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INTERNAL GRAVITY WAVES IN THE EQUATORIAL PACIFIC

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1. INTRODUCTION

Mixing in the ocean surface layer is an important process in the transport of heat, momentum, and CO_2 into the deep ocean. For example, the flux of heat into the cold, upwelling water in equatorial regions provides one of the major heat sources driving the ocean circulation. Direct measurements of the ocean mixed layer have provided good estimates of the bulk layer properties. However, estimates of the small-scale effects of internal waves and related turbulence have remained ambiguous because of the difficulty in observing these processes (Moum et al. 1992; Wijesekera and Dillon 1991). Until more detailed measurements become available, numerical models can provide a convenient and cost-effective way to analyze the details of the surface mixed layer. .

Modeling the surface layer of the equatorial Pacific Ocean is challenging because of the strong vertical current shear and density stratification common to the region (Figure 1). The primary zonal current is the eastward flowing Equatorial Undercurrent (EUC) centered at roughly 120 m depth, with a speed of about 1.5 ms^{-1} as shown in Figure 1. The EUC is forced by a zonal pressure gradient resulting from the westward directed surface wind stress (McCreary 1981). Above the EUC, the wind stress directly forces the South Equatorial Current (SEC), which flows westward with a speed of about 0.5 ms^{-1}. The shear zone generated by these currents is marginally stable and exhibits a diurnal cycle of turbulence dependent on convection forced by surface cooling (e.g., Moum et al. 1989). In addition, surface convection forces internal gravity waves, which can transport momentum away from the surface current to deeper waters (Wijesekera and Dillon 1991). In this report, we discuss recent modeling results for the equatorial Pacific showing the generation of convection, turbulence, and internal waves.

2. RESULTS

The model is based on the two-dimensional, nonhydrostatic, Boussinesq equations of motion with an incompressible flow assumption. Subgrid scale processes are parameterized using a version of the standard Smagorinsky (1963) subgrid-scale model. A full equation of state for sea water is used; however, in the work described here salinity was held constant. The domain size for the simulation was 1280 x 200 m with 5 m resolution. Initialization of temperature and zonal velocity from the surface to 150-m depth was based on vertical profiles reported in Hebert et al. (1991) for a location on the equator near 135° W. A linear extrapolation was assumed for both temperature and velocity below 150 m. A constant surface wind stress of 0.05 Nm^{-2} was prescribed along with an idealized surface heat flux representing typical conditions at 135° W. A 3-day simulation period was chosen, allowing the model to adjust during the first day and still provide two full heating cycles for analysis.

Plots of the model vertical velocity and temperature structure are shown for times representing surface heating (Figure 2) and surface cooling (Figure 3) from simulation day 2 and 3. The model reveals that the surface mixed layer and thermocline undergo a diurnal cycle with internal waves active throughout the simulation. The character of these waves is strongly affected by the surface heating cycle. During the day, the waves outside the thermocline region are almost random in scale with wavelengths between 50 and 200 m and phase speeds ranging from near zero to about 0.1 ms^{-1} (Figure 2). At night, the wave field becomes more coherent with a dominant wave-

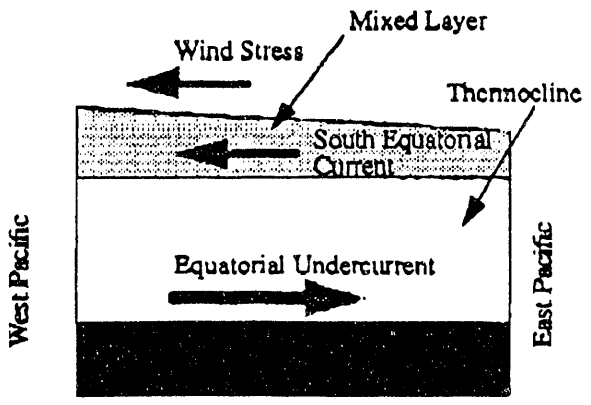

Figure 1. Schematic of the equatorial current structure.

Figure 2. Temperature and vertical velocity after 2 simulation days. Contour interval for vertical velocity is 0.02 m^{-1} and for temperature is 2° C.

Figure 3. Same as Figure 2 for simulation day 2.5 and with a vertical velocity contour interval of 0.04 ms^{-1}.

(Figure 3). Both the wavelength and strength of the nocturnal waves are similar to a wave event observed using a towed thermistor chain as reported in Hebert et al. (1992). Also, the dominant wavelength of about 200 m agrees with the time series analysis presented in Moum et al. (1992) for the 4-day period beginning on April 14, 1987.

