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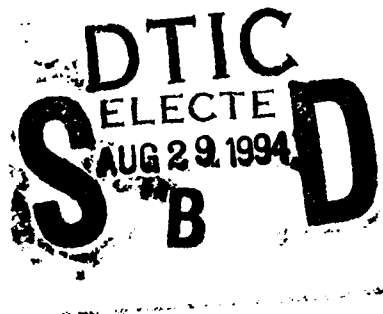


Determination of the Similarity Theory Scaling Parameters from the Vertical Gradients

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13. ABSTRACT (Maximum 200 words) Dynamic similarity of flows theory as applied to the surface boundary layer of the atmosphere must be treated as an interwoven quadrupartite function of trivariant orthogonal turbulent motion, the thermal structure and associated vertical gradients, evaporation at the interface, and vertical transport of water vapor with respect to the specific humidity as a function of height. Physical studies of the surface layer are typically conducted with respect to the buoyant heat flux and the momentum flux. Buoyancy is associated with the sensible and latent heat fluxes. The key parameters for establishing the vertical fluxes are the scaling velocity, scaling temperature, and scaling humidity. An interpretation of the similarity theory developed from the basics of the mean value theorem of calculus allows evaluation of the scaling parameters without utilizing the mathematically complex diabatic influence functions, but directly from the vertical gradients of wind, temperature, and specific humidity.					
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1. Introduction

Surface boundary layer processes, in particular the turbulent kinetic energy budget and the surface energy balance, are cohesively linked to the vertical profiles of wind, temperature, and specific humidity that characterize the mean flow and turbulent structure of approximately the first 100 m of the atmosphere. The shapes of the vertical profiles are deemed to be controlled by the nondimensional parameters of the dynamic similarity theory of Obukhov. [1] Priestley [2] suggested that the Richardson [3] number was the sole controlling parameter of similarity near the earth's surface. Two significant entities that have a profound effect on dynamic similarity and the surface energy balance must be added to Priestley's corollary, the aerodynamic roughness length, and the latent heat flux via local evapotranspiration rates. Surface roughness length has significant effects on vertical wind shear; larger shears are associated with larger values of surface roughness. Large wind shears and increases in the vertical latent heat flux caused by increased evapotranspiration have a stabilizing effect on the atmosphere. The addition of water vapor to the atmosphere reduces the sensible heat flux, which leads to the stabilizing phenomenon.

An orderly and comprehensive examination of the surface energy balance requires detailed analyses of the vertical profiles of wind, temperature, and humidity with respect to atmospheric stability. Application of dynamic similarity principles to the micrometeorological characteristics of the surface boundary layer is interwoven with the nature of turbulent flow in which the instantaneous orthogonal velocities exhibit irregular, random fluctuations so that only the statistical properties can be subjected to analyses. Thus, profile analysis is generally based on averages formed over some arbitrary time interval and at preselected measurement levels on an instrumented tower or mast. Analytical arguments concerning the best approaches to defining the profiles are undertaken.

2. Buoyancy in the Surface Layer

Buoyancy is an atmospheric property of an air parcel that allows it to ascend and remain suspended in a compressible fluid such as the atmosphere. In the atmosphere, a buoyant force on a parcel of air may be directly attributed to a local increase in temperature resulting from the heating of the atmosphere at the air-earth interface by conduction, which immediately leads to a vertical energy transfer by convection. The work done by buoyancy against gravity results in a decrease with height of potential temperature, a characteristic of an unstable atmosphere.

Conversely, during the nocturnal hours, in the absence of incoming solar energy, the kinetic energy of the turbulent fluctuations will decrease, because of the extraction of energy from the mean motion by the Reynolds stresses. The atmosphere becomes quiescent and the mean flow approaches laminar flow conditions. The atmosphere is said to be stable.

The stability of the atmosphere may be expressed as the ratio of mechanical production to the convective production of energy. This ratio is generally known as the gradient form of the Richardson [3] number or the Obukhov [1] scaling length. Richardson, in the development of his criterion of turbulence, did not fully consider the effects of water vapor on atmospheric stability. Obukhov also omitted this important detail in his arguments leading to the dynamic similarity of flows theory. Lumley and Panofsky [4] discussed the importance of water vapor to the dynamics of the atmosphere. Busch [5] redefined the Obukhov length L with respect to the affect of water vapor and the latent heat flux on atmospheric stability as

$$L = - \frac{u_*^3 c_p \rho \theta}{kg(H + 0.07 \mathcal{L} E)} = \frac{u_*^2 \bar{\theta}_v}{kg \theta^*} \quad (1)$$

where

- u_* = a scaling or friction velocity
- c_p = the specific heat of air at constant pressure
- ρ = density

k	=	Karman's constant
g	=	gravitational acceleration
H	=	vertical sensible heat flux
$\mathcal{L}$	=	latent heat of vaporization
E	=	evaporation rate
$\mathcal{L} E$	=	latent heat flux
$\bar{\theta}$	=	mean potential temperature
$\bar{\theta}_v$	=	virtual potential temperature
θ^*	=	virtual scaling temperature.

By definition, the scaling length L is related to the gradient Richardson number, R_i , by

$$\frac{z}{L} = R_i \phi_M \frac{K_H}{K_M} = R_i \frac{\phi_M^2}{\phi_H} \quad (2)$$

where

z/L	=	Monin and Obukhov [6] scaling ratio
K_H	=	exchange coefficient for heat
K_M	=	exchange coefficient for momentum
ϕ_M	=	dimensionless lapse rate.

Based on the O'KEYPS* function, it may be demonstrated that in unstable flow $z/L = R_i$, because $\phi_H = \phi_M^2$, and in a thermally stratified stable regime, $z/L = R_i \phi_H$, because $\phi_H = \phi_M$. Therefore, in an unstable regime,

$$\frac{z}{L} = R_i = \frac{g \Delta \bar{\theta}_v}{\bar{\theta}_v (\Delta \bar{v})^2} z_g \Delta \ln Z \quad (3)$$

*O'KEYPS is an acronym formulated by Panofsky [7] and Yaglom [8] and recognizes the major contributors to the development of contemporary dynamic similarity theory Obukhov, [1] Kazanski and Monin, [9] Ellison, [10] Yamamoto, [11] Panofsky, [12] and Sellers. [13]

where

$$\begin{aligned}\bar{v} &= \text{mean horizontal wind speed} \\ z_g &= \text{geometric mean height of the layer } z_2 - z_1.\end{aligned}$$

For stable mean flow conditions, Hansen [14] found that $\phi_M = (1 + 15 R_i)$, allowing z/L to be expressed as

$$\frac{z}{L} = R_i (1 + 15 R_i). \quad (4)$$

The use of virtual potential temperature for surface boundary layer physical processes is basic to proper and accurate solutions. Virtual potential temperature is the temperature necessary for dry air to produce the actual density at ambient pressures, expressed as

$$\bar{\theta}_v = \frac{\bar{\theta}}{1 - 0.379 e/p} \quad (5)$$

or

$$\bar{\theta}_v = \bar{\theta}(1 + 0.61 \bar{q}) \quad (6)$$

where

$$\begin{aligned}e &= \text{partial pressure of water vapor} \\ p &= \text{atmospheric pressure} \\ \bar{q} &= \text{mean specific humidity.}\end{aligned}$$

3. The Mechanics of Vertical Profiles

Investigation and examination of surface boundary layer phenomena usually involve the collection of representative experimental data. In the interest of economy, not only in collecting experimental data, but also in processing samples, vertical profiles are usually observed. This means data are analyzed using bulk aerodynamic methods.

Traditionally, the instrumentation of towers and masts has been at heights above the surface established by application of the law of natural growth. This results in measurements at instrument levels spaced in a geometric or logarithmic progression. It also suggests that the vertical gradients are tangent to the profiles at the geometric mean height of each layer. Rachele, Tunick, and Hansen [15] investigated the validity of this premise via the mean value theorem of calculus; finding that the true point of tangency was not at the geometric mean height of a layer, but 8.2 percent above $z_g = (z_1 \cdot z_2)^{1/2}$. Equation (3) for R_i is now written as

$$R_i = \frac{g}{\bar{\theta}} \frac{\Delta \bar{\theta}_v}{(\Delta v)^2} \Delta z \quad (7)$$

because, from the mean value theorem,

$$z = \frac{\Delta z}{\Delta \ln z} \quad (8)$$

This is only of academic interest because values of R_i calculated for heights of z_g or z translate into identical values of L . The important point is that data analysis, using either the law of natural growth or the mean value theorem, yields cogent estimates of atmospheric stability.

The stochastic nature of turbulent flow implies that statistical methods must be invoked to temporarily smooth the data for each tower or mast level. Customarily, the Richardson numbers or Monin-Obukhov scaling ratios are formed at this point in the analysis. An inspection of the stability parameters usually reveals a nonuniform distribution with height, and on

conversion to Obukhov lengths, a variable L with height. Dynamic similarity requires a constant scaling length throughout the surface layer. The variability of L with height can be attributed to production and dissipation of mechanical and convective energy and eddies of several orders of magnitude in size impinging on or completely bypassing tower instrumentation. Irregular buoyancy and subsidence effects on anemometers or temperature and humidity sensors yield nonstationary appearing mean profile data.

Smoothing the profile heights can be accomplished in various ways. The simplest method is smoothing by eye (graphing the averaged data points as a function of height) and hand fitting a smooth curve to the points. Values picked from the curve are used to recalculate the stability parameters. A better fit can possibly be determined using a technique attributed to Businger et al. [16] For unstable flow conditions, differentiated second-order polynomials in $\ln z$ are fitted. For the stable regime, the polynomial $a + b \ln z = cz$ was used. The smoothed values for each level of interest are used to evaluate the stability.

Additional smoothing of the stability parameters is required to provide best estimates of the Obukhov length. Lettau [17] introduced the concept of the approximate height derivative of the gradient Richardson number, that may be used as a smoothing technique, and is given by

$$L^{-1} = \frac{\Sigma(R_i)_z}{\Sigma(z)_i} \quad (9)$$

where

$$i = 1, 2, 3$$

which is valid for unstable conditions. For the stable regime

$$L^{-1} = \frac{\Sigma(z/L)}{\Sigma(z)} \quad (10)$$

after determining z/L from equation (4). This technique is referred to as a stationarity forcing function.

4. Dynamic Similarity in the Surface Layer

The O'KEYPS designation has been reserved to describe the functions associated with the linear-quartic profiles in unstable flow. We chose to use the acronym as an all-encompassing descriptor for both the unstable as well as the log-linear stable regime. The O'KEYPS originators, however, were not the sole contributors to the investigation of stable mean flow conditions in the surface boundary layer; there were many others, including Webb, [18] Oke, [19] and Hicks. [20]

During this investigation concerned with water vapor effects on stability, it soon became apparent that a number of simplifying shortcuts could be used in the data analyses. These abridged procedures were unique enough to deserve a separate moniker and were dubbed *Mariah*.^{*} O'KEYPS and *Mariah* yield identical results, but *Mariah* requires much less computational effort.

The wind, temperature, and specific humidity profiles in the surface boundary layer of the atmosphere, written in differential form are

$$\frac{\partial \bar{V}}{\partial z} = \frac{u^*}{kz} \phi_M \quad (11)$$

$$\frac{\partial \bar{\theta}}{\partial z} = \frac{T^*}{kz} \phi_H \quad (12)$$

**Mariah has no relationship to any meteorological phenomenon or individual, but was unabashedly lifted from the Broadway Musical and motion picture "Paint Your Wagon", in which Uncle Willie sang*

"....way out there they have a name for rain and wind and fire, the rain is Tess, the fire is Joe, and they call the wind Mariah."

and

$$\frac{\partial \bar{q}}{\partial z} = \frac{q^*}{kz} \phi_w \quad (13)$$

where

T^* = a scaling temperature associated with potential temperature
 q^* = a scaling humidity.

Within the framework of dynamic similarity, it is presumed that the transfer of heat and mass are identical, so $\phi_w = \phi_H$. From the O'KEYPS function, it is easily demonstrated that for unstable conditions

$$\phi_M = (1 - 15 z/L)^{1/4} \quad (14)$$

and

$$\phi_w = \phi_H = (1 - 15 z/L)^{1/2}. \quad (15)$$

Integrating equations (11), (12), and (13) using equations (14) and (15) yields

$$\bar{V} = \frac{u^*}{k} [\ln z/z_o + \psi_M(z/L)] \quad (16)$$

$$\bar{\theta} - \theta_o = \frac{T^*}{k} [\ln z/z_o + \psi_H(z/L)] \quad (17)$$

and

$$\bar{q} - q_0 = \frac{q^*}{k} [\ln z/z_0 + \psi_M (z/L)]. \quad (18)$$

The diabatic influence functions are given by

$$\psi_M (z/L) = - \int_{z_0}^z \frac{1 - \phi_M}{z/L} d (z/L) \quad (19)$$

and

$$\psi_H (z/L) = - \int_{z_0}^z \frac{1 - \phi_H}{z/L} d (z/L) \quad (20)$$

and solved using Benoit's [21] method

$$\psi_M (z/L) = - \left\{ \ln \left[\frac{(\mathcal{L}_0^2 + 1) \mathcal{L}_0 + 1)^2}{(\mathcal{L}^2 + 1) (\mathcal{L}_0 + 1)^2} \right] + 2 [\tan^{-1} (\mathcal{L}) - \tan^{-1} (\mathcal{L}_0)] \right\} \quad (21)$$

where

$$\mathcal{L} \equiv \left[1 - 15 \frac{z}{L} \right]^{1/4}$$

$$\mathcal{L}_0 \equiv \left[1 - 15 \frac{z_0}{L} \right]^{1/4}$$

and

$$\psi_H (z/L) = 2 \ln \left(\frac{\lambda_o + 1}{\lambda + 1} \right) \quad (22)$$

where

$$\lambda = \left[1 - 15 \frac{z^{1/2}}{L} \right]$$

$$\lambda_o = \left[1 - 15 \frac{z_o}{L} \right].$$

The scaling temperature for unstable flow may be found from

$$T^* = \left(\frac{z}{L} \right) \frac{\theta u_*^2}{k z g} \quad (23)$$

and according to Myrup [22], q_o , the specific humidity at z_o , is determined from

$$q_o = \frac{RH}{1000} \left[3.74 + 2.64 \left(\frac{T_o}{10} \right)^2 \right] \quad (24)$$

where

T_o = the temperature at Z_o in degree Celsius.

Equations (16), (18), and (23) now can be evaluated for u^* , q^* , and T^* , the initial scaling parameters required to solve the surface energy balance. This approach is laborious because it is necessary to first evaluate the diabatic influence functions in equations (21) and (22).

A second approach is to guess a value for the Obukhov length L (negative for unstable conditions and positive for stable). Using the value for L , a solution is found for equations (21) and (22). For known values of $\bar{V}$, θ , and q , u^* , T^* , and q^* are evaluated using equations (16), (17), and (18). The values are used to determine $H = -c_p \rho u^* T^*$ and $\mathcal{L}E = -\mathcal{L} \rho u^* q^*$ that are substituted in equation (1) to obtain a new estimate of L . This value of L is used to form new estimates of u^* , T^* , and q^* . The process is repeated until the scaling constants converge. The second approach is not very efficient, even with machine processing.

In a thermally stratified stable regime in which $z/L = R_i \phi_M$ and $\phi_H = \phi_M$, the integration of equations (11), (12), and (13) yields

$$\bar{V} = \frac{u^*}{k} \left[\ln z/z_0 + \beta \frac{z}{L} \right] \quad (25)$$

$$\bar{\theta} - \theta_0 = \frac{T^*}{k} \left[\ln z/z_0 + \beta \frac{z}{L} \right] \quad (26)$$

and

$$\bar{q} - q_0 = \frac{q^*}{k} \left[\ln z/z_0 + \beta \frac{z}{L} \right] \quad (27)$$

where

β = a variable according to Hansen. [14]

If $\phi_M = 1 + \beta z/L$, then from equation (14)

$$\beta = \frac{15}{\phi_M} \quad (28)$$

where

$$\phi_M = (1 + 15 R_i).$$

The scaling parameters may then be found from

$$u^* = \frac{k \Delta \bar{V}}{\Delta \ln z + \beta \frac{\Delta z}{L}} \quad (29)$$

$$T^* = \frac{k \Delta \theta}{\Delta \ln z + \beta \frac{\Delta z}{L}} \quad (30)$$

and

$$q^* = \frac{k \Delta q}{\Delta \ln z + \beta \frac{\Delta z}{L}} \quad (31)$$

As with the unstable case, a value of L is needed to obtain values of u^* , T^* , and q^* using equations (29), (30), and (31). This is done by employing an iterative approach similar to the unstable case.

5. The Mariah Approach

The Mariah approach evolved from a study by Hansen. [23] From the definition of the Obukhov [1] scaling length of equation (1), repeated here as

$$L = \frac{u_*^H \bar{\theta}_v}{kg \theta^*} \quad (32)$$

where

$$\theta^* = T^* + 0.61 \theta q^*, \quad (33)$$

we may substitute equations (11), (12), (13), and (32) in equation (1), yielding

$$L = \frac{\phi_H \bar{\theta}_v (\Delta V)^2}{g \Delta \ln z \phi_M^2 (\Delta \theta + 0.61 \theta \Delta q)}, \quad (34)$$

which is valid in both the stable and unstable regimes, and allows a direct calculation of L from the observed gradients. However, in stable conditions $\phi_H = \phi_M$; therefore, the denominator contains ϕ_M . Manipulation of equation (33) reveals that in stable flow

$$L = \frac{\bar{\theta}_v (\Delta V)^2}{g \Delta \ln z (\Delta \theta + 0.61 \theta \Delta q)} - 15 \frac{\Delta z}{\phi_M \ln z}, \quad (35)$$

which leads to

$$L = \frac{B^2}{60 Z + 2B} \quad (36)$$

with $z = \Delta z / \Delta \ln z$, and

$$B = \frac{2 \bar{\theta}_v (\Delta V)^2}{g \Delta \ln z (\Delta \theta + 0.61 \bar{\theta} \Delta q)}, \quad (37)$$

a solution for the Obukhov length independent of the dimensionless wind shear.

The scaling constants u^* , T^* , and q^* are calculated using the discrete forms of equations (11), (12), and (13),

$$u^* = \frac{kz \Delta V}{\phi_M \Delta Z} = \frac{k \Delta V}{\phi_M \Delta \ln Z} \quad (38)$$

$$T^* = \frac{kz \Delta \theta}{\phi_M \Delta Z} = \frac{k \Delta V}{\phi_M \Delta \ln Z} \quad (39)$$

and

$$q^* = \frac{kz \Delta q}{\phi_H \Delta Z} = \frac{k \Delta q}{\phi_M \Delta \ln Z}, \quad (40)$$

and while evaluating ϕ_M using equation (14), noting that $z = \Delta z / \Delta \ln z$.

6. Discussion

The Mariah approach is simpler than previous solutions involving similarity. Aside from directly estimating the Obukhov scaling length L , Mariah produces results that do not differ numerically from the O'KEYPS equations. Mariah operates using measured gradients, and O'KEYPS operates using chosen trial values and iterating until convergence. Hand calculation using the O'KEYPS equations is tedious.

Figure 1 shows a comparison of scaling specific humidities determined using both schemes. The significant feature is the evaluation of q^* in the stable regime. During the nocturnal hours, evaporation continues to the point before the humidity profiles go isohumic and invert. Inversion of the specific humidity profiles is the first step toward dew fall. The transfer of water vapor from the atmosphere to the surface appears to have little or no effect on the application of dynamic similarity to the surface boundary layer.

Figure 2 shows specific humidity effects on atmospheric stability. An increase of water vapor in the atmosphere in conjunction with larger latent heat fluxes reduces absolute values of the Richardson number or Obukhov scaling length 4 to 10 percent.

Brunner [24] suggests that in the Planetary Boundary Layer relative humidity can be considered a constant in approximately the first 100 m of the atmosphere. Thus, relative humidities calculated from wet and dry bulb temperatures at shelter heights can be easily translated to vapor pressures and specific humidities at tower heights. Although it is not the most exacting method, it will provide the moisture corrections needed for surface energy balance studies.

