Seasonal variations in the surface energy and CO₂ flux over a high-rise, high-population, residential urban area in the East Asian monsoon region

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Abstract
Using the eddy covariance method, this study reports the one-year turbulent fluxes of momentum, energy, and CO₂, and their seasonal variations over a recently redeveloped high-rise, high-population, residential area in the metropolitan city of Seoul, Korea. The study area is affected by the Asian monsoon, which is accompanied by long rain spells and a related mid-season depression of solar radiation in the summer. Our analysis shows that the urban surface energy balance and turbulence characteristics demonstrate typical urban properties. Unstable conditions dominate all day, and the storage heat flux (night-time and morning) and sensible heat flux (afternoon) significantly affect the diurnal variations in the urban surface energy balance. Owing to the rough urban surface, the turbulence intensities are higher than those reported previously in other cities. The annual CO₂ emission rate is approximately 13.1 kg CO₂ m⁻² year⁻¹ with traffic, which is the major source of CO₂ (+2.3 μmol·m⁻²·s⁻¹ per 100 vehicles). Ecosystem respiration, including that by vegetation, soil, and humans, becomes dominant in the night-time (00:00–05:00), thus contributing significantly to the annual CO₂ budget. Further analysis indicates a unique coupling of urban surface energy partitioning and CO₂ emission rates with the seasonal progression of the Asian monsoon: (a) surface albedo has annual minima in late summer when the sun elevation angle, is relatively higher and the urban surface condition is wetter than in other seasons; (b) the Bowen ratio ranges from 1.7 (summer) to 7.0 (winter); and (c) CO₂ emission rates show seasonal variations with the progress of the summer monsoon.

KEYWORDS
Asian monsoon, CO₂ emission, eddy covariance method, high-rise residential area, Seoul, surface energy balance, urban climate

1 | INTRODUCTION
Recently, interest has grown regarding the effects of urbanization on local, regional, and global changes in biogeochemical cycles. With over half of the world’s population living in cities (UN, 2014), anthropogenic activities in cities can threaten societal sustainability by accelerating environmental changes. By fossil fuel
com bustion and cement production, about 10 billion tons of carbon in CO₂ (Ballantyne et al., 2012) and equivalent waste heat are emitted into the atmosphere annually; the majority of these anthropogenic emissions are closely related to urban metabolisms (Grimm et al., 2008). Recent climate change is a major challenge for cities and surrounding areas as they are vulnerable to climate-induced scarcities of water and food, degradation in air quality, and extreme weather. These challenges are elevated as cities expand and consequently, the interactions between a city and the climate system require immediate attention and study. The long-term monitoring of urban surface fluxes to the atmosphere will enable us to improve the prediction of future weather, climate systems, and their effects on our societies. Direct observations are necessary for developing and evaluating urban-atmosphere exchange models. Furthermore, data-based models of urban-atmosphere interactions can be used by policymakers to establish effective policies for climate mitigation and adaptation.

Monitoring of the exchanges of momentum, energy, and mass at the urban-atmosphere interface has received considerable attention in recent decades. The eddy covariance method is one of the most widely used techniques for the direct measurement of surface fluxes. In the last 30 years the eddy covariance method has developed rapidly, with approximately 50 applications used on the study of urban surfaces worldwide (Grimmond and Christen, 2012). Previous studies have suggested the following commonalities of urban surface energy balance (SEB) and CO₂ emission rates (FC): (a) the sensible heat flux (QH) is dominant over the latent heat flux (QL), (b) the heat storage term (dQs) is important in determining SEB, and (c) the urban surface is a net CO₂ source, and FC is controlled by traffic counts and vegetation areal fractions.

Most urban flux towers are concentrated in developed countries in Europe and North America. Therefore, previous investigations have not assessed urban surface fluxes extensively regarding their component, magnitude, and temporal variation for developing countries in Asia and Africa. Although more than half of the world’s urbanites live in Asian countries, only a few studies have investigated short-term urban SEB and FC in Asian cities (e.g., Tokyo, Japan (Moriwaki and Kanda, 2004); Beijing, China (Liu et al., 2012; Miao et al., 2012; Song and Wang, 2012); Singapore (Velasco et al., 2013; Roth et al., 2017); Shanghai, China (Ao et al., 2016); Osaka, Japan (Ueyama and Ando, 2016; Ando and Ueyama, 2017); and Seoul, Korea (Hong and Hong, 2016; Hong et al., 2019)). The lack of directly observed data from developing countries, especially in the Asian monsoon region, hinders our understanding of the interactions between urban structures and their functions with the environment.