A number of conceptual models have been proposed to explain the formation of internal waves in convective, sheared flow. The most popular hypothesis treats convective plumes as independent structures that act as obstacles to the mean flow, generating trapped internal waves. In the numerical model, this process is active only during the initial spin-up phase of the simulation. As soon as internal waves are generated, the convection becomes a forced internal mode, which weakens during the daytime stabilization period, but does not completely dissipate.

The key to the internal mode forcing is the presence of critical layers at 35 and 150 m depth, where the mean velocity is about 0.1 ms^{-1}, matching the dominant internal wave phase speed. During the cooling period, the Richardson number in the upper critical layer drops below 0.25, creating an unstable gravity wave mode with energy being transferred from the mean flow to the internal wave. When surface heating is imposed, the internal waves decrease in strength and act as a drag to the mean flow near the critical layers (Figure 4). The magnitude of the wave drag is comparable to the turbulent dissipation rate and the large-scale zonal pressure gradient estimated by Hebert et al. (1991), implying that the internal waves may play a significant part in the zonal momentum budget. The critical layer instability also promotes the generation of small scale waves with wavelengths between 30-60 m that are similar in appearance to Kelvin-Helmholtz billows. These disturbances may represent the beginnings of hydrodynamic instability, explaining the diurnal cycle in turbulent dissipation observed in the upper portion of the thermocline. Also, measurements of turbulent dissipation taken beneath the EUC by Peters et al. (1988) support the existence of a critical region near 150 m depth.

Figure 4. Average rate of change for horizontal velocity during day 3 of the simulation.

3. CONCLUSIONS

In conclusion, numerical simulations have shown that internal waves are produced by critical layer interaction in the equatorial surface layer and can transport eastward momentum downward by producing a wave drag at two critical layers. The strength of the drag is of the same order as other terms in the zonal current budget equation.

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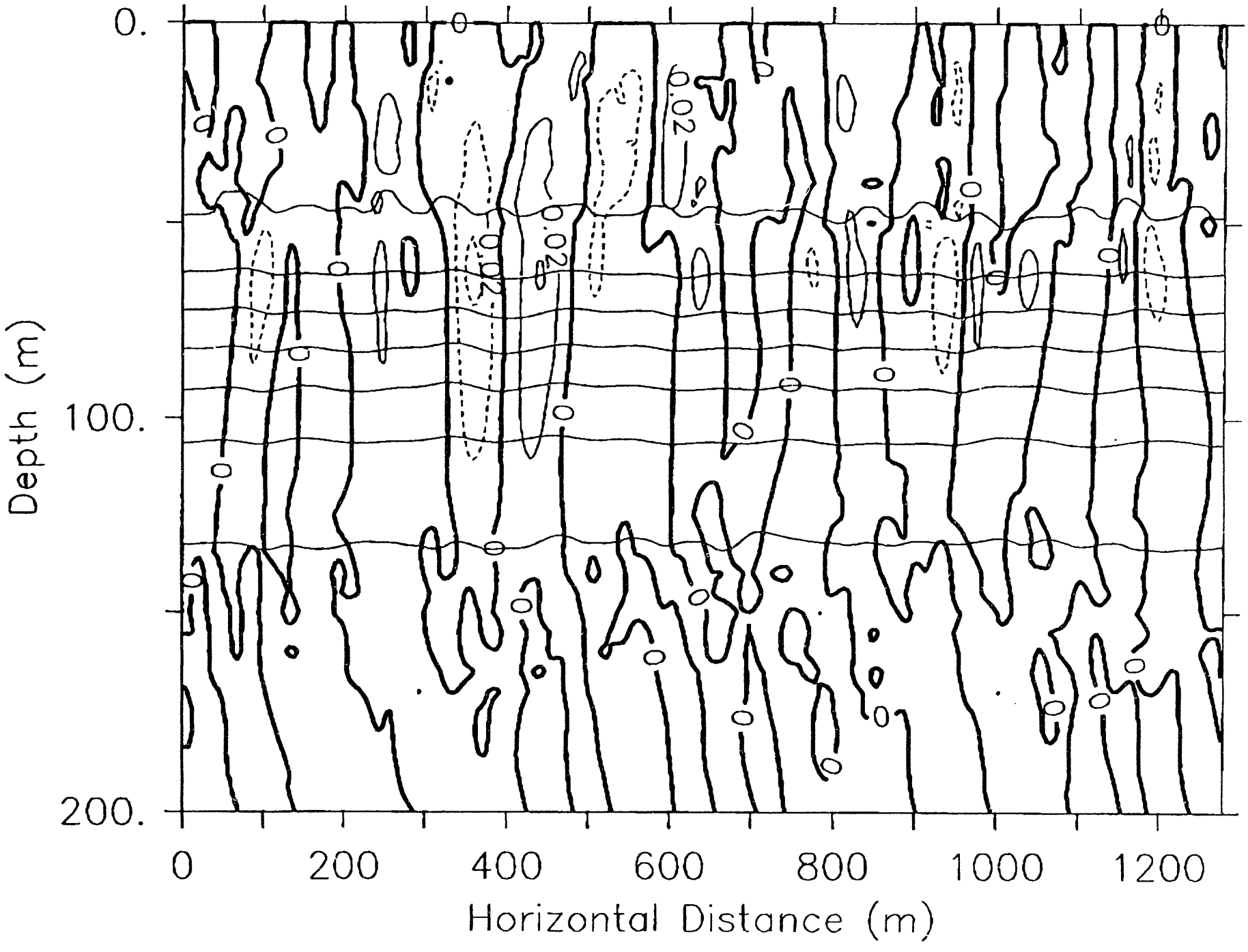
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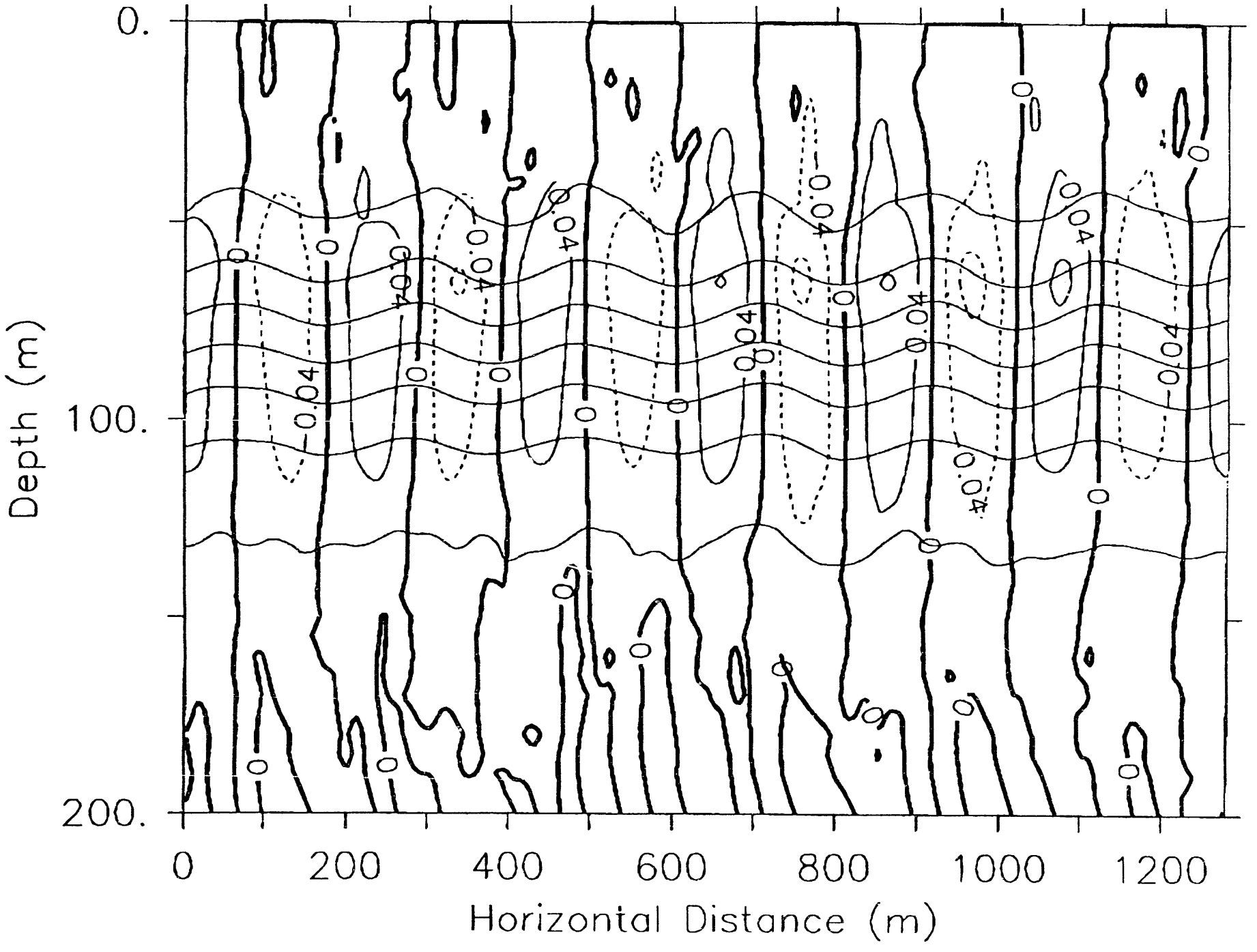
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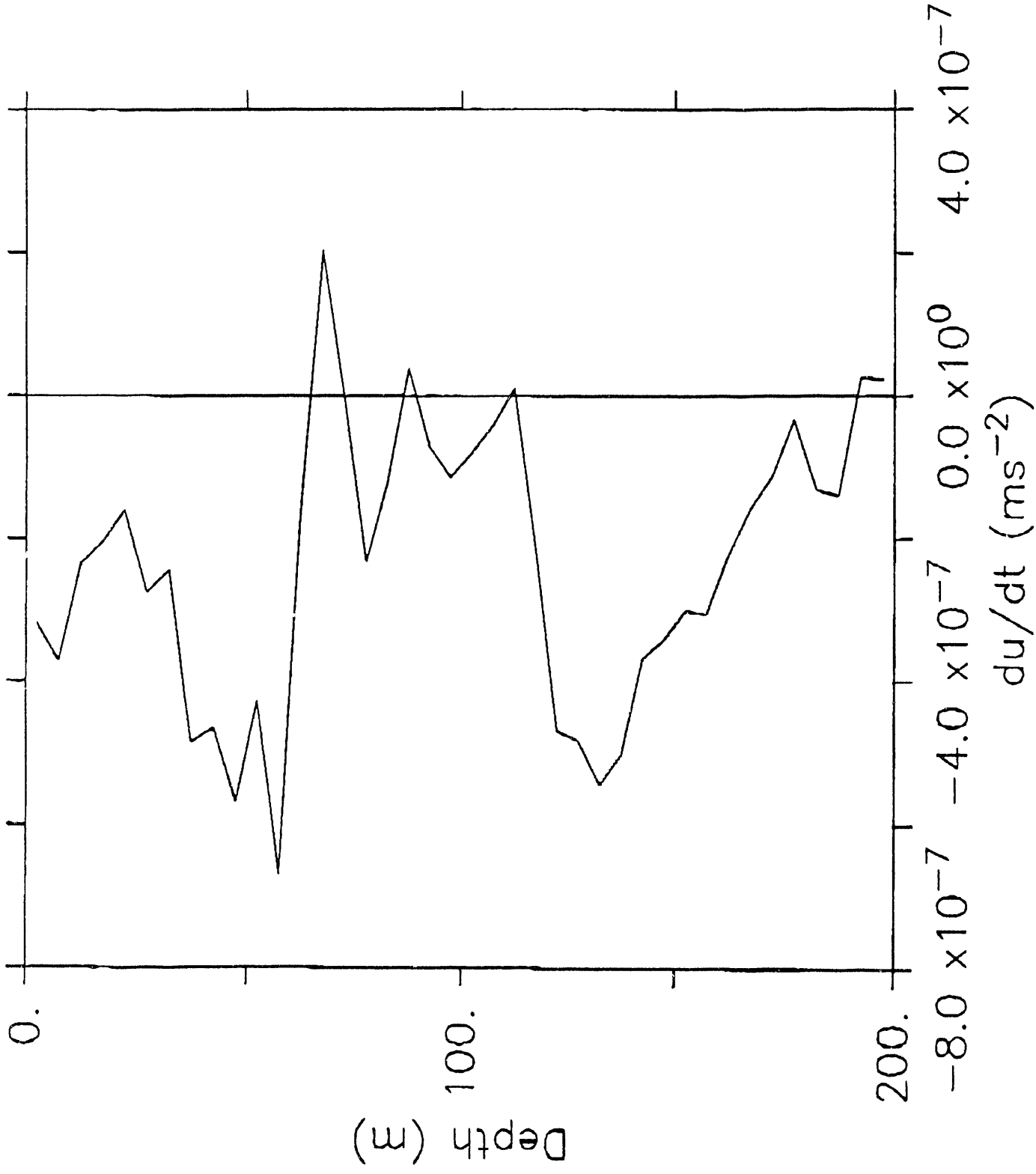
Fig 4.



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