Relative Efficiencies of Turbulent Transfer of Heat, Mass, and Momentum over a Patchy Urban Surface

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ABSTRACT

This study uses observational data from a suburban site in Vancouver, British Columbia, Canada, to investigate the relative facility with which heat, water vapor, and momentum are transported by turbulence in the unstable surface layer. The ratios of linear correlation coefficients $-r_{wT}/r_{uw}$ and $-r_{wq}/r_{uw}$ increase approximately linearly with instability and are generally smaller than typical rural values due to bluff-body effects. The ratio r_{wT}/r_{wq} is greatest near neutral and larger than unity at all stabilities. This inequality may be caused by the complex source/sink patterns of the urban surface, cloud effects on the radiative forcing, and the unusually well-developed interaction between the surface and the upper portions of the urban boundary layer. Inequality of transfer between T and q will make it difficult to measure turbulent fluxes for cities using standard gradient approaches.

1. Introduction

Indirect approaches to the evaluation of the turbulent fluxes of heat, water vapor, and momentum in the surface layer, such as the aerodynamic or Bowen ratio–energy balance methods, use assumptions regarding the relative transfer efficiencies of the entities involved (usually expressed through the turbulent diffusivities). While there may not be universal agreement, there is a fairly general consensus founded on comprehensive empirical evidence (e.g., Dyer and Hicks 1970; McBean 1970; Businger et al. 1971; Pruitt et al. 1973; Dyer 1974) regarding the transfers of momentum, heat, and water vapor over extensive and relatively homogeneous surfaces with low roughness. A similar body of work is not available regarding the validity of such assumptions or relationships for the surface layer over urban terrain.

McCormick and Kurfis (1966) used measured profiles of direct solar radiation and air temperature in the lowest 100 m over Cincinnati to infer effective vertical exchange coefficients of aerosol and heat. They conclude that the coefficients are not equal. The evidence, however, is neither direct nor comprehensive. More recently, Rotach (1993) showed that for slightly unstable conditions the nondimensional gradient of potential temperature measured in the vicinity of an urban canyon is generally lower than usually observed in the surface layer. The same was true for the dimensionless gradient of wind shear except that the use of local values resulted in good agreement with expressions for the surface layer.

Roth and Oke (1993) and Roth (1993) show that turbulent transfer over a suburban surface in Vancouver, British Columbia, Canada, generally conforms to Monin–Obukhov similarity (MOS) scaling during near-neutral and unstable conditions, but there are differences from results over more homogeneous terrain. With the exceptions of longitudinal wind speed and momentum flux, the shapes and position of the peaks of suburban (co)spectra agree with reference data, but the normalized dissipation rates and spectral correlation coefficients have different magnitudes. While the general behavior of the integral statistics (e.g., turbulence intensities, normalized standard deviations, and linear correlation coefficients) also agree with reference results from rural sites, again there are differences in magnitude.

The objective of this paper is to examine turbulent exchange mechanisms (especially those of heat and water vapor) over urban surfaces in more detail. This is done through analysis of the turbulent linear correlation coefficients defined by

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$$r_{uw} = \overline{u'w'}/\sigma_u\sigma_w \quad (1a)$$

$$r_{wT} = \overline{w'T'}/\sigma_w\sigma_T \quad (1b)$$

$$r_{wq} = \overline{w'q'}/\sigma_w\sigma_q \quad (1c)$$

$$r_{Tq} = \overline{T'q'}/\sigma_T\sigma_q, \quad (1d)$$

where u, w are the longitudinal and vertical wind velocity, respectively; T is air temperature; q is the absolute humidity; and σ stands for standard deviation. The prime denotes the instantaneous deviation from the time-mean value (indicated by the overbar). The correlation coefficients can be viewed as a measure of the overall efficiency of the transfer and varies between 0 (no correlation) and 1 (optimally efficient transfer).

The results are presented as relative transfer efficiencies to facilitate the comparison of the individual transfer mechanisms:

$$-\frac{r_{wT}}{r_{uw}} = -\frac{u_*T_*}{\sigma_w\sigma_T}\cdot\frac{\sigma_u\sigma_w}{u_*u_*} = \frac{\sigma_u/u_*}{\sigma_T/T_*} \quad (2a)$$

$$-\frac{r_{wq}}{r_{uw}} = -\frac{u_*q_*}{\sigma_w\sigma_q}\cdot\frac{\sigma_u\sigma_w}{u_*u_*} = \frac{\sigma_u/u_*}{\sigma_q/q_*} \quad (2b)$$

$$\frac{r_{wT}}{r_{wq}} = -\frac{u_*T_*}{\sigma_w\sigma_T}\cdot\frac{\sigma_w\sigma_q}{u_*q_*} = \frac{\sigma_q/q_*}{\sigma_T/T_*}, \quad (2c)$$

where u_*, T_*, and q_* are the usual surface-layer scales defined through $-\overline{u'w'}$, $\overline{w'T'}/u_*$ and $\overline{w'q'}/u_*$, respectively. Equation (2) shows that the correlation coefficients in (1) and their respective ratios are related to their normalized standard deviations. It follows that any differences observed in the suburban results compared to reference data can be discussed in terms of the more fundamental normalized standard deviations.

The following relationships, which are based on MOS scaling laws, are used to represent the observations from the homogeneous surface layer:

$$\sigma_u/u_* = 2.2(1 - 3z'/L_v)^{1/3} \quad (3)$$

$$\sigma_w/u_* = 1.3(1 - 3z'/L_v)^{1/3} \quad (4)$$

$$\sigma_T/T_* = 0.95(-z'/L_v)^{-1/3} \quad (z'/L_v < -0.03) \quad (5)$$

$$\sigma_q/q_* = 1.04(-z'/L_v)^{-1/3} \quad (z'/L_v < -0.1), \quad (6)$$

where z' is the effective height above the zero-plane displacement and the Obukhov length $L_v = u_*^3\overline{T}_v/(gk\overline{w'T_v'})$ with $\overline{T}_v$ the mean absolute virtual temperature, g is the gravitational acceleration, and k is von Kármán's constant (here to be taken as 0.4).

Expressions (3)–(6) are from flat, grassy sites and were obtained by De Bruin et al. (1993), Panofsky et al. (1977), Wyngaard et al. (1971), and Högström and Smedman-Högström (1974), respectively. While there is general consensus regarding the applicability of (4) and (5), evidence that (3) should follow MOS is weak. It is often observed that statistics of horizontal wind components scale better with boundary layer height than with height above surface (e.g., Panofsky et al. 1977; Panofsky and Dutton 1984). In the case of (6) many studies report a lot of scatter, and it is questionable whether humidity can be treated as a local quantity, that is, described with MOS variables (see below). McBean (1971) concludes that humidity scales better with a stability parameter based on the moisture flux than on the heat flux. However, this scaling applied to the observations presented below did not result in improvement such as reduced scatter.

2. Observational data

The site for this study is a suburban area (called Sunset) located in south Vancouver. It consists of mainly single-family residential housing. The 1–2-story houses have a mean height of 8.5 m. Within a circle of 2-km radius centered on the site about 43% of the active or 3D surface area (which is approximately 1.5 times the map area) is green space, 13% is roof, 11% is paved, and 33% is wall (or canyon) (Grimmond et al. 1991).

The instantaneous velocities of the wind field were measured with a three-dimensional sonic anemometer, KD (Kaijo Denki, model DAT-310, probe: TR-61C), at z' = 19 m (22.5 m above ground minus 3.5 m zero-plane displacement). About 1 m to one side was a one-dimensional sonic anemometer and fine-wire thermocouple, SAT (Campbell Scientific, model CA27T), to measure fluctuations of vertical velocity and air temperature, and a Lyman-alpha hygrometer, Ly-α (ERC, model BLR, modification F1), for humidity fluctuations. A second set, consisting of a SAT and a Krypton hygrometer, KH (Campbell Scientific, model KH20), was mounted on the other side of the KD. An additional SAT/KH combination was installed at z' = 11 m. The temperature and humidity statistics at z' = 19 m presented below are averages from the two sensor combinations SAT/Ly-α and SAT/KH.

Signals were recorded at 25 Hz over a 60-min record length on a PC-XT acquisition system with a low-pass filter cutoff frequency of 10 Hz using sixth-order Butterworth maximum flat filters. All sensors were rotated into the mean wind before each run to minimize flow interference. The KD sonic anemometer data were further corrected for the flow distortion effects caused by the transducers themselves and the bulk of the array, and due to any misalignment of the sensors, according to the procedures given by Wyngaard (1981), Wyngaard and Zhang (1985), and Rotach (1991). The output of both types of hygrometer was corrected for density effects (Webb et al. 1980), and the KH units were corrected for temperature-induced fluctuations in oxygen following Tanner and Green (1989). A photograph and complete description of the instrumentation and data handling can be found in Roth and Oke (1993).

Observations were gathered on eight days in the period 5–15 July 1989. A total of thirty-eight 60-min runs were collected during 0800–2000 LAT [local apparent time; see Oke (1987)]. In this period the range of wind speeds was 1.2–8 m s^{-1}, air temperature 16°–28°C, and absolute humidity 8–12 g m^{-3}. The results are from unstable conditions in the range $-1.80 < z'/L_v < -0.05$.

3. Analysis of correlation coefficients

Plots of the ratios $-r_{wT}/r_{uw}$, $-r_{wq}/r_{uw}$, and r_{wT}/r_{wq} and atmospheric stability z'/L_v are given in Fig. 1. The ratio for heat/momentum increases with instability in an almost linear fashion—from a value near unity at neutral to about 4 at $z'/L_v = -1.8$ (Fig. 1a). A similar pattern is evident for the water vapor/momentum ratio. The ratio is approximately 0.4 near neutral and 3.0 at $z'/L_v \approx -1.5$ (Fig. 1b). Both ratios are smaller than the MOS predictions (especially for $-r_{wq}/r_{uw}$), which are derived by substituting (3) and (5) into (2a), and (3) and (6) into (2b), but the form of the trend is similar. The present observations are also lower but parallel in trend to those observed by McBean and Miyake (1972) for $-0.1 > z'/L_v > -0.7$ over a flat, extensive grass-covered site (Fig. 1a), and the rural data of McBean (1970) for $0.0 > z'/L_v > -0.7$ (Fig. 1b).

To understand the observed differences in Fig. 1 it is helpful to consider the respective normalized standard deviations, which are related to the correlation coefficients as shown in (2). From (2a) and (2b) and the data in Figs. 2a and 3, it follows that the lower suburban ratios in Figs. 1a and 1b are mainly a consequence of the relatively low σ_u/u_* values (note the suburban σ_T/T_* agrees well with the reference in Fig. 3a). In the case of the humidity/momentum ratio the relatively high σ_q/q_* values (Fig. 3b) result in a further reduction.

The relatively low normalized standard deviations compared to reference data (solid lines) of u (and also w and the transverse velocity v) plotted in Fig. 2 imply an enhanced momentum transport over the urban surface relative to representative rural values, which is due to increased roughness. Similar results have been obtained in urban–rural turbulence studies such as Bowne and Ball (1970), Ramsdell (1975), and Clarke et al. (1982) (Fig. 2a) and in the urban study by Rotach (1993). Further, Roth (1993) shows that $-r_{uw}$ at this suburban site is larger than the rural reference data of, for example, McBean (1970), decreasing from about 0.4 near neutral to 0.2 at $z'/L_v = -1$ and matching the reference data at large instabilities.

Increased momentum transport may not be the only explanation for the differences observed in Fig. 2, in particular for the ratios involving the horizontal wind components. Over a city surface, close to the roughness elements, production of mechanical energy is enhanced, but that of buoyant energy is reduced. Examination of urban u spectra (e.g., Jackson 1978; Roth and Oke 1993) shows that in comparison to rural data less energy (variance) is contained at the buoyancy-produced large scales. This will contribute to a decrease of the normalized velocity components. The uw cospectra are not affected in the same way because they relate mainly to mechanical energy, which is produced at smaller scales.

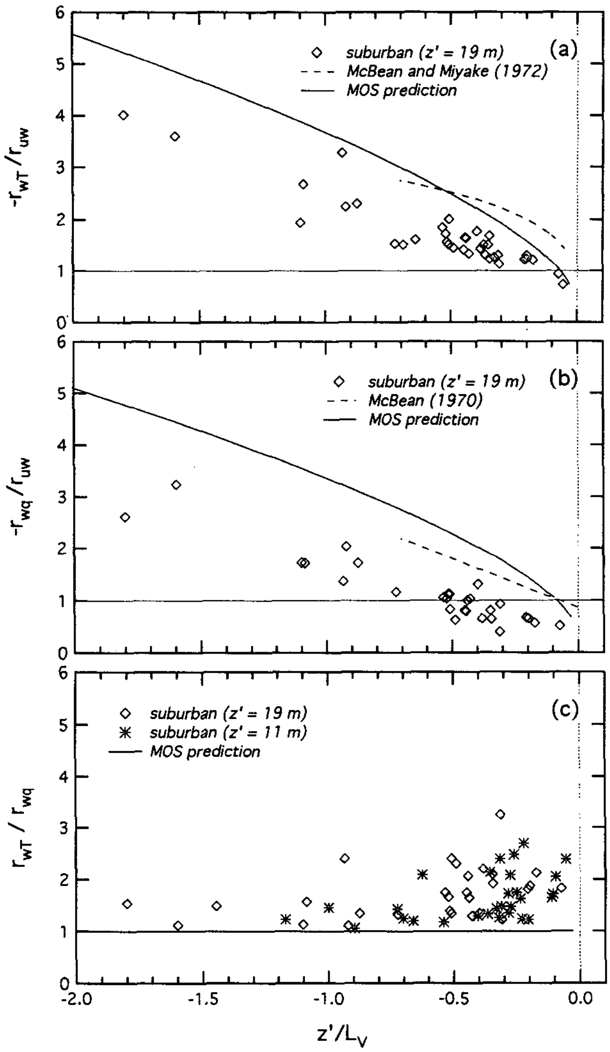

FIG. 1. Ratio of correlation coefficients versus stability for (a) heat/momentum, (b) water vapor/momentum, and (c) heat/water vapor transport. Dashed lines in (a) and (b) are empirical fits to the rural data of McBean and Miyake (1972) and McBean (1970). The solid lines represent the MOS prediction derived from (2) and the respective normalized standard deviations in (3)–(6).

At all stabilities the heat/water vapor ratio is greater than unity (Fig. 1c), and therefore the transfer of heat is more efficient than that of moisture. This is in contrast to the MOS prediction of unity, which implies that heat and water vapor are transported by the same mechanism in the homogeneous surface layer (see below).

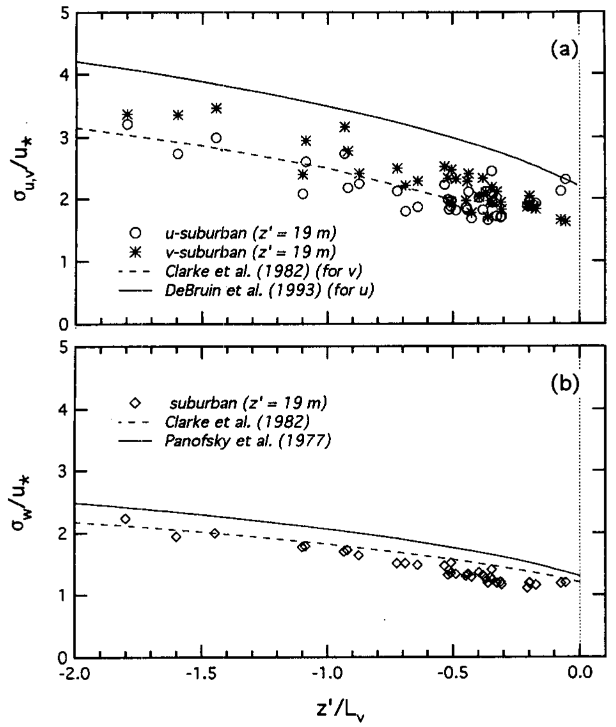

FIG. 2. Normalized standard deviations of (a) u, v and (b) w versus z'/L_v. The solid lines represent homogeneous surface-layer data [Eqs. (3) and (4)]. The dashed lines in (a) and (b) are from suburban data presented by Clarke et al. (1982) and given by $\sigma_v/u_* = 2.5(-z'/L_v)^{1/3}$ (for $z'/L_v < -1$) and $\sigma_w/u_* = 1.2(1 - 2.5z'/L_v)^{1/3}$, respectively (after Roth 1993).

It follows that the transfer mechanisms for the two variables are different at the present site. Although it is unwise to generalize too much, given the scatter of these data, it seems that the ratio is greatest in near-neutral conditions and declines toward unity for large instabilities. This generally agrees with results from a rural study by McBean (1970), except that the suburban ratios are larger. His data suggest a r_{wT}/r_{wq} value of about 1.4 near neutral dropping to about 1.2 at $z'/L_v = -0.6$. Figure 1c also shows that differences between the two observation levels are small. Note, however, that in this figure r_{wT}/r_{wq} at the lower level is plotted against z'/L_v evaluated at the upper level since u_* was not measured at $z' = 11$ m. The effect of using local variables is not certain but probably includes a slight shift of the data toward larger instabilities.

From (2c) and the data of Fig. 3 it follows that the larger ratios are mainly a consequence of the relatively large σ_q/q_* values. This can be due to enhanced momentum transfer because

$$\sigma_q/q_* = \sigma_q u_*/\overline{w'q'}. \tag{7}$$

Second, based on data measured over surfaces with different homogeneities and flux magnitudes, Weaver (1990) hypothesized that, if a flux contains contributions from sparsely distributed sources of a variable at the surface, they may become entrained in the flow and increase the variance of that variable but not necessarily its covariance. From (7) it follows that in this case the normalized standard deviation will also increase but the correlation coefficient would decrease [Eq. (1c)] as observed by Roth (1993).

Hill (1989) shows that, if MOS is valid for humidity as well as temperature, then $r_{wT} = r_{wq}$, and he suggests that differences between these correlations can be attributed to the differential action of buoyancy on the humidity and temperature field. This can occur only if temperature and humidity fluctuations are not perfectly correlated. This was the reasoning originally forwarded by Swinbank and Dyer (1967) to attribute the equality of the eddy diffusivities in their rural study to high $T - q$ correlations ($r_{Tq} > 0.9$).

Additional observational support on this point is given by Phelps and Pond (1971), using observations obtained over water, and by McBean and Miyake (1972) and McBean (1973), who suggest that, when r_{Tq} is close to unity and moisture is a passive scalar, the fluxes of heat and moisture have similar characteristics; however, when the r_{Tq} correlation is small the efficiency of the moisture transport is decreased. This fits

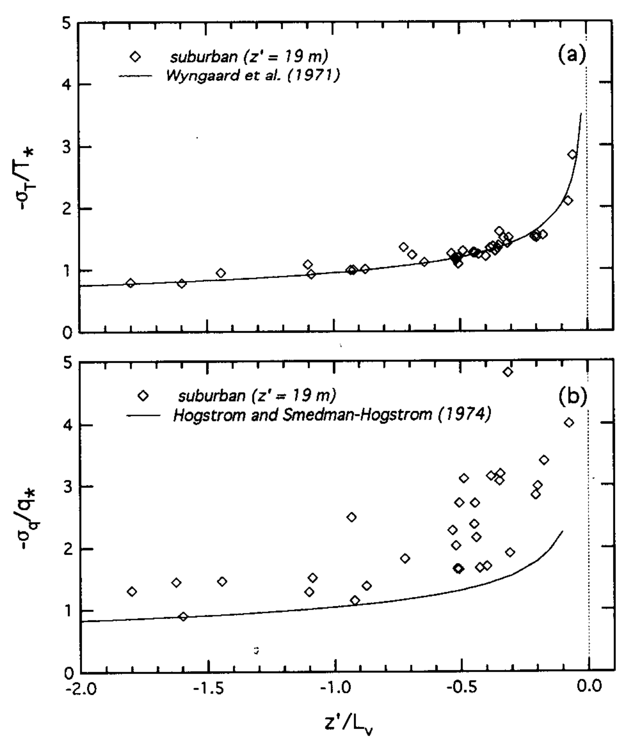

FIG. 3. Normalized standard deviations of (a) T and (b) q versus z'/L_v. The solid lines represent homogeneous surface-layer data [Eqs. (5) and (6)] (after Roth 1993).

exactly with the results of the present study. Figure 4 demonstrates that as r_{Tq} increases, r_{wT}/r_{wq} approaches unity. It also shows that the urban environment is generally characterized by low correlation between the temperature and moisture fluctuations (r_{Tq} typically < 0.80).

We conclude that there are several differences between the ratios of transfer efficiencies at suburban and rural sites. In near-neutral and slightly unstable conditions the transport of momentum over built-up regions is extremely efficient, leading to smaller ratios of $-r_{wT}/r_{uw}$ and $-r_{wq}/r_{uw}$ than reference data. Further, the transfers of heat and water vapor are different, such that $r_{wT} > r_{wq}$. In the following two sections we try to elucidate the processes likely to be responsible for the observed differences.

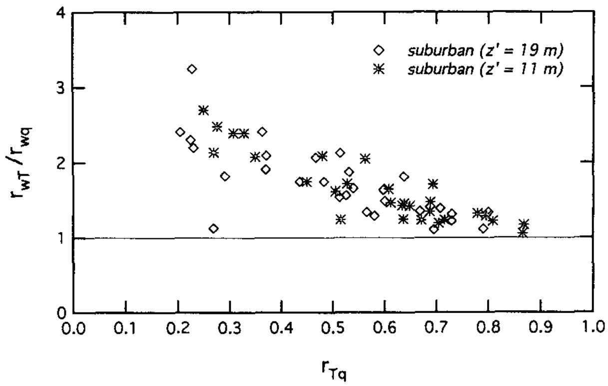

FIG. 4. Ratio of linear correlation coefficients for heat and water vapor (r_{wT}/r_{wq}) versus that for temperature and humidity (r_{Tq}).

4. Controls on urban diffusion

When the aerodynamically rough and inhomogeneous urban surface interacts with the airflow above and within the roughness elements, the following processes, which are usually not found (or are less explicit) over more homogeneous rural surfaces, occur.

1) Turbulent wakes are generated by the individual roughness elements, which convert the mean turbulent kinetic energy of the flow into turbulent kinetic energy (Raupach and Thom 1981).

2) Momentum is absorbed from the flow by both form and skin-friction drag on individual roughness elements.

3) Organized structures in the overlying boundary layer are important agents that determine the nature of the near-surface turbulence (see section 5).

Wake diffusion (Thom et al. 1975), that is, mixing generated by turbulent wakes behind individual roughness elements, acts as an additional turbulent diffusion mechanism for flow properties: Vortices shed from the edges and corners of buildings seed the shear flow with interacting wakes characterized by small eddies that efficiently mix and diffuse momentum, heat, moisture, or any other scalar quantity. *Form drag* due to bluff bodies (i.e., pressure differences across the individual roughness elements), on the other hand, augments the transport of momentum to the surface but has no analog in the transport of heat or mass. It follows that the transfer of properties other than momentum occurs by turbulent diffusion alone and is usually less efficient than momentum transfer (Thom 1972). The more efficient urban momentum transfer compared to the transport of heat and water vapor can therefore be attributed to the additional bluff-body forces. Similarly, McBean (1987) notes that much of the momentum transfer in cities is due to pressure perturbations associated with form drag on structures rather than viscous forces associated with skin friction.

Although wake diffusion has been identified by Raupach et al. (1980) and Hosker (1986) as an efficient mechanism for the diffusion of momentum, it is not certain whether it could contribute to the observed imbalance in Figs. 1a and 1b. Thom et al. (1975), for instance, suggest that mixing generated by turbulent wakes behind individual roughness elements acts as an additional diffusing mechanism, contributing more effectively to heat than momentum. This would be in contradiction to the present observations; however, Thom et al.'s suggestion applies only for a narrow range of instabilities close to neutral, and their findings have been questioned by Garratt (1978). He suggests that the direct influence of the surface upon turbulent diffusion is the same for both momentum and heat transfer. It seems that the only possibility of a differential wake diffusion effect would be the case where the principal momentum sink is not co-located with the sources/sinks of the scalars—for example, the heat (humidity) sources are located outside the zones affected by wake diffusion. Section 5c does show that this occurs over urban areas where the temperature and humidity fields close to the surface are strongly modulated by mixed-layer processes.

Since form drag does not contribute to the diffusion of heat and moisture, the transfers of these entities remain dependent upon turbulent mixing. The following section examines possible mechanisms for the differences observed in Fig. 1c.

5. Sources of poor r_{Tq} correlation

In general we may expect that the equality of transfers of heat and water vapor is applicable only in a homogeneous surface layer with humidity acting as a passive scalar and a $T - q$ correlation near unity (e.g., Hill 1989). From Fig. 4 it is concluded that the inequality of the eddy diffusivities for heat and water vapor in the city is related to poor correlations between T and

q. In the following analysis we follow the conclusion of McBean (1973) that when considering the transfer of passive scalars, the external influences that govern the active-scalar/passive-scalar correlation (in this case T/q) must be examined.

The ratio of the spectral correlation coefficients plotted in Fig. 5a demonstrates that the transfer efficiency of heat is larger than that of moisture at all frequencies, with little difference between the two measurement levels. The ratio increases slightly toward the larger scales, which coincides with a decrease of the $T - q$ spectral correlation coefficient (Fig. 5b). (Little significance is placed on the differences at the lowest frequencies where values become statistically unreliable.) These data therefore indicate that a variety of processes, which involve a range of scales, contribute to the dissimilarities and low $T - q$ correlations. The following section explores the possibility that these less-than-perfect correlations are caused by spatial inhomogeneities of the surface source/sink distribution and/or the surface radiative forcing or, for unstable conditions, the downward advection of air from the mixed layer.

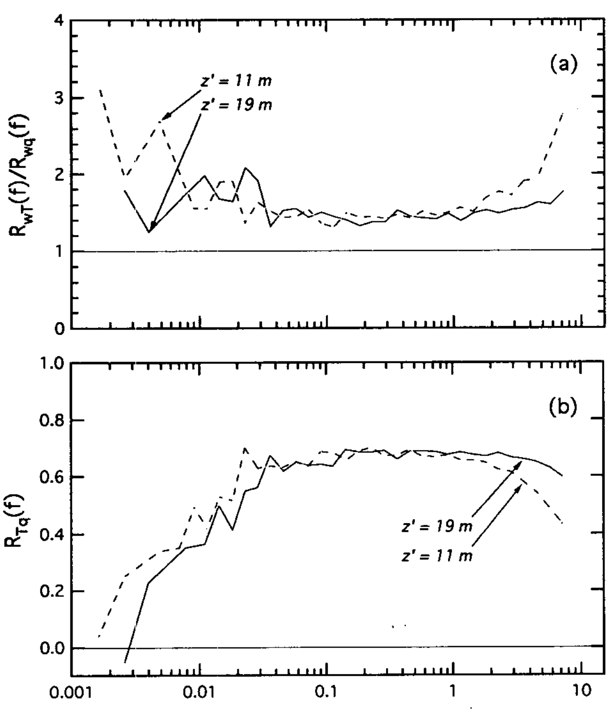

FIG. 5. Composite (average over all runs) (a) ratio of spectral correlation coefficients for heat and water vapor (R_{wT}/R_{wq}) and (b) spectral correlation coefficients for temperature and humidity (R_{Tq}) measured at two levels (z' = 11 and 19 m) versus nondimensional frequency f (n—natural frequency in Hz) [data from Roth and Oke (1993)].

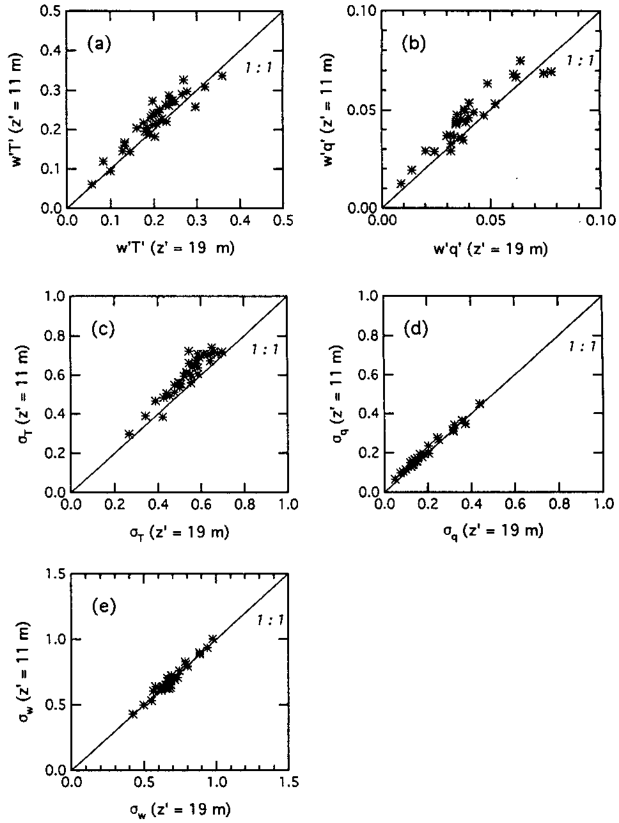

FIG. 6. Comparison of kinematic fluxes of (a) sensible heat, (b) water vapor and standard deviations of (c) temperature, (d) humidity, and (e) vertical velocity measured at two levels (z' = 11 and 19 m) over the suburban site.

a. Surface heterogeneity of the source/sink distribution

Inhomogeneity in surface structure can result in the sources of heat and water vapor being separate. This is very likely to be the case in the urban environment. For example, whereas all surfaces are sources of heat, they do not all possess water. Further, the availability of water for evaporation or transpiration may vary from unrestricted (open water or irrigation-wetted surfaces) to tightly held (dry soils or building materials). As a result, individual surfaces with Bowen ratio values ranging from large negative to large positive values coexist side by side. As a result, the surface thermal field of an urban area is dominated by the presence or absence of water: dry (wet) areas are associated with warm (cool) temperatures [e.g., the infrared imagery of Schmid and Oke (1992)].

Over an extensive dry surface the absence of water leads to a very low or negative r_{Tq}; over a large wet surface r_{Tq} is positive. Therefore, a system whose surface is made up of both types is likely to have a composite r_{Tq} that is significantly less than unity. Further, and just as importantly, the city consists of a succession of wet and dry patches: irrigated lawns, buildings, trees, streets, etc., so microscale advection at each leading edge is the norm. Airflow across a dry–wet interface results in a positive temperature but negative humidity change. The reverse is true in the case of a wet–dry interface. Note that both advective transitions tend to reduce r_{Tq}.

There are two other characteristics of the surface of a city that are likely to lead to low r_{Tq}. First, many of the urban sources of heat and moisture are arranged at different heights, even with strong vertical zonation (e.g., ground level versus roof level). In the present study it was noted that the fluxes of sensible and latent heat decreased upward (Fig. 6). In the case of sensible heat (latent heat) the mean flux densities were 272 (111) W m^{-2} at the lower and 248 (98) W m^{-2} at the upper level. The index of agreement (d_i) and root-mean-squared difference (rmsd) were 0.97 (0.97) and 25 (13) W m^{-2}, respectively [for a discussion of the statistical indices see Willmott (1981)]. In the case of the sensible heat flux sensors, the latter value compared to the inter-sensor rmsd of 11 W m^{-2} obtained at the same site, and $z' = 19$ m suggests that the observed differences are real and not only an instrumental bias. No such inter-sensor comparison is available for the humidity sensors. Whereas the standard deviation of temperature was generally higher at the lower level, the standard deviation of vertical velocity did not change appreciably between $z' = 11$ and 19 m (Fig. 6). The observed variation with height may well be due to the difference in the source areas for the two heights [see Schmid and Oke (1990) and Table 1 below].

Second, in addition to natural sites of heat and moisture release there are anthropogenic sources related to fuel combustion for heating, cooling, transportation, and industrial processing. These emanate from domestic and industrial chimneys, windows, doors, vehicle tailpipes, cooling towers, cooling ponds, etc. Their disparate source locations produce different average profiles of scalars within the urban "canopy" (Oke 1989) and imply differing transport out of this layer.

b. *Surface heterogeneity due to cloud effects on radiation*

About half of the observations were taken under cloudy conditions. A broken cloud field advecting across the landscape induces changing patterns of shade on the ground that result in spatially and temporally varying radiative input at a given location. In partly cloudy weather McNaughton (1992, personal communication) reports watching T and q traces that are in phase when the site is sunlit but in antiphase when first in shadow. Of itself this source of low r_{Tq} is not uniquely urban. However, it is possible that in combination with the greater and somewhat organized patterns of shade produced by buildings and streets and the wide range of thermal inertias of building materials (e.g., roofs versus roads), this is a greater source of poor correlation in a city. In addition to the example used here, where a cloud field moves across the landscape, it is also possible that air advecting through shade patterns cast by almost stationary clouds could operate like the mechanism suggested in relation to microscale advection and thereby lead to lower r_{Tq}.

c. *Large eddy convection*

In about 50% of the measurement runs there were signs that the signals were influenced by convective structures that extended well above the surface layer. Dissimilarity in the turbulent exchange can then be attributed in part to interaction with air from the mixed layer, or even above the capping inversion, where the $T - q$ correlation may be smaller, or have the opposite sign, to that near the ground.

An instructive example is provided by the time series given in Fig. 7. This trace is typical of about 25% of the runs, all of which are associated with partial cloud cover. During this hour the following mean conditions prevailed: net all-wave radiation, 349 W m^{-2}; sensible heat, 260 W m^{-2}; latent heat, 200 W m^{-2}; temperature, 21.4°C; absolute humidity, 8.3 g m^{-3}; wind speed, 3.66 m s^{-1}; wind direction, 165°; friction velocity, 0.49 m s^{-1}; $z'/L_v - 0.45$; and a stratocumulus cover.

The temperature trace shows a combination of high-frequency fluctuations together with some low-frequency (large eddy), ramplike structures that last for 120–360 s. They develop from a below-average temperature during a downdraft (indicated by the dashed lines on the T' panel in Fig. 7) followed by a slow, irregular increase that amplifies then ends with a sharp drop. These are probably associated with the passage of a large rising thermal followed by a cold downdraft. Note that the pattern observed in the present temperature trace is similar, but not the same, as often observed over flat, rural surfaces (e.g., Wilczak 1984). In the homogeneous case the temperature "ramps" only last for about 100 s or less and exhibit a clear sawtooth appearance. The $w'T'$ trace shows that these structures (positive w' and T' in thermals or negative w' and T' in the downdrafts) contribute a relatively large proportion of the net sensible heat flux. These are typical of well-developed convection cells in the mixed layer (e.g., Stull 1988). This interpretation is supported by the fact that the typical mixed-layer depth at this site and time of day is about 500 m (Steyn and Oke 1982), giving a convective mixed-layer velocity of about 1.5 m s^{-1} and a convective timescale of about 330 s.

The humidity signal is dominated by a few large negative deviations (up to 1.8 g m^{-3}), which last for 15 to