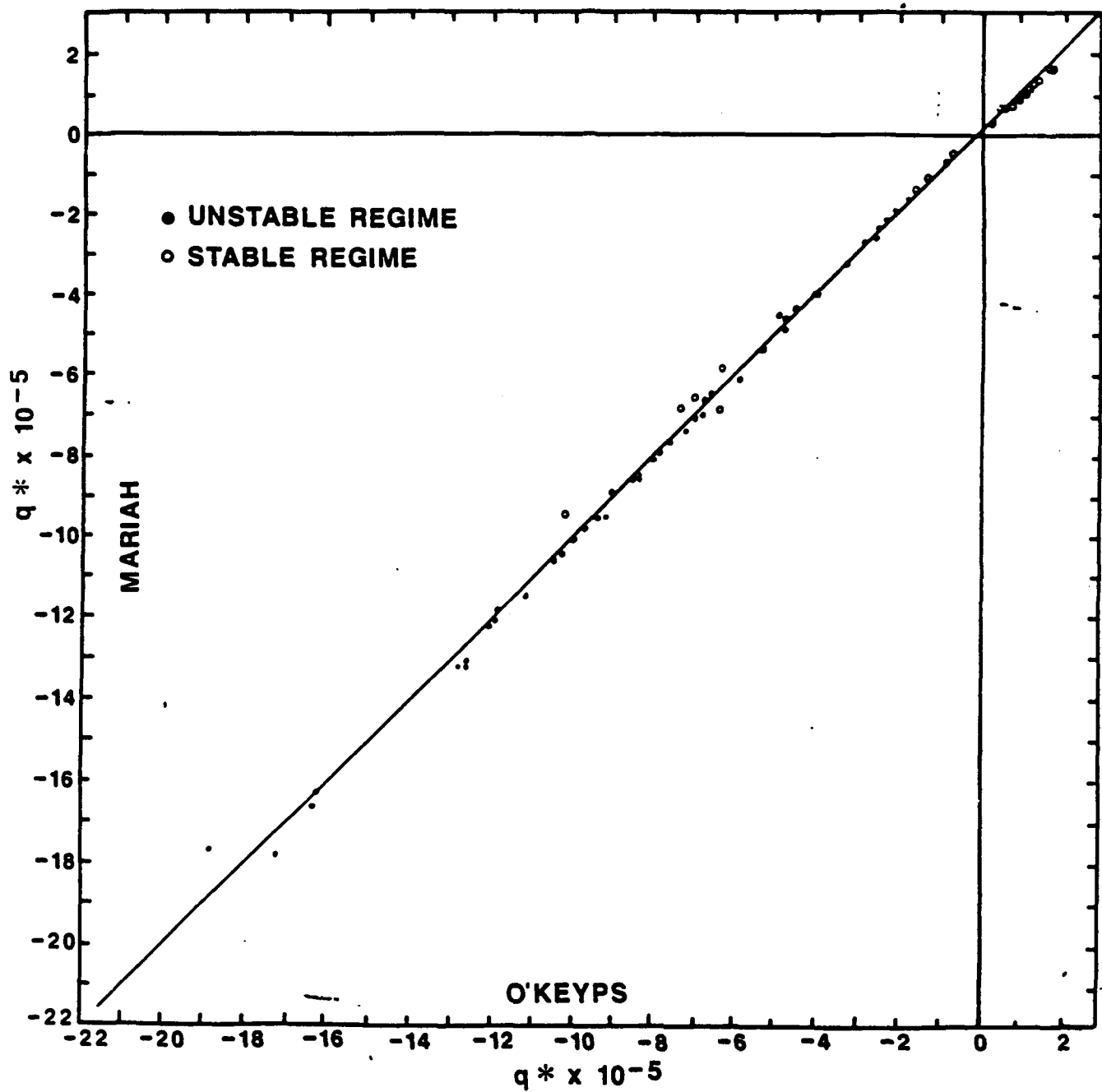


Figure 1. Comparison of scaling specific humidities determined using both schemes.

