Study of vertical wind profiles in an urban area with complex terrain (Tehran)

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In this paper monthly trends of vertical wind profiles within and above an urban area with complex topography (Tehran) have been investigated using data from a Sodar, a 100-m and two 28-m towers and some surface stations. It includes the occurrence, evolution, dissipation time, peak time and maximum height for katabatic-anabatic winds due to topography, return flow and urban circulation. Vertical distributions of wind and the heat and momentum fluxes up to 600 m were also considered.

The vertical wind profiles which have diurnal and seasonal variations show southwesterly daily anabatic wind and northeasterly nocturnal katabatic wind. Daily vertical wind profile structure which has two layers and two jets contains a decrease approximately at 300–400 m which may be due to the return flow and a daily thermal convective flow in the urban convergent terrain. At night the nocturnal wind profile structure, in the majority of months, has two layers and sometimes three layers containing northeasterly flows. In two layers structure, a decrease can be seen between two layers. In three layers structure, the middle layer has a different direction that indicates the return flow and urban circulation more clearly. Katabatic flows also develop in successive phases varying from 1 to 3 phases in different months. Investigation on surface wind demonstrates that in cold months a drainage flow from a valley located in the west of the station can affect wind speed and direction especially at evening transition time. These flow patterns can influence the way air pollutants disperse in this area.

1. Introduction

Local mesoscale air flows such as sea breezes, mountain, valley winds (slope-winds) and the urban heat-island circulations partly control the local weather and pollutant transport in many parts of the world. This may be quite significant in cities situated in valleys during stagnant weather periods, and may lead to severe air quality problems as the valley may prevent horizontal ventilation. In an urban area the local slope winds when combined with the urban heat-island circulation may re-circulate the urban air, especially during

winter and night-time, when vertical diffusion is small (Atkinson 1981). So study of airflows in complex terrain has been particularly intense over the past fifteen or so years. A major component of this activity has been observations that are required for proper understanding of airflow over an urban area with complex topography.

Slope flows which play an important role in the diurnal circulation of complex terrain regions caused by the thermal forcing on topographic surfaces, belong to the well-known class of winds known as ‘thermal circulation’. Thermal circulations span a range of scales, from a few kilometers

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scale such as flow over a simple slope to mesoscale diurnal circulation systems with horizontal scales on the order of 100 km or more (Bossert and Cotton 1994). Thermal circulation in complex terrain has been studied by some investigators, principally using field experiments. There is no lack of examples of theoretical, numerical and experimental studies of slope flows in the literature. Theoretical (Prandtl 1952; McNider and Pielke 1981; Doran and Horst 1983; Arritt and Pielke 1986; Doran *et al* 1990), numerical (Yamada 1989; Tripoli and Cotton 1989) and laboratory-scale (Meroney 1990) studies have their own merits and drawbacks. A variety of factors affecting slope wind development, include valley wind systems, ambient winds, surface heterogeneities and the radiation budget (e.g., Drobinski *et al* 2003a, 2003b). The basic mechanism that generates slope flow is horizontal temperature difference between air adjacent to a mountain slope and the ambient air at the same altitude over the center of the valley. Heating during the day and radiative cooling at night near the surface contribute to this temperature difference (for a review see Whiteman 2000). Slope flows are characterized by sustained wind speeds, typically 1–5 m/s while the transition between slope flows in the early morning and in the late afternoon is characterized by periods of very low winds, 0.5 m/s (Papadopoulos and Helmis 1999).

Nocturnal airflows and thermal circulations in valleys have also been widely observed, such as those reported by Banta and Cotton (1981); Whiteman (1982); Clements *et al* (1989); Banta *et al* (1995); Triantafyllou *et al* (1995) and Whiteman *et al* (1996). It is not surprising, given that katabatic winds are stably stratified flows with small vertical exchange coefficients. Many nocturnal observations have shown velocity and temperature fluctuations during katabatic flow conditions in a variety of locations (Horst and Doran 1986, 1988; Helmis and Papadopoulos 1995).

Another wind class that belongs to the thermal circulation which has an important effect on urban wind pattern is the heat-island circulation (Oke 1982). Such thermally driven wind systems have been reported for different cities (e.g., Graz city in southeastern edge of the Alps: Lazar and Podesser (1999), Bochum city in Germany: Barlag and Kuttler (1990/91); Toronto city: Findlay and Hirt (1969); London city: Lee (1977)). Because such low wind reduces dispersion and transport, it contributes to serious air pollution problems (Whiteman 1990). Hence, diffusion and air quality studies in urban areas focus on the movement and diffusion of tracer substances or pollutants. It is clear that the main application for the study of turbulence and the effects of local wind systems and modelling of atmospheric conditions especially

in urban area or mountainous terrain is air quality prediction. Similarly preparing knowledge of basic features of the governing flows and the thermal inversion conditions for local authorities is very essential and has led to several experimental campaigns in order to analyze topographic flows by modelling approach in recent years as a powerful method.

Other than these wide investigations all over the world, a few scholarly works have been done for the city of Tehran and are often in local language. For example, Bidokhti and Noroozi (2004) showed that katabatic wind oscillation has a period of about 40 min and vertical mode of low frequency internal wave generated by katabatic wind in a stratified environment might have created shear layers. The metropolitan area of Tehran acts as an important man-made source of pollutants that creates a poor air quality condition. On some days, the pollution loading of the atmosphere is so high that the impressive Alborz Ranges become invisible from most vantage points. It has been recently stated by Tehran's Clean Air Committee that 10,000 people die every year due to air pollution related cardio-pulmonary disease. Shafie-Pour and Ardestani (2007) estimated that the total health damage from air pollution in major urban centres of Iran cost 7 billion US\$; equivalent to 8.4% of Gross Domestic Productivity (GDP); clearly making air quality a serious concern for this country.

The aim of this paper is to present an in-depth analysis of mesoscale wind structure (horizontally and vertically) for Tehran with its topographical situations and its climatology. The area is described in section 2, its database network in section 3, result of observational wind pattern is discussed in section 4, and finally, concluding remarks are given in section 5 which may be applicable to the cities with similar settings.

2. Area characteristics

Tehran, the capital of Iran (35°42'N, 51°25'E), has an area of 700 km² and is situated in a semi-enclosed basin in south of Alborz Mountain range. The southern sloped down to the low lying flat terrain (figure 1a, b). The mountainous nature of Tehran has a dominating control on low level wind climatology. The location for a major city is unusual since it is not near a river or even close to the sea. The average annual rainfall is approximately 230 mm, with most precipitation falling in autumn and winter months. Due to high elevation (approximately 1140 m), aridity and latitude, the city experiences four seasons. Climate can be extremely hot in summer (with

Figure 1. Map of Tehran Metropolitan Region (upper and middle part), districts of Tehran showing the measuring stations (lower part). Numbers denote (1) Tehransar, (2) Aghdasieh, (3) Geophysics, (4) Fatemi, (5) Bazar and (6) Resalat, respectively.

mid-day temperatures ranging between 30° and 40°C), and cold in winter when night time temperatures can be below the freezing point. Local precipitation is absent for 6 months of the year

on the low lying areas. Therefore aside from the summer when large scale flow is easterly, a westerly direction is preferred in other times but is thought to be modified by the mountains. In this study

Figure 2. Estimated roughness at Geophysics station according to the wind directions.

the stability parameter (z/L , $z = 15$ m) range varies from -0.1 to 1 . Hence Tehran experiences a more stable weather condition compared to that of Moriwaki *et al*'s (2002) results for Tokyo (-0.2 , 0.2) and Haizhen *et al*'s (2006) results (-0.1 , 0.1) for idealized simulations in numerical experiments on the east coast of Vancouver Island. The static stability may be affected by mountainous surroundings of the city. Another characteristic is its inhomogeneous surface that gives large value of surface roughness in various directions (figure 2). As figure 2 shows, easterly wind experiences the most roughness ($z_0 + d = 10.8$ m) compared to southeasterly wind ($z_0 + d = 5.8$ m) which shows the complexity of the surface area of the city. Since mixing height (MH) is an important parameter for meteorological research especially for air pollution,

it varies for Tehran from 454 m in January to 1820 m in June.

