CHAPTER 94

COMPUTING EOLIAN SAND TRANSPORT FROM ROUTINE WEATHER DATA

S. A. Hsu
Coastal Studies Institute, Louisiana State University
Baton Rouge, Louisiana 70803

ABSTRACT

Extensive wind profile measurements have been made over beaches, tidal flats, and small dune fields in Barbados, Ecuador, Florida, Texas, and on Alaskan Arctic coasts by the author and his associates during the past several years. The linear relationship between shear and wind velocities was further verified by these measurements, in conjunction with Bagnold's data from experiments in a Libyan desert. From these measurements, a simple method is developed for computing eolian transport of the most common sand particle sizes which occur on coasts and deserts by using only routinely available wind observations from nearby weather stations.

INTRODUCTION

In the coastal environment one of the main unsolved problems of sand supply and loss is that of calculating the rate of sand transport by wind action. Recently, Hsu (1971a) introduced a method based on limited field measurements by which this rate can be scaled using a special Froude number. Extensive wind profile measurements have been made over beaches, tidal flats, and small dune fields in Ecuador, Florida, Texas, and on Alaskan Arctic coasts by the author and his associates during the past several years. The linear relationship between shear and wind velocities was further verified by these measurements and also by Bagnold's (1941) data from experiments in a Libyan desert (Hsu, 1973).

To add to the measurements previously obtained, an experiment to measure the wind stress on the air-beach interface of an island beach was conducted on Barbados, West Indies, during July-August 1973. The results are synthesized in this paper, along with those from similar experiments on continental coasts (Hsu, 1973). These results are related to routine wind observations from weather stations, so that anyone who is interested in computing the rate of eolian sand transport can use the simple formula provided in this paper for general application. The necessity for complicated measurements of atmospheric shear velocity is eliminated.

DEVELOPMENT OF THE METHOD

The relationship formulated by Hsu (1971a) for computing the rate of sand transported by the wind is (see Fig. 1)