Studying the effects of rapid urban development on the urban microclimate in Asian cities that are influenced by the Asian monsoon is uncommon. Hong and Hong (2016) reported on the effects of residential regeneration on the surface heat environment using long-term direct measurements of QH and radiative fluxes. Following the urban redevelopment into a compact, high-rise residential building, the urban heat island intensified by approximately ~0.6°C, and the fractions of QH, anthropogenic heat emission (QF), and dQs increased (Hong and Hong, 2016). However, without QF, the SEB was inevitably uncertain, and the FC over urban areas was left for future study. In addition, most Asian cities are influenced by the East Asian monsoon system. The Asian monsoon is characterized by heavy precipitation events and tropical cyclones (i.e., typhoons) during the monsoon period, which affects natural ecosystems and their water cycles (e.g., Kwon et al., 2010; Hong and Kim, 2011; Hong et al., 2014). Several studies reported that QF was important in the SEB during the summer monsoon season in Asian cities (Moriwaki and Kanda, 2004; Liu et al., 2012; Miao et al., 2012; Ando and Ueyama, 2017). In previous urban flux measurement efforts, the role of the Asian monsoon in urban microclimates and FC was not extensively investigated. The monsoon period coincides with the main summer growing season for vegetation and prior studies have suggested the importance of urban vegetation as a regulator for energy and water cycles over urban ecosystems. However, little attention has been paid to the role of urban vegetation on FC in this unique region.

With the above background, the main objective of this study is to report the one-year turbulent exchanges of energy and CO₂ over a high-rise, high-population residential area in Seoul, Korea, by focusing on their seasonal variations and controlling factors. We highlight the impacts of the Asian summer monsoon on the urban SEB. After the methodology and site description are presented (Section 2), the flux measurement results are reported with random error characteristics (Section 3). We investigate temporal variabilities of surface fluxes by classifying the observation data into seasons (December–February, winter; March–May, spring; June–August, summer; and September–November, autumn), working days (weekdays, not holidays) and non-working days (weekends and holidays), and daytime (>0 W m⁻² of net radiation) and night-time periods.

2 | MATERIALS AND METHODS

2.1 | Urban surface energy–CO₂ balances

For the ideal volume of the urban ecosystem, SEBs could be expressed as the following equation (Oke, 1987):
\[ Q^* + Q_F = Q_H + Q_E + dQ_S + dQ_A \ (W \cdot m^{-2}) \]  

(1)

where \( Q^* \) is the net radiation estimated by the sum of incoming and outgoing, or reflected, short- and longwave radiation (\( K_1, K_2, L_1, \) and \( L_2 \)):

\[ Q^* = K_1 - K_2 + L_1 - L_2 \ (W \cdot m^{-2}) \]  

(2)

This study presents direct observations of surface albedo (\( \alpha = K_2 / K_1 \)) by a net radiometer, analysed with several potential issues to understand the neighbourhood-scale \( \alpha \) and its seasonal variability.

Similar to previous studies, this study estimates \( dQ_S \) as the residual (\( RES = Q^* - Q_H - Q_E \)). This approximation can include \( Q_F \), measurement errors, and footprint mismatches. The annual mean \( Q_F \) is approximately 20 W m\(^{-2}\) from an inventory estimation (Lee and Kim, 2015). However, since the temporal variation of \( Q_F \) is a remaining question, the \( Q_F \) is considered as a part of \( RES \).

The urban CO\(_2\) budget equation is expressed as (Feigenwinter et al., 2012):

\[ FC + dS = C + RE - P \ (\mu mol \cdot m^{-2} \cdot s^{-1}) \]  

(3)

Here, \( dS, C, RE, \) and \( P \) are the concentration change of CO\(_2\) in the control volume, CO\(_2\) emission from fossil fuel combustion, respiration by soil, vegetation, and humans, and CO\(_2\) uptake by photosynthesis, respectively. \( dS \) can be neglected by the stationary assumption of the eddy covariance method; therefore, the observed \( FC \) is the well-mixed sum of \( C, RE, \) and \( P \), representing the neighbourhood spatial–temporal scale in the upwind direction.

**FIGURE 1** (a) The location of the flux tower with urban land cover map from GRUMP version 1. (b) Aerial photograph around the flux tower (●) and the location of the traffic survey (★). (c and d) Photographs show landscape surrounding the flux tower for the prevailing wind direction.
2.2 Site characterization

The flux tower is located in the north-western part of Seoul, Korea (Figure 1; 37.6350°N, 126.9287°E). The district around the tower was redeveloped in 2009 following the New-Town Plan, a representative housing regeneration policy in Seoul and its satellite cities. The total population density around the site was approximately 15,000 people km
². For the week of 3–9 November 2014, traffic counts were conducted on a major road near the flux tower (star in Figure 1b), reaching 11,835 and 10,834 vehicles day⁻¹ on weekdays and weekends, respectively.

Figure 2 shows urban morphological information estimated from the high-resolution LiDAR data and indicated directional heterogeneity around the site. The prevailing wind comes from the high-rise building complex (i.e., the south to west direction, 135–315°), and impervious surfaces such as buildings and roads occupