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How Sluggish Are the Waters of the Dead Sea?

A. Huss[1], T. Ezer[2] and A. Hecht[2]

[1] *Department of Atmospheric Sciences, The Hebrew University of Jerusalem, Jerusalem 91904, Israel and*

[2] *Department of Physical Oceanography, Israel Oceanographic and Limnological Research, National Institute of Oceanography, Haifa, Israel.*

Abstract

In the course of a series of numerical simulations of the circulation in the Dead Sea, it was found that wind forcing generates currents which are about 15% less intense in waters as dense as the water of the Dead Sea than those generated by identical forcing in the Mediterranean Sea.

Introduction

The density of the Dead Sea water is very high higher by about 20% than that of the Mediterranean water, which is also rather high in comparison with that of the water of the open oceans. This is due to the extremely high salinity of the Dead Sea, about 280‰ as compared to that of the surface layers of the open oceans, about 35‰. While temporal and spatial variations in the density of the surface layers of the oceans are relatively small, they could be considerably larger in the Dead Sea, due to the large daily and seasonal temperature variations and to the influx of flood waters during the winter.

Due to the peculiar chemicals dissolved in the Dead Sea waters and the relative proportions between these chemicals (e.g. Neev and Emery, 1967), the usual definition of salinity is meaningless and the density of its water cannot be computed according to the internationally accepted procedures (e.g. Millero et al., 1980; Millero and Poisson, 1981). A tentative empirical equation of state was proposed by Steinhorn (1980).

$$\rho_T = \rho_S + (S - S_s)\frac{\partial \rho}{\partial S} + (T - 25)\frac{\partial \rho}{\partial T} \qquad (1)$$

Where ρ_T denotes at the temperature T (in °C) and salinity S, and $\rho_S = 1.23179$ gr cm^{-3} is the density at 25°C, and at a salinity of 275.8‰;

$\frac{\partial \rho}{\partial s} = 0.00092$ kg cm^{-3}; $\frac{\partial \rho}{\partial T} = -0.00042$ g cm^{-3} deg^{-1}.

A legitimate question arises: To what extent does the larger density influence the dynamics of the wind-driven currents? An appeal to observations would encounter the difficulty of identifying similar forcing, which would permit comparison between different water bodies. Moreover, the unique morphology of the Dead Sea would be an additional handicap. On the other hand, computational simulation may at least provide a tentative answer.

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Ezer (1984) has made a series of simulations of currents in the Dead Sea in which he employed a model described by Huss and Feliks (1981) (involving both atmosphere and sea) with minor modifications, i.e. reducing considerably the number of atmospheric levels. The "atmospheric" section of the model computed the wind field at 3 levels from a height of 5 m above the sea surface, where the winds were measured, to the surface of the sea. The governing equations consist of two horizontal components of the equations of motion, to the hydrostatic equation, the thermodynamic energy equation and the continuity equation, with appropriate parameters for the two different media. For the adoption of turbulent diffusion coefficients, K-theory was applied. The "air-sea" interaction section of the model computed the currents at 13 levels below the sea surface. The sea was assumed to be homogenous in the experiments described here down to the bottom, as the spatial and temporal variation of density (for the length of the run) would only have secondary effects. Thus the vertical diffusion term in this model is independent of the density.The density also cancels out in the horizontal gradient force. Therefore the different densities in the different runs will show up only in the sea-atmosphere interaction at the interface. This has been formulated by assuming a contiunuous flux of momentum through the material interface, as described by Pandolfo and Jacobs (1973):

$$\rho_w K_{mw}\left(\frac{\partial V_h}{\partial z}\right) = \rho_a K_{ma}\left(\frac{\partial V_h}{\partial z}\right) \tag{2}$$

with ρ – the density; K_m – the turbulent diffusion coefficient; V_h – the horizontal velocity; z – the vertical coordinate; and the subscripts "w" and "a" referring to water and air, respectively. In Eqn. (2):

$$K_{mw} = l_w\left(\frac{\partial V_h}{\partial z}\right); K_m = l_a\left(\frac{\partial V_h}{\partial z}\right) \tag{3}$$

where the mixing lengths l were computed according to Lettau (1962):

$$l = k\,(2 + z_o)\left[1 + 4\left(\frac{z+z_o}{z_m}\right)^{5/4}\right]^{-1} \tag{4}$$

where K = 0.4 is the Von Karman constant and with appropriate values for z and z in the sea and the atmosphere, involving also an empirical relation which takes into account the influence of the surface waves, as derived by Pierson (1964), and which was found to be a reasonable approximation in the Dead Sea.

z_m is the value of $z + z_o$ at which l has its maximum value. z_m was taken as 250 m in the atmosphere, according to Lettau (1962), and equal to 20 m in the sea, based on trial simulations.

The z_o of the atmosphere was taken to be 1 mm. For the sea in the upper half of the total depth it was taken to be H/2, where H is the mean wave-height, which, according to Pierson (1964) is given by

$$H = 1.5416 \times 10^{-4}\,(u^2 + v^2)\ \text{m} \tag{5}$$

where u and v are the velocity components at a depth of 20 m. In the lower half of the sea z_o is taken as 10 cm. Further details are given in Ezer (1984).