In this semi-arid environment, the climatic conditions regularly produce fine, cloudless weather conditions with weak winds which can approximately be considered ideal for being affected by local mountains. This is especially true during summer when depression disturbances are almost nonexistent.

3. Data description

Our investigation is based on observations made over a 12-month period using two 28-m towers, a Sodar (Sonic Detection and Ranging), a 100-m tower and four surface stations.

The Doppler Sodar used is a PA1 Sodar that is located at the Geophysics Institute of Tehran University (figure 1). The Sodar worked almost continuously from January to December of 2006. Each Sodar cycle is defined by different pulse sequence which is sent up in uni-beam directions and its vertical resolution is 30 m. During this study the data were obtained from 15 m to around 600 m. Sodar measurements include three wind components, their standard deviation, and momentum fluxes which were stored every 15 minutes.

To find knowledge of hourly surface wind, data from a 2D sonic anemometer measures wind speed, direction and temperature at Geophysics station for whole of 2006 were used. Also data from other surface stations (northern: Aghdasieh, central: Fatemi and low-lying: Bazar) and 2 towers (in upstream: Tehransar and in downstream: Resalat) were also used to demonstrate the existence of monthly temperature gradients over Tehran (figure 1).

Table 1. Days with calm weather in each month for 2006 that were used in this study.

Month	Days	Total samples	Selected samples
January	1,2,3,4,5,6,7,10,11,12,13,18,19,20,22,23,24,25,26,28,29,31	58908	41960
February	1,2,3,5,6,7,8,10,11,12,13,18,19,20,21,22,23,24,25,26,27,28	54720	43200
March	1,5,6,7,8,9,12,13,14,15,22,23,26,29,31	41920	23720
April	1,5,6,9,10,11,14,15,19,20,21,22,28,29,30	57060	28780
May	1,2,3,4,5,6,7,8,9,10,11,12,13,16,17,18,24,25,26,27,28,30,31	53720	40000
June	1,2,3,4,5,6,7,8,9,10,11,12,13,14,15,16,17,18,19,20,21	52200	34940
July	13,18,16,20,21,22,23	59500	13440
August	13,14,15,16,17,18,19,20,21,22,24,25,26,27,28,29,30,31	59480	35540
September	2,3,4,6,7,8,9,10,11,12,13,14,17,18,19,20,21,22,23,24,25,28,29,30	48540	40080
October	1,2,3,4,5,9,10,11,12,13,14,15,16,17,18,19,20,21,22,23,24,25,26,27,28,29,30,31	50980	45220
November	1,2,3,4,5,6,7,8,9,10,11,12,13,16,17,19,20,21,22,23,24,25,26,27,28,29,30	55480	49720
December	1,2,3,4,5,6,7,8,9,10,11,12,13,14,15,16,17,18,19,20,21,22,23,25,26,29,30,31	36840	36840

For the heat flux investigations, data from a 100-m tower which is located near the Institute of Geophysics in a uniform low-storied residential area lying approximately 150 m south of Sodar site was used. Four 2D sonic anemometers installed on the tower at 2, 10, 45 and 100 m that measure wind speed, direction, temperature and humidity.

