure sensors.

Figures 4 and 5 show the corrected pressures for the two flights corresponding to figures 2 and 3. In both cases the difference between the surface and 1600 m pressures decreases with distance north of the equator. This result is expected from the hydrostatic equation, as the air in the marine layer becomes significantly warmer as it moves to the north over warmer waters.

South of 6° N, the pressure distributions look quite similar on the two days, with a negative pressure gradient at the surface. This pressure gradient drives the meridional flow at low levels against surface friction. However, north of 6° N the surface pressure gradient is much stronger in the flight 10 case than it is for flight 14. This correlates with the much stronger surface winds and deep convection north of 6° N for flight 10.

3 CONCLUSIONS

The above results show that the surface pressure gradient, and hence the surface wind, are not controlled completely by hydrostatic effects in the marine layer, which is largely confined to below 1600 m in this region. In particular, the surface pressure gradient north of 6° N in figure 4 results from an additive combination of hydrostatic effects in the marine layer and a negative north-south pressure gradient imposed from above. In figure 5, the pressure gradient aloft opposes that produced by the marine layer, resulting in a much weaker surface gradient. To the extent that convection is forced by surface heat fluxes, which are stronger when the wind is stronger, the factors contributing to variations in surface pressure gradients over the warm water north of 6° N will also contribute to variations in convection there.

Acknowledgments. We thank UCAR/JOSS, NCAR/RAF, NOAA/AOC, NOAA/PMEL, and UNAM for their excellent work on EPIC2001. EPIC is a part of the US CLIVAR program, with funding from NSF, NOAA, and CONACyT.