When Eqn. (2) is written in finite-difference form and the z-increments just above and just below the interface are equal, $(\Delta z)_a = (\Delta z)_w = 125$ cm in our case, we obtain, with the initial currents equal to zero, after the first time step, the current velocity at the interface:

$$V_o = V_a\left[1 + \frac{l_w}{l_a}\left(\frac{\rho_w}{\rho_a}\right)^{1/2}\right]^{-1} \tag{6}$$

Under the model assumption:

$$\frac{l_w}{l_a} = 0.38 \tag{7}$$

for the mixing lengths adjacent to the interface. Thus we obtain for the Mediterranean and the Dead Sea, respectively:

$$\left(\frac{V_a}{V_o}\right)_{Med} = 13.0\ ;\ \left(\frac{V_a}{V_o}\right)_{D.S.} = 14.3 \tag{8}$$

The initial currents generated by the same wind are therefore perceptibly smaller in the Dead Sea. This estimate is, of course, only valid for the initial time and for the interface.

A better estimate can be obtained by comparing the output of the model for two runs with different densities. This has been carried out with a two dimensional version of the model mentioned above, i.e. applied to a vertical plane in the E-W direction. It was felt that for the problem posed the results from this model, not requiring the large amounts of computer-time as the 3-dimentional model, would be adequate.

The water density for the first run corresponded to a temperature of 27°C and a salinity of 40‰, i.e. "normal" eastern Mediterranean sea water. The water density for the second run corresponded to the same temperature, i.e. 27°C, but the salinity was assumed to be 280‰ and the density was computed from (1). In both cases, a constant southeasterly wind with a speed of 3.5 m s^{-1} was assumed to blow at 5 m above the sea surface. It should be mentioned that currents measured in the Dead Sea in corresponding environmental conditions (for details see Ezer, 1984) appear to indicate that the model produces reasonably realistic results.

After only a three hour run the curents in the Mediterranean simulation were about 13% stronger than those in the Dead Sea simulation. After a 12 hour run the difference increased to about 15% (Fig. 1).

Moreover, the model runs indicated that the difference between the speeds of the currents in the two environments increases with depth.

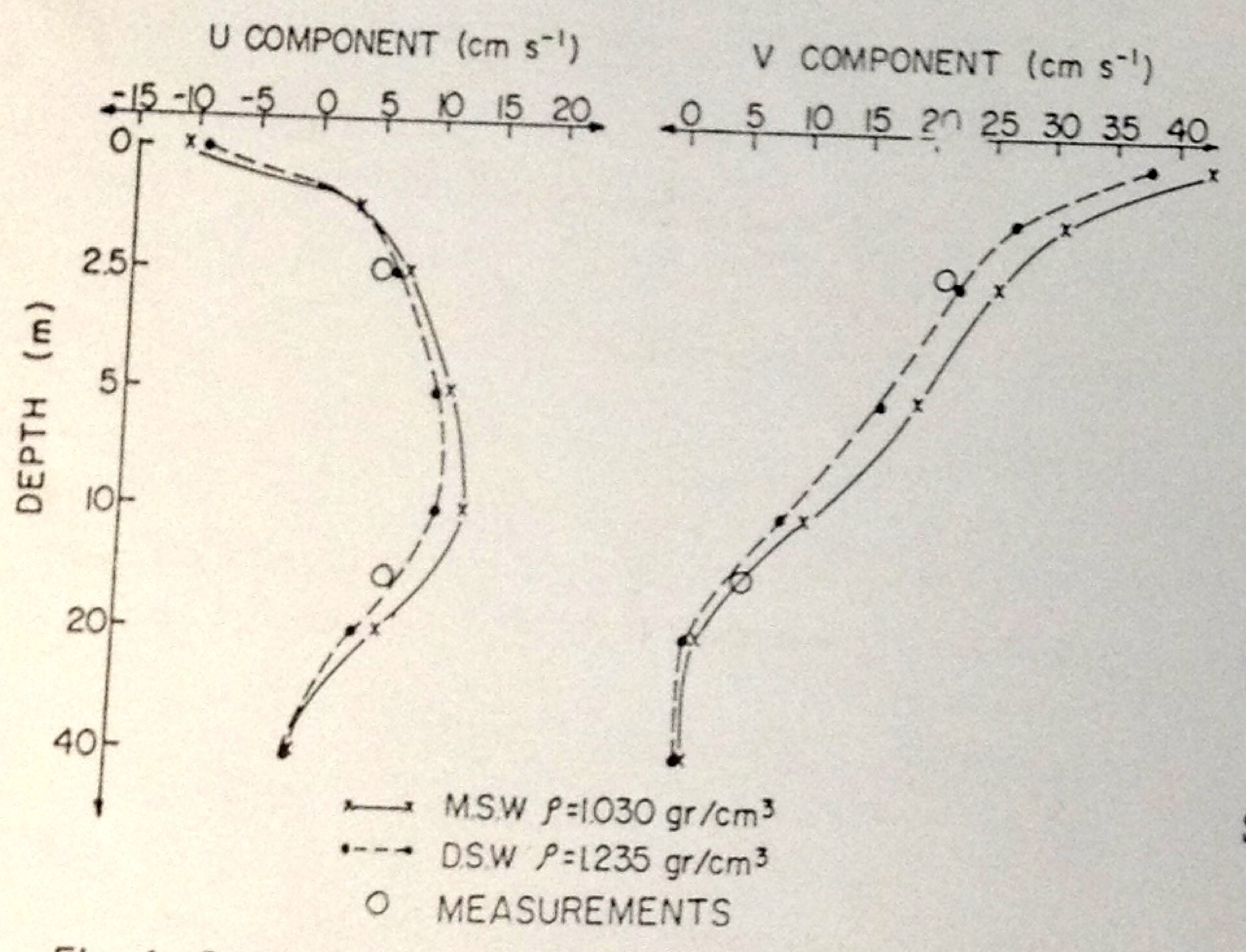

Fig. 1. Current profiles in the Mediterranean and the Dead Sea waters, forced by a constant wind at 5 m above sea surface ($u = -250$ cm s^{-1}, $v = 250$ cm s^{-1}) after 12 hours.

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