In this paper, data processing contains two steps including the criteria as follows:

- Step 1: Fine, cloudless, weather conditions with weak wind ($u < 5$ m/s at 15 m), which can be approximately considered ideal for being affected by mesoscale circulations. Derived knowledge from this step demonstrates that background wind has a dominant influence on mesoscale flow and sometimes it overwhelms (figures not shown). To exclude the effects of mesoscale flow which is mainly associated with ambient flow, the second step for data processing was implemented.
- Step 2: In this step two criteria were added to step 1. At the first, data from days when

synoptic scale systems are active over the station were excluded (table 1). Also Pielke (2002) mentioned (see 4.11) any variable can be expressed as follows:

$$\bar{\varphi} = \varphi_0 + \varphi', \quad (1)$$

where $\bar{\varphi}$ denotes any dependent variable, φ_0 is an averaged variable representing the synoptic scale atmospheric conditions value and φ' shows the mesoscale deviations from this large scale. Influence of background flow by subtracting the monthly average (φ_0) from the data was removed. Consequently as we expected the effects of mesoscale flow are clearly appeared in the results.

3.1 Descriptions of measured variables

It is known that the wind profile in an urban area does not follow a logarithmic profile in the surface layer (Rotach 1995) and is usually divided into an inertial sub-layer and a roughness sub-layer (RSL). In the vertical, RSL extends from the surface up to a level at which horizontal homogeneity of the flow is achieved. So knowledge of flow

Figure 3. Monthly average of hourly Sodar wind (u,v) profile (cm/s) and wind speed contour for (a) January, (b) February, and (c) March 2006, for local time. Contours interval is 30 cm/s for speeds lower than 100 cm/s and 50 cm/s for speeds larger than 100 cm/s.

characteristics within the RSL where most of the air pollution problems occur is necessary in urban areas. It is noticeable that obtaining knowledge of urban boundary layer over Tehran, that has a mountainous nature and high separated building with undefined average heights with dominating control on low level wind, is really difficult. To show the importance of mesoscale circulation, detailed analysis of wind data were performed on an annual and monthly basis. In this work at first we consider the vertical wind profile at the Geophysics station situated in the northern half of Tehran (south of Alborz Mountain) to show the mesoscale features of wind profile. Figures 3–6 show monthly average of horizontal wind profile ($U(u, v)$) for 24 hours for days listed in table 1 (January, February and March in figure 3), (April, May and June in figure 4), (July, August and September in figure 5) and (October, November and December in figure 6). In addition to obtain more information on mesoscale flow, the time evolution of monthly average of vertical component of wind ($w(\text{cm/s})$) again for days listed in table 1 were also plotted in figure 7 (for first six months of 2006) and in figure 8 (for second six months of 2006).

To characterize the mean atmospheric circulation at the surface, data from a 2D sonic anemometer were also analyzed. It is obvious that in urban area, the combination of thermal and dynamical factors affect the wind structure. As Zawar-Reza *et al* (2010) showed, one of the features that likely plays a significant role in the low level wind field and has associated control on particulate pollution in Tehran is the heat-island effect. In order to show the urban heat-island effect over Tehran wind circulations, monthly average of temperature for six surface stations in different parts of Tehran were calculated (figure 9). It should be mentioned that these stations have different elevations from sea level. The highest elevation belongs to Aghdasieh and the lowest one is for the southern station (Bazar). The largest temperature difference is between Geophysics and Bazar stations because of heat-island effect and also difference in elevation from sea level. Temperature of Fatemi and Tehransar that have approximately the same height from the sea level are different by $0.2^\circ\text{--}1.2^\circ\text{C}$ (figure 9). This implies that the thermal urban circulation may have an important effect on wind structure.

Figure 4. Similar to figure 3, but for (a) April, (b) May, and (c) June.

Figure 5. Similar to figure 3, but for (a) July, (b) August, and (c) September.

The most important aim in this research is to investigate wind pattern over the Tehran urban area with complex topography. In daily variations as Whiteman *et al* (1999) recognized, there are four regimes in diurnal cycle in this mountainous terrain: night time, morning transition, day time and evening transition. The results indicate that the mesoscale feature of wind structure over Tehran is bi-modal due to katabatic, anabatic, their return flows and urban circulation (figures 3–8). Hence, we divided the wind profile study periods to daily, nocturnal classes and transition categories which will be discussed respectively.

4. Results and discussion

4.1 Daily wind profile

The anabatic wind appears as one of the most frequent phenomenon over Tehran which has two layers and contains two jets (figures 3–6). Daily wind direction in these layers does not change with height and time, only its speed varies and has a minimum at around 300–400 m height. In particular, from figures 7 and 8 it can be seen that

magnitude of vertical wind component (w (cm/s)) containing seasonal variations present a decrease in middle height (around 300 m). Wind (w) profiles in first 100 m above the ground after 12:00 LST varies from 10 cm/s in November to 50 cm/s in May. In upper levels ($z > 500$ m) w varies from 10 to 40 cm/s respectively for December and March.

As Haizhen *et al* (2006) indicated, the decrease in wind speed is also associated with convergence areas having scales around 2–3 km. They showed that the flows during the midday, even in simulations, might result from convective thermal overturning. So the decrease in the horizontal wind speed and in w profiles may be as a result of convergence zone of urban circulation as much as overturning (due to return flow).

The next interesting events which are depicted in daily wind profile are two visible jets, at about 200 and 550 m above the surface (figures 3–6). It is also noticeable that the jets locations and their strengths have seasonal variations. A similar low-level jet profile characteristic was observed by Oettl *et al* (2001) for Graz city at southeastern edge of Alps Mountain. Figures 3–6 show that jet speed at around 200 m varies from 150 cm/s in January

Figure 6. Similar to figure 3, but for (a) October, (b) November, and (c) December.

to 300 cm/s in April. The minimum height of the lower jet is about 150 m in January and its maximum height is about 250 m in May. The second jet at higher than 500 m with southwesterly direction has a maximum value of 300 cm/s in June and August. Its position, like the first jet, also changes in different seasons.

Figures 3–8 illustrate that daily wind profile may be divided into two similar layers as a result of urban circulation and return flow.

Near sunset the wind patterns show a quick transition in wind direction and a reduction of wind speed similar to Haizhen *et al* (2006) results. When the day time thermal convection ceases over land, the down slope flow commences to form circulations which are much shallower and weaker than the daytime circulations.

4.2 Nocturnal wind profile

Here we do not discuss the force that causes down slope flows, but consider its seasonal characteristics like speed, depth and evolution. As figures 3–6 reveal, near the sunset, elimination of surface heating allows a very quick transition (in about 1 hour)

from upslope to down slope flows. This rapid shift just after sunset could cause strong variations and a change in sign of the terms involving horizontal and vertical heat flux convergences.

Figures 7 and 8 show that the nocturnal wind over Tehran is light and is dominated by the down slope flow. Skillingstad (2003) showed that slope flows typically occupy only a very shallow layer; they can cover a broad region of sloping terrain. The reason for substantially elevated katabatic flow up to around 600 m in this region refers to Alborz bare mountain in the northern sector of the station with height of about 3900 m. The best feature of the two-layer nocturnal wind profile is captured in summer (figures 3–6). Haizhen *et al* (2006) indicated that for Vancouver this flow extends to almost 200 m high with the maximum values of 6 m/s near the ground and a jet around 100 m above the ground. In this study lower part of down slope flows reach winds with $w = -46$ cm/s in April and $|u| = 250$ cm/s in June and $w = -8$ cm/s in October and $|u| = 30$ cm/s in January.