$$q = K \cdot Fr^3 = K \left[\frac{U_*}{(gD)^{1/2}}\right]^3 \tag{1}$$

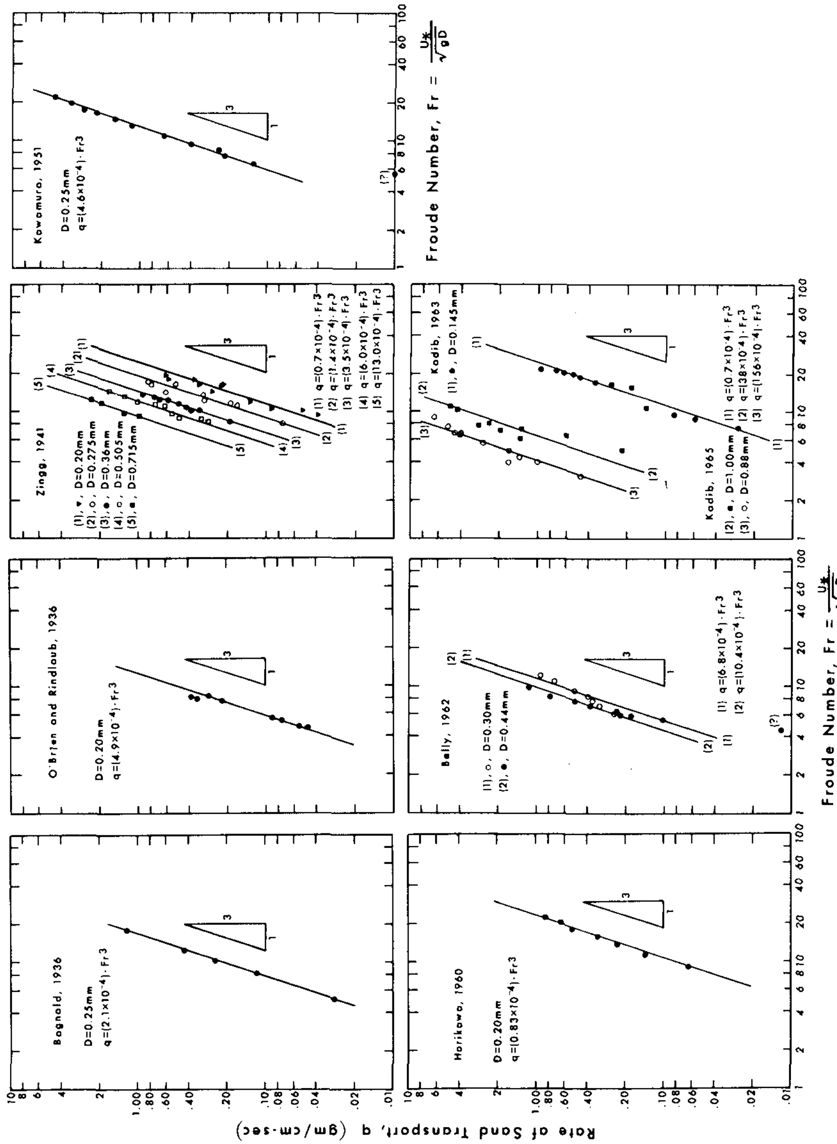

Figure 1. Froude number scaling in the rate of eolian sand transport (from Hsu, 1971a).

where q (gm/cm-sec) is the rate of sand transported by the wind and Fr is a special Froude number. Fr is a function of the atmospheric shear velocity U_* (cm/sec), the acceleration of gravity g (980 cm/sec^2), and the mean grain size of the sand particles D (mm). K is defined as the dimensional eolian sand transport coefficient and has the same dimensions as q. The values of K are delineated in Fig. 2. The value of Fr is explicitly determined by the bracketed term in equation (1).

If the value of U_* is known, the rate of eolian sand transport can be computed. In order to utilize routine wind observations to compute U_* values, the following procedure was developed.

In the lowest turbulent layer of the atmosphere over land, under homogeneous steady-state conditions, the logarithmic horizontal wind velocity increase with height has been observed over beaches (Hsu, 1971b), over tidal flats (Hsu, 1972), over deserts (Bagnold, 1941), and in laboratory channels (Kadib, 1965).

The logarithmic wind profile law states that

$$\bar{U}_z = \frac{U_*}{\kappa} \ln \frac{z}{z_o} , \qquad (2)$$

where $\bar{U}_z$ is the mean horizontal wind velocity at any given height z, U_* is the shear (or friction velocity, equivalent to $[\tau/\rho]^{1/2}$, where τ is the surface wind stress and ρ is the air density), κ is the von Kármán constant ($\approx$0.4), and z_o is the aerodynamic roughness length defined under the boundary condition that $\bar{U}_z = 0$ at $z = z_o$. The value of z_o depends upon the characteristics of the underlying surface.

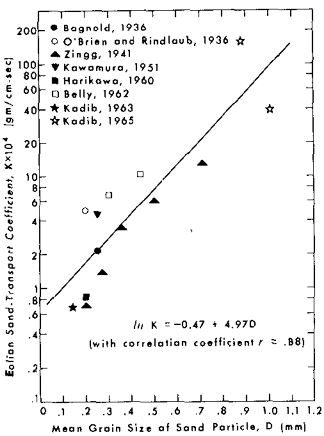

Figure 2. Determination of eolian sand transport coefficient from mean grain size of sand particle (from Hsu, 1971a)

Under eolian sand transport conditions, however, equation (2) should be modified such that

$$\bar{U}_z - U_t = \frac{U_*}{\kappa} \ln \frac{z}{z_{ot}} , \qquad (3)$$

where U_t is the threshold velocity (see Bagnold, 1941) and z_{ot} is the roughness length defined under the boundary condition that $\bar{U}_z = U_t$ at $z = z_{ot}$. The value of z_{ot} depends upon the mean grain size of the surface sand under question.

Figure 3 verifies the validity of equation (3) on coasts and in deserts by showing some examples of wind profile measurements under eolian sand transport conditions. For instrumentation and wind data reduction and analysis procedures on coasts and in deserts, see Hsu (1971b, 1972) and Bagnold (1941), respectively. Similar criteria have been used in investigations in laboratory channels (e.g., see Kadib, 1965).

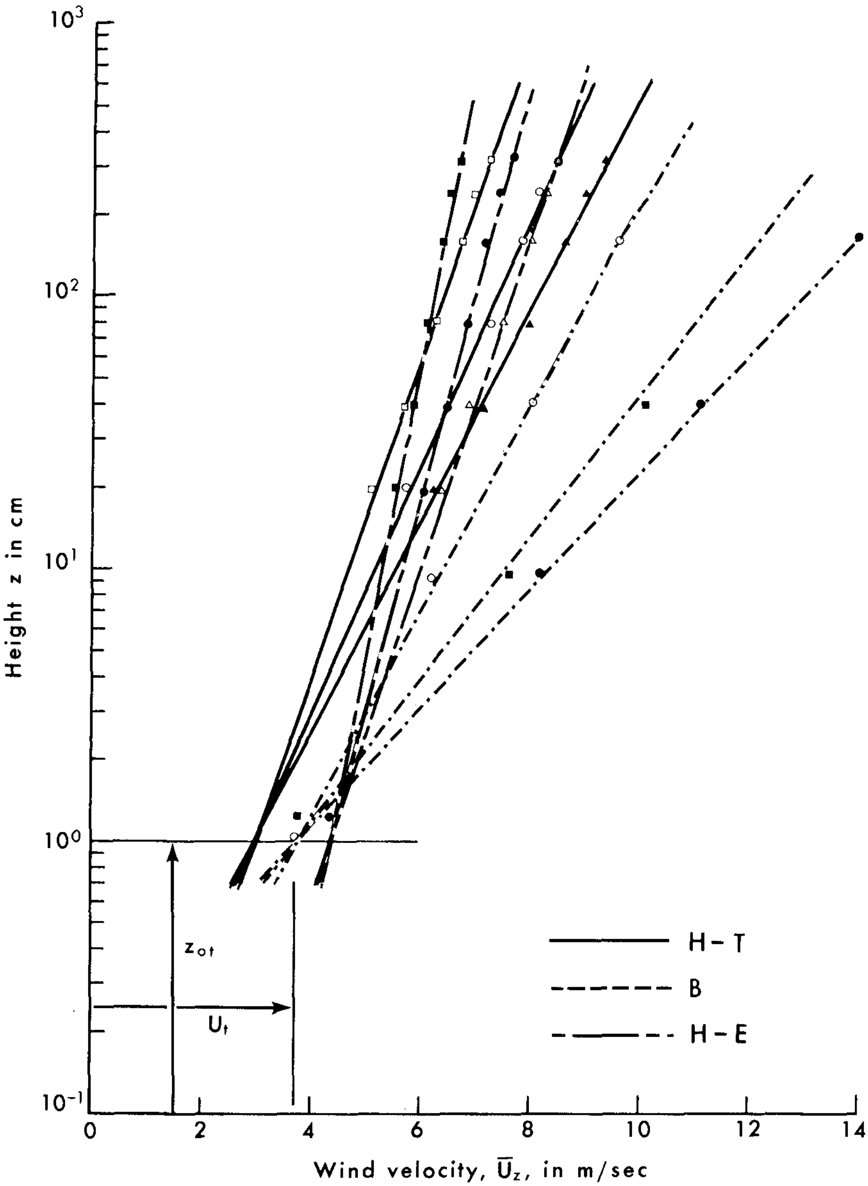

Figure 3. Examples of observed wind velocity vertical distribution over sand surfaces when the sand was in motion. The data lines are indicated by H-E, H-T, and B for measurements made by Hsu over a beach near Playas, Ecuador, by Hsu over a beach on northern Padre Island, Texas, and by Bagnold in the Libyan desert (from Hsu, 1973).