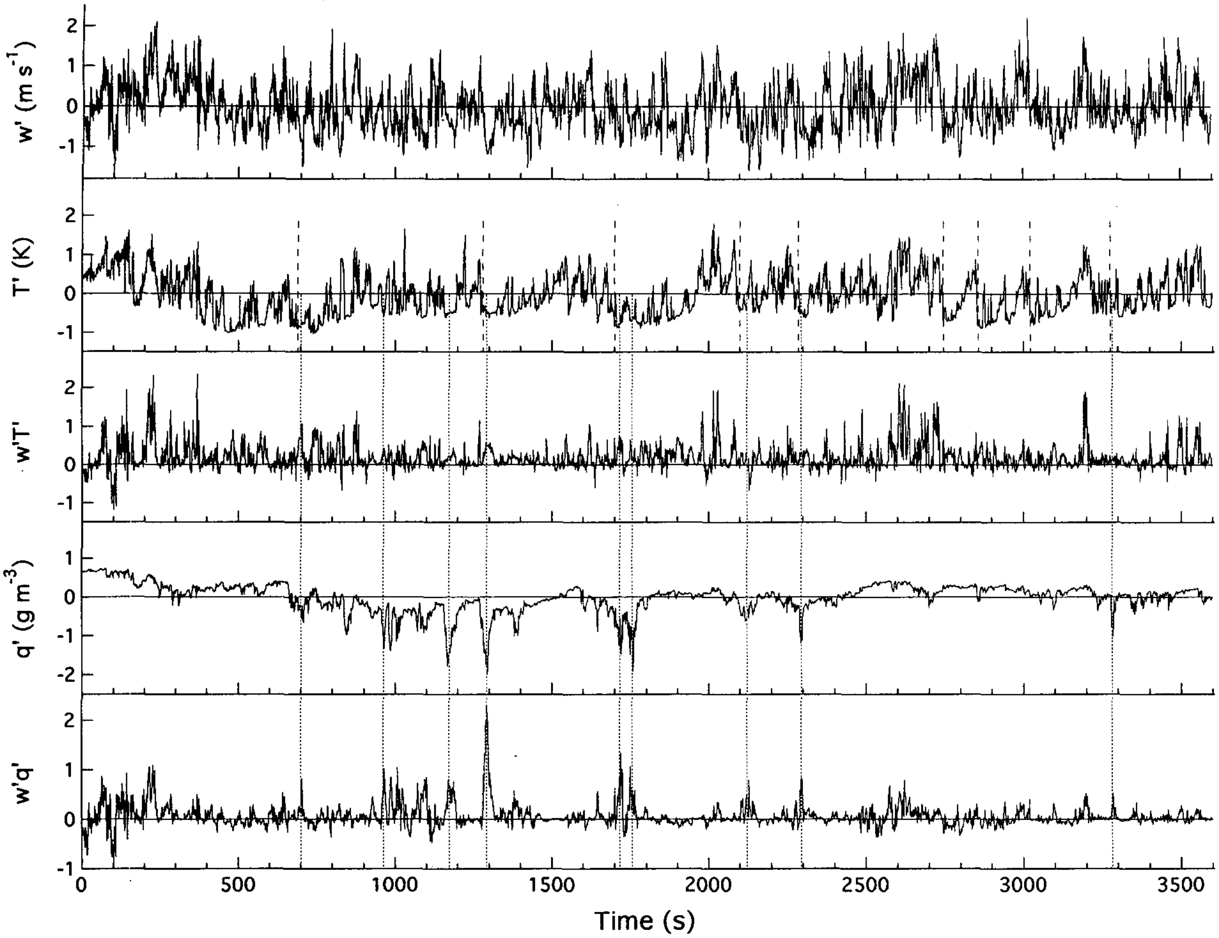

FIG. 7. Time series of vertical velocity (w'), air temperature (T'), absolute humidity (q'), and the covariances for heat ($w'T'$) and water vapor ($w'q'$) transport at the Sunset site on YD 188 between 1550 and 1650 LAT. Dashed lines on T' panel indicate approximate starting times of large eddy structures; dotted lines connecting T' and $w'q'$ panels visualize correlation between cool, dry downdrafts and positive moisture flux (see text).

60 s in conjunction with the downdraft portion of the temperature structures (indicated by the dotted lines connecting the T' and $w'q'$ panels in Fig. 7). These dry downdrafts account for the bulk of the evaporative flux, the rest being conducted by relatively quiet periods with small moist updrafts. The correlation coefficients involving q are low ($r_{Tq} = 0.38$ and $r_{wq} = 0.32$), and the q spectrum for this run (not shown) is characterized by considerable variance at large scales and the absence of a peak. This is in contrast to the observations over crops by Högström and Smedman-Högström (1974), Ohtaki (1985), and Anderson and Verma (1985), which show decreasing variance at large scales (but marked by a lot of scatter). In general the humidity spectra from the present urban study exhibit relatively large amounts of low-frequency energy, whereas the temperature spectra agreed fairly well with rural reference results (see Roth and Oke 1993).

Figure 7 suggests that evaporation is being driven by the import of dry air from higher in the mixed layer, which in turn may have been entrained from above the capping inversion and transported to the surface by large eddies. The degree of coupling between the surface layer and the rest of the planetary boundary layer (PBL) has been quantified by McNaughton and Jarvis (1983), who develop a parameter (Ω) that gives the appropriate weighting to the radiative and advective components in the Penman–Monteith evaporation model. It has values between zero (strongly coupled) and unity (decoupled). For the run in Fig. 7, with measured aerodynamic and surface conductances of 67 and 6.7 mm s^{-1}, respectively, the coupling is very efficient ($\Omega = 0.25$). This value is close to that suggested by McNaughton and Jarvis (1983) and observed by Lee and Black (1993) for rough forest surfaces. Typical values for low-plant covers are in the range 0.5–0.8

(McNaughton and Jarvis 1983). Low $T - q$ correlations have also measured by De Bruin et al. (1993) over a dry rural site. They suggest that in the case of strong coupling and dry conditions at the surface (Bowen ratio ≈ 2 at the Sunset site) large eddies will have a relatively large effect on evaporation at the surface. Production of turbulent kinetic energy through shear and buoyancy over rough and warm cities is usually larger than over nonforested rural areas and likely to lead to enhanced surface–PBL interaction including evaporation.

These interpretations fit well with previous urban boundary layer observations of turbulent transfer from aircraft surveys over St. Louis. A spatial analysis of the correlation coefficients at a height of 150 m found the local maximum r_{wT} to be large (about 0.8) and centered over the urban core in exactly the same location as the minimum in r_{wq} (about 0.1) (Ching 1985). Briggs (1987), commenting upon the same dataset, notes it appears that dry heat is the organizing agent for the convection in this area while humidity takes a passive role, and suggests the areal distribution of sources of sensible and latent heat may be poorly correlated in the city. In a separate study over the same city, Hildebrand and Ackerman (1984) obtained profiles of turbulent fluxes through the depth of the mixed layer during days with fair weather cumulus cloud. Contrary to rural values they found the $\overline{w'q'}$ profile to increase with height and show considerable scatter. This was attributed to increased entrainment at the capping inversion due to greater turbulence over the city. Inspection of their time series (Ackerman and Hildebrand 1978) reveals similar structures to those shown in Fig. 7. A schematic of the suggested boundary layer dynamics is presented in Fig. 8.

6. Implications

This study shows that the ratios of the transfer efficiencies $-r_{wT}/r_{uw}$, $-r_{wq}/r_{uw}$, and r_{wT}/r_{wq} over suburban terrain are not in agreement with those from rural reference data. Because of the absence of appropriate data (information on gradients), the respective eddy diffusivities (K) could not be evaluated. However, since the transfer efficiencies and eddy diffusivities are related to each other and both are measures of the transfer mechanism, it seems appropriate to convey implications with respect to the use of flux-gradient methods to obtain turbulent fluxes in urban meteorology.