The results demonstrate that after transition time wind structure starts to form a two or sometimes three-layer structures over the station. It is

Figure 7. Monthly average of hourly vertical component of wind (w (cm/s)) for (a) January, (b) February, (c) March, (d) April, (e) May, and (f) June of 2006. Contours interval for negative values is 10 cm/s from minimum value and for positive value -10 cm/s from maximum value.

noticeable that the difference between two structures confines to middle layer that in three layers, one has different direction but in two layers, one appears only as a decrease in wind speed without any change in directions. Haizhen *et al* (2006) showed that in dense canopies this upslope flows are because of daily anabatic flows that are still dominant early at night. Monti *et al* (2002) found a 100–300 m high flow with speed of 0.5 m/s for the reverse flow. In this study if the second layer is visible it has a depth of around 250–400 m above ground and a value less than 30 cm/s.

The first layer forms at the beginning of the night with a small depth, which is contemporaneous to the depth increase of the inversion layer. This layer contains the katabatic wind flow from northeast. Figures 3–6 indicate that the horizontal wind speed at around 300–400 m decreases in some hours that the mean wind speed is low. Direction of the decreased wind is different (e.g., figure 1c at 01:00–04:00 LST and 19:00 LST) that may be due to the prevailing return flow. Figures 7 and 8 indicate the vertical structure of the wind profile,

specially the depth and strength of the katabatic wind flows.

In all months the down slope flow usually develops in one phase except in December (3 phases), January (3 phases), February (3 phases) and May (2 phases). In aforementioned months, the first phase formed before sunset and lasts for about 1–2 hours. Sometimes the down slope flows did not reach the lower levels (January and February). After a time gap, the second phase of katabatic flow is formed. With temperature decrease, the rest of upslope flow that formed in the last day disappears and the third phase commences to form. Similar structure has been seen for Athens (Flocast *et al* 1998). Figures 3–8 also demonstrate two maximum centers for katabatic flow related to two jets. The jet position in the first layer is about at 100–150 m, but its strength varies from 60 cm/s in December to 300 cm/s in July. However w (cm/s) associated with this jet also reach different values for each month.

Monti *et al* (2002) reported thickness of ~ 100 m and speed of 3.5 m/s for the layer containing the

Figure 8. Similar to figure 6 but for (a) July, (b) August, (c) September, (d) October, (e) November, and (f) December of 2006.

Figure 9. Monthly average of temperature ($^{\circ}\text{C}$) for four surface stations (Geophysics, Fatemi and Bazar, Aghdasieh) and two towers (Resalat and Tehransar).

katabatic flow for late evening in mountainous terrain. He also used Nappo and Rao (1987) relations namely:

$$h \cong 0.073(\sin \alpha)^{2/3} s, \quad (2)$$

$$u \cong 2.5(\sin \alpha)^{2/9} B_0^{1/3} s^{1/3}, \quad (3)$$

where h is the katabatic flow depth, α is the slope angle, s is the along-slope distance, u is the katabatic flow velocity and B_0 is the stabilizing buoyancy flux, for estimating the speed and height of katabatic flow. Here for this terrain with about 13.5° slope angle and 5000 m from the mountain top, in neutral condition the constant varies from 0.073 to 0.29 in equation (2).

This difference may show the effects of urban circulation and complex topography over mesoscale wind regime in this area.

4.3 Morning and evening transitions

Another wind regime defined for complex terrain is the wind pattern in transition times when

it switches from upslope to down slope or *vice versa*. Particularly for both to appear, the influence of the katabatic flow on the surface wind structure and to investigate wind pattern at transition times, data from a 2D sonic anemometer were used (figures 10 and 11). Here the transitions are analyzed in terms of wind fields and temperature only in one area (Geophysics Institute) using monthly average of surface horizontal wind and temperature for 12 months of 2006.

The time variation of the temperature suggests that the anabatic flow is marked by a rapid heating of air, results from radiative heating (figures 10 and 11). This process takes 1–2 hours. At night a rapid cooling is the main reason for down slope flow that causes katabatic flow. It is worth noticing that near the surface, forming of a reverse temperature gradient due to compressional warming, decreases the speed of katabatic flow compared to the upper levels. Also the rate of cooling decreases substantially that leads to stabilizations and small temperature increase. This process may lead to the formation of down slope flow in different phases.

Figure 10. Monthly average of temperature, wind speed and wind direction for surface data in Geophysics station for (a) January, (b) February, (c) March, (d) April (e) May, and (f) June of 2006. Dashed red lines denote wind direction, black lines present wind speed and dotted blue lines illustrate temperature.

Figure 11. Similar to figure 10 but for (a) July, (b) August, (c) September, (d) October, (e) November, and (f) December of 2006.

After the morning transition time to noon, during the calm days of the whole year, wind direction is from southeastern sector and from noon to evening transition time, it is confined to the south-westerly direction (similar to anabatic wind, seen in the observations). Observed wind directions in figures 10 and 11 show that in the majority of months, anabatic wind from southern to south-western sector reaches the stations. Near the sunset and sun rise when wind switches from anabatic to katabatic or *vice versa*, two peaks appear in the wind direction as the most evident cases in these figures. Evening peak in cold months (from October to May) has considerably different behaviour compared to other months. It seems that the reason for this may be due to a topographic channeling of wind. This flow commence from a valley located in the west of the station. The channeling wind in cold months gets strong and deviates the wind direction from that of the upper level wind.

One interesting point in figures 10 and 11 is that the surface wind speed is more than upper level wind speed captured by Sodar. This paradox is also due to aforementioned drainage flow. (This channel wind was investigated in a local research work and was not published in international journals.) Existence of this drainage flow and its height shows

the importance of terrain topography on wind profile over Tehran.

4.4 Heat and momentum fluxes

Surface fluxes are important because they are used as inputs and boundary conditions in the advanced air pollution dispersion models especially for urban air quality assessments. Undoubtedly, sensible surface heat flux Q_c near the ground and radiation divergence over the katabatic layer (ΔR) affect the buoyancy flux, speed and thickness of katabatic layer. For example, the buoyancy flux of this layer is given by Manins and Sawford (1979a) and Nappo and Rao (1987) as:

$$B_0 = \left(\frac{g}{\rho c_p \bar{\theta}} \right) (\Delta R - Q_c). \quad (4)$$

Thus one of our interests in this research is to have some estimations of surface fluxes. To calculate heat flux, the data from the 100 m tower positioned 150 m away from Sodar station was used. We used the gradient relation:

$$\overline{\theta'w'} = \frac{-K_h \partial \bar{\theta}}{\partial z}, \quad K_h = 0.81 \text{ m}^2/\text{s}. \quad (5)$$

Figure 12. Monthly average of sensible heat flux (captured from a 100 m tower near Geophysics station) at 70 m from surface for first 8 months of 2006.

Figure 13. Storage term as a ratio of for sensible heat flux at 70 m to sensible heat flux at 6 m according to the stability parameter.

Figure 14. Momentum fluxes against wind gradients for u component (a) and v component (b) for three levels including 15–45 m, 45–75 m and 75–105 m.

Here K_h was calculated using Prandtl number (Businger *et al* 1971). Small values of turbulent thermal diffusivity (K_h) are one of the characteristics of this area. It shows that how urban area

and also topography can decrease eddy size over the terrain.

Field study for eight months started from January 2006 shows that heat flux lies mainly in

range -1.3 at night to 7.4 (W/m^2) in day at 70 m height (figure 12).

Also heat flux at 6 m varies from -103 at night to 42 (W/m^2) in day. The values at 6 m are large compared to that of Galinski and Thomson (1995) results that cover -50 to 0 (W/m^2) for night time. Grimmond and Oke (2002) obtained maximum value of 200 – 300 (W/m^2) for summer time sensible heat flux for Vancouver.