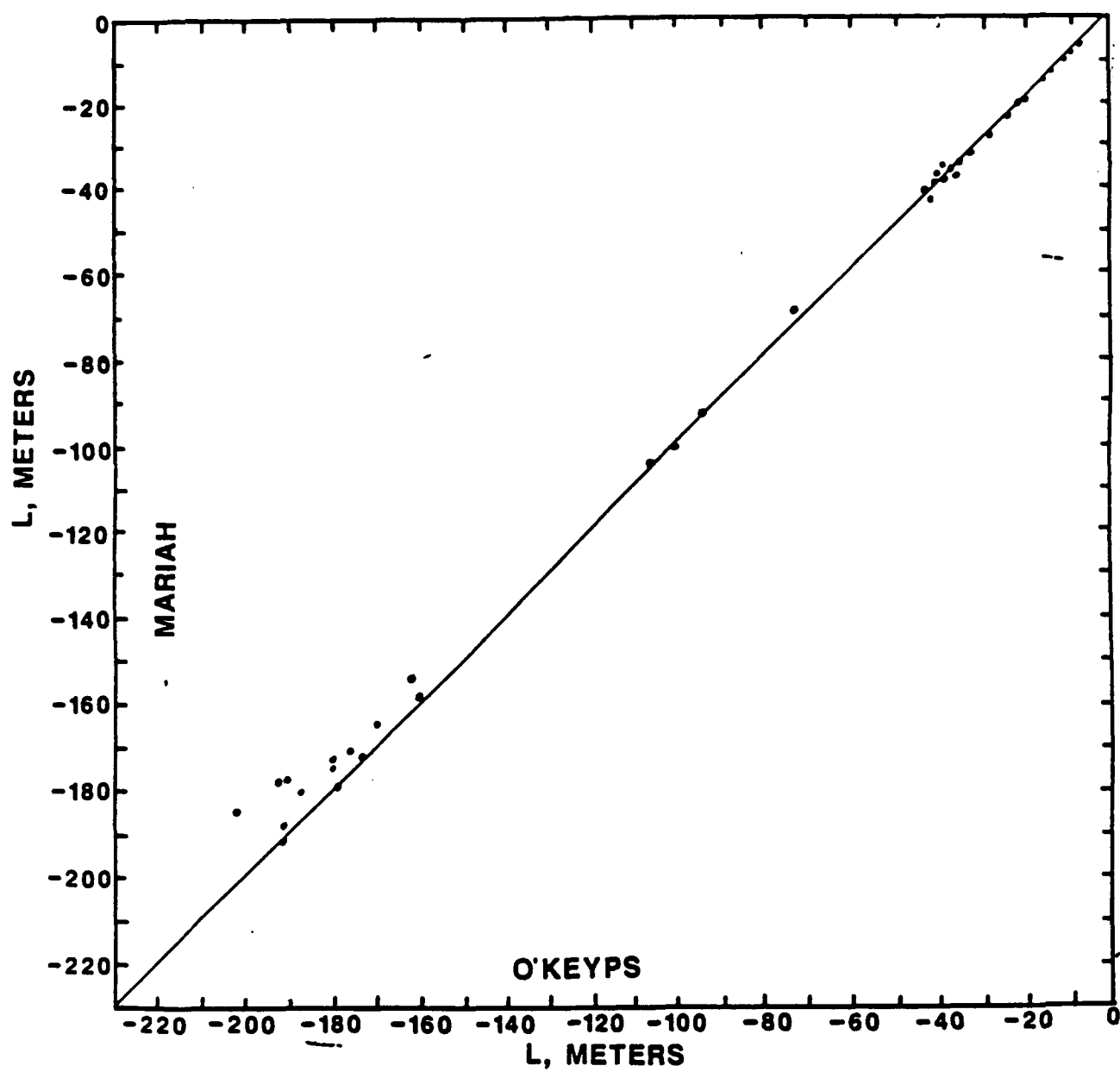


Figure 2. Specific humidity effects on atmospheric stability.

7. Conclusions

The Mariah approach to similarity theory profile analysis is simpler than that of the O'KEYPS methodology, especially if the more formal analyses involving the diabatic influence functions $\psi_M(z/L)$ and $\psi_H(z/L)$ are used to evaluate the scaling parameters u^* , T^* , and q^* in the unstable regime.

O'KEYPS and Mariah appear to function well in the stable regime — especially after inclusion of water vapor in the models. The approach to an isohumic state and the inversion of the specific humidity profiles seems to have little or no effect on the profile analyses schemes, indicating that expansion of the Mariah scheme by estimating specific humidity profiles for shelter height relative humidities is highly feasible for the entire stability spectrum.

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