In general the aerodynamic method has not been employed to evaluate urban fluxes, but the Bowen ratio–energy balance approach has, especially in Vancouver (Kalanda et al. 1980; Oke and McCaughey 1983; Oke and Cleugh 1987; Grimmond and Oke 1991) and Bonn (Kerschgens and Hacker 1985). In these studies the sensible and latent heat fluxes are evaluated using the gradient Bowen ratio β_G, based on measurements of

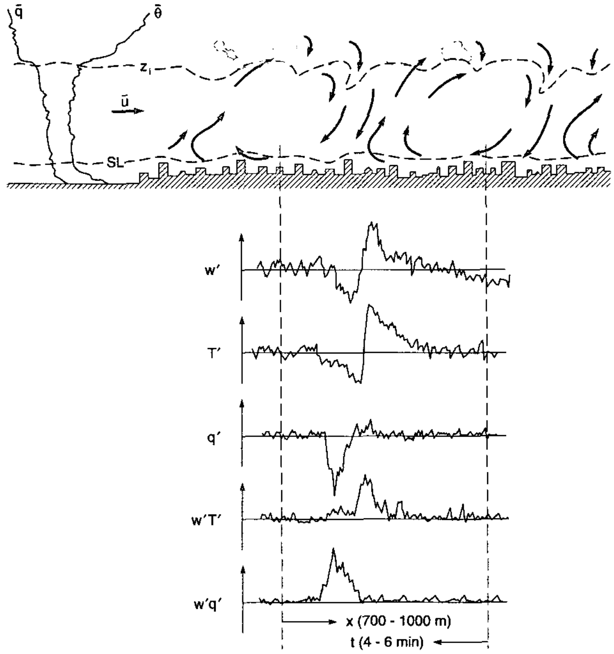

FIG. 8. Conceptual depiction of boundary layer dynamics over a city during cloudy conditions. Shown is the interaction of large eddies with the entrainment layer at boundary layer height z_i and associated entrainment of dry air at the downwind edge of clouds, which results in the characteristics observed in the time series measured within the surface layer (SL) (see text and Fig. 7). Typical mean profiles of temperature ($\overline{T}$) and humidity ($\overline{q}$) are also shown.

vertical differences of temperature and humidity. Evaporation is also sometimes obtained using the Bowen ratio in combination with eddy correlation values of the sensible heat flux density:

$$Q_E = Q_H/\beta_G \qquad (8)$$

(Oke and Cleugh 1987; Grimmond et al. 1991).

The results of the present study put the validity of such estimates in doubt because the derivation of β_G assumes similarity of transfer, and Fig. 1c indicates that the transfers are not equal.

Problems in estimating β_G may also arise because of the large vertical separation of sensors necessary in the urban surface layer. This is the outcome of the very efficient mixing found over the rough city. A large height interval is therefore required in order to obtain sufficient temperature and humidity differences to keep measurement errors small (Kalanda et al. 1980). This often means that at least the lowest sensor may have been located in the roughness sublayer. For example, in the present study the fluxes of sensible and latent heat were noted to increase toward the surface in the daytime (Fig. 6).

In the lower portion of the surface layer the flow is influenced by individual surface elements and is three-dimensional. Thus, point measurements may not be representative of the mean surface-layer fluxes, and it becomes important to know the size and location of the source area contributing to the flux. A relationship between the spatial distribution of sources/sinks at the surface and the measured signal at some height has been developed by Schmid and Oke (1990).

Using a version of the model for scalar fluxes (FSAM) (Schmid 1994) the 50% source areas (i.e., the area enclosing 50% of the surface influence) for an instrument at $z' = 19$ m were calculated. The dimensions, describing the shape of the source ellipsoid, are given in Table 1 together with the necessary input variables for five cases. The results show that as instability increases, the size of the 50% source area decreases and moves closer toward the measuring point. Thus, with increasing instability the sources/sinks in the immediate vicinity of the suburban site become more important. We also note that the area of the 50% source area (A_r) is small in all cases and probably does not include a representative sample of the surface characteristics of a complete suburban landscape. This result supports our suggestion that the observations reported herein are probably influenced by the small-scale inhomogeneities of the surface characteristics. This effect would be more pronounced at the lower level; however, the model was not used for this case because it would be outside its range of applicability.

Schmid et al. (1991) show that as the source area of a flux increases (e.g., by increasing the sensor height) so does the spatial averaging of the flux contributions, and the measurements become more spatially representative and temporally less variable. Also, the changing fetch provided by different wind directions results in additional averaging (in the present study runs were accepted only if the approach flow came from a particular 90° sector). Even though horizontal variability diminishes with increasing height, the present study shows that the turbulent exchange near the surface (e.g., of moisture) may still not be homogeneous because it is also affected by large-scale convective structures that inject the properties of the mixed layer and mixed-layer top into the surface layer.

7. Conclusions

The present study provides additional support to a better understanding of the complex nature of turbulent fields in urban environments. The similarity of transfer efficiencies in the surface layer is investigated, and the results indicate that the behavior of the city atmosphere is unlike that of rural reference surfaces. Ratios involving the momentum flux are relatively smaller than reference sites near neutral and at small instabilities because of bluff-body effects. The ratio of the diffusivities for heat and water vapor is greater than unity at all stabilities. This is probably due to the inherent inhomogeneity of the urban surface, which presents a spatially uneven and complex distribution of sources and sinks of heat and moisture to the lower atmosphere. It is also likely due to the greatly enhanced coupling between the urban surface and mixed layers, which injects the effects of mixed-layer convective structures, and the secondary influence of cloud dynamics, down to the surface where they play a role in determining the surface flux field.

Inequality of the transfers of heat and water vapor means that the Bowen ratio method of obtaining turbulent fluxes over the city is open to error. This suggests, as pointed out by Mulhearn and Finnigan (1978), that to obtain reliable flux estimates over very rough terrain such as a city it is preferable to use eddy correlation methods. In addition, if only point measurements are available, it is important to ensure sufficient averaging over different directions of fetch in order to obtain a spatially representative sample. Considering the likely distortion of mean gradients, it follows that the gradient Richardson number is not entirely suitable as a measure of atmospheric stability in the urban atmosphere.

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TABLE 1. Input parameters and dimensions of the 50% turbulent source area calculated using FSAM (Schmid 1994) and evaluated at $z' = 19$ m. Here z_0—roughness length (0.52 m at the Sunset site); a—near end; e—far end; d—maximum lateral half-width; x_d—upwind distance of d; and A_r—surface area of ellipse.

z'/z_0	z'/L_v	σ_v/u_*	a [m]	e [m]	d [m]	x_d [m]	A_r [m^2]
37	−0.07	1.66	59.8	165.4	43.9	91.5	7733
37	−0.20	1.89	39.6	110.2	35.2	61.4	4191
37	−0.45	1.97	24.3	70.2	25.6	39.4	1979
37	−0.72	2.50	17.0	50.2	25.2	26.7	1406
37	−0.93	3.16	13.6	40.7	27.0	21.3	1236

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