To test height dependency of heat flux, ratios of fluxes measured at two heights (70 and 6 m) according to the atmospheric stability parameter (z/L , $z = 10$ m and L is the Obukhov length) were calculated and plotted. In the surface layer, the vertical fluxes should be constant if we ignore the storage term as a ratio of heat fluxes at two different heights (Haizhen *et al* 2006). Figure 13 shows that this term over Tehran decreases with stability parameter and has a linear behaviour for stable and unstable stratifications. Moriwaki *et al* (2002) indicate that the storage term in Tokyo for unstable condition equals to around 1 between 21 and 29 m. Judging from these results, existence of surface layer with height dependent heat flux in this residential area is evident. This dependency is because of complex three dimensional source/sink distributions which lead to spatial and temporal variations in wind profile and temperature.

In addition, the momentum flux and nondimensional mean wind profile ($\varphi_m = -(kz/u_*)\partial\bar{u}/\partial z$, $\overline{u'w'} = K_m\partial\bar{u}/\partial z$) were calculated for this period over Geophysics station. Nondimensional wind gradient (φ_m) was less than 0.75 for unstable and 6.4 for stable conditions (figures not shown). These values are larger than Al-Jiboori *et al's* (2002) results for Beijing.

Also the monthly average of hourly momentum fluxes for whole of 2006 were estimated (figures not shown) to show the differences between down slope and upslope flows. The derived momentum flux has a seasonal and a diurnal variation. Sodar results show that in daytime it varies from -110 (cm^2/s^2) in summer to -20 (cm^2/s^2) in winter. The nocturnal momentum fluxes also vary from -30 (cm^2/s^2) in summer to 0 (cm^2/s^2) in winter. Another importance of the momentum flux knowledge is to determine the eddy diffusivity

$$\overline{u'w'} = -K_{m_u}(\partial u/\partial z), \quad (6)$$

$$\overline{v'w'} = -K_{m_v}(\partial v/\partial z). \quad (7)$$

As figure 14 shows the calculated coefficients of turbulent diffusivity (K_m) especially for u component is small. Eddy diffusivity in x direction is half of that in y direction, but both of them increase vertically. This difference indicates the effect of

west-east arrangement of mountain in the north that affects the turbulent structure in this terrain. Their anisotropic values illustrate the importance of field study in this complex terrain especially from the point of view of air pollution.

5. Conclusion

Urban air circulation is a suitable ground for testing mesoscale meteorological and air pollution models over cities. In this paper we have presented results from surface layer measurements taken during 2006 for Tehran area to obtain knowledge of wind profile and heat flux over this urban area with complex topography.

Analysis of wind profile in Tehran shows that the mountainous topography of Tehran has a mesoscale control over wind regime in this region. Basically, three main wind circulations in the vertical wind profile can be identified (down slope–upslope flows, their return flows and thermal circulations due to urban effects). The frequency distribution of wind direction over Tehran is bimodal, nocturnal and daily.

Nocturnal wind profile contains 2 and sometimes 3-layer structures. Both structures are similar in first and last layers. The middle layer in 3-layer structure, appears as a layer with different directions (due to more strong return flow and urban circulations), but 2-layer structure presents only a decrease in wind speed between the two. First and last layers due to down slope flow from Alborz Mountain have a northeasterly direction and also contain a jet at around 250 and 550 m. It is worth noting that at night the katabatic flow phase develops in first phase in warm months to three phase in cold months.

Daily wind profile contains two layers with the same direction and two jets around 150 and 550 m and a decrease at around 300 – 400 m (due to the return flow and urban circulation). The first layer is affected by anabatic flow from southwestern of Tehran.

Analysis of the surface data showed that a drainage flow has large control on the surface wind especially after evening transition. The drainage flow increases wind speed at the surface compared to that from lowest level of the Sodar data and also make sharp variation in the wind directions.

Therefore it is clear that the Tehran topography has a strong influence on the wind pattern horizontally and vertically that certainly is important in air pollution dispersion.

We know that our knowledge of the effects of wind and large sources of anthropogenic heat and air pollution over Tehran is little and there is plenty of scope for work; however, these will come

with adding more equipped stations and testing meteorological and air pollution models.

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