Figure 4 shows that the linear relationship between shear and wind velocities is further verified by wind profile measurements for eolian transport of sands with mean diameters ranging from 0.2 to 0.3 mm. This relationship was derived using data collected from a variety of environments, including those over beaches, tidal flats, and small dune fields in Barbados, Ecuador, Florida, and Texas by Hsu (1971a, 1972, and unpublished data), over an Arctic beach by Walters (1973), and in the Libyan desert by Bagnold (1941). Some detailed information was given in Hsu (1973, 1974).

Thus the rate of eolian sand transport can be computed by a given value of D and wind speed at 2 m above the surface by using Figs. 2 and 4. However, it is not always convenient for an anemometer to be located 2 m above the surface. In this case, equation (3) may be used. At heights z = 2 m and 10 m we obtain from equation (3) and Fig. 4 the following set of equations to calculate U_* as a function of U_{10m}:

$$U_{10m} - U_t = \frac{U_*}{\kappa} \ln \frac{1000}{z_{ot}} \tag{4a}$$

$$U_{2m} - U_t = \frac{U_*}{\kappa} \ln \frac{200}{z_{ot}} \tag{4b}$$

$$U_* = 0.044\ U_{2m} \tag{4c}$$

Equation (4) yields

$$U_* = 0.037\ U_{10m}\ . \tag{5}$$

Note that the difference between equations (4c) and (5) is rather small. Because an anemometer at a height between 2 and 10 m is usually available at nearby weather stations, for general applications the average of equations (4c) and (5) may be useful:

$$U_*\ (\text{cm/sec}) = 4.0\ U\ (\text{m/sec}) \tag{6}$$

where U is the known hourly averaged wind velocity in the constant direction of transport at an anemometer height between 2 and 10 m.

Therefore, by using Fig. 2 in conjunction with equation (6), the rate of eolian sand transport can easily be computed for a given value of D. An example is given in the following section.

APPLICATION OF THE METHOD

It is simple to apply the method developed in the previous section. For example, to compute the most commonly occurring and well-sorted standard

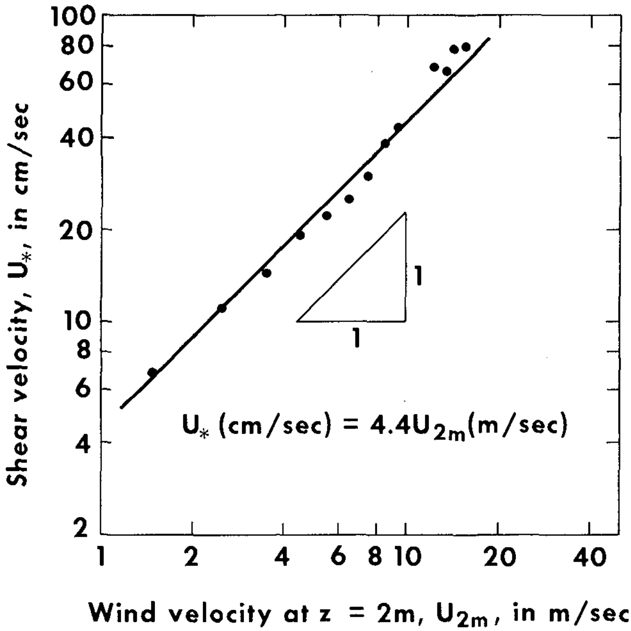

Figure 4. Relationship between shear and wind velocities synthesized from many measurements, ranging from tropical to arctic continental coasts (for details see Hsu, 1973 and 1974) and the Libyan desert, as well as a windward beach near Bath, on the island of Barbados, West Indies.

sand particle size D (= 0.25 mm; see, e.g., Bagnold, 1941), the following steps may be followed:

1. From Fig. 2 we have $K = 2.17 \times 10^{-4}$ by setting D = 0.25 mm.

2. Substituting values of K, g, D, and equation (6) into equation (1) we get

$$q = 1.16 \times 10^{-4} U_*^3 . \qquad (7)$$

Equation (7) is shown in Fig. 5.

The accuracy of this simple method should be within the experimental error arising from field and laboratory measurements because equation (7) is developed from the data sets. For those sand particle sizes not well sorted, Bagnold's correction criterion may be used. For nonstandard sizes,

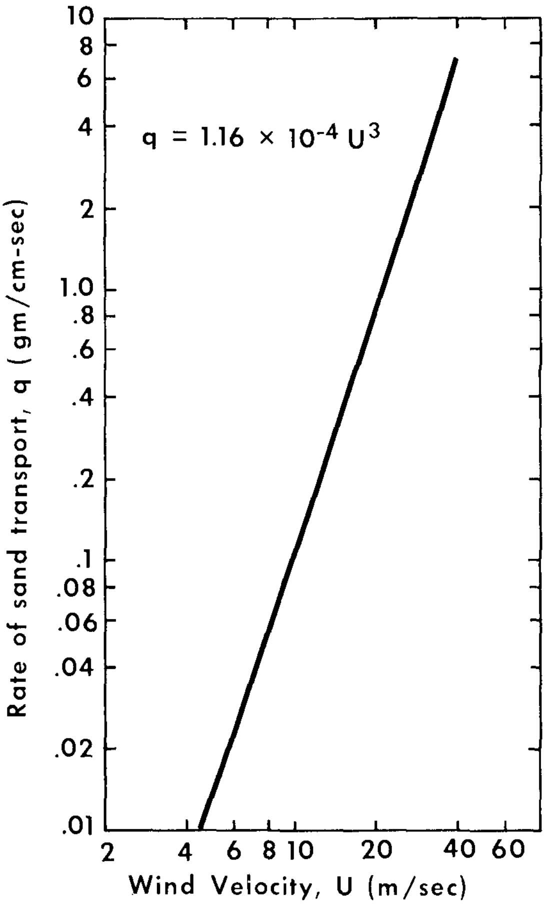

Figure 5. Computation of eolian sand transport for standard particle size (D = 0.25 mm) from routine wind observations at height 2 to 10 m above the ground obtained from nearby weather stations. U is the known hourly averaged wind velocity in the constant direction of transport.

it is recommended that Kadib's collection of U_t and z_{ot} be applied to the computation of U_*, which will be used in conjunction with K to calculate q. Further research should be conducted on the transport mechanism in the wet sand environment (e.g., swash zone) and sand movement around manmade coastal structures.

ACKNOWLEDGMENTS

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REFERENCES

Bagnold, R. A., 1941, The physics of blown sand and desert dunes. New York (Morrow), 265 pp.

Hsu, S. A., 1971a, Wind stress criteria in eolian sand transport. J. Geophys. Res., 76:8684-8686.

________, 1971b, Measurement of shear stress and roughness length on a beach. J. Geophys. Res., 76:2880-2885.

________, 1972, Boundary-layer trade-wind profile and stress on a tropical windward coast. Boundary-Layer Meteorol., 2:284-289.

________, 1973, Computing eolian sand transport from shear velocity measurements. J. Geol., 81:739-743.

________, 1974, Experimental results of the drag-coefficient estimation for air-coast interfaces. Boundary-Layer Meteorol., 6:505-507.

Kadib, A. A., 1965, A function of sand movement by wind. University of California, Berkeley, Tech. Rept. HEL-2-12, 91 pp.

Walters, C. D., 1973, Atmospheric surface boundary layer wind structure studies on an Alaskan Arctic Coast during the winter and summer seasons. Unpublished Master's thesis, Louisiana State University, 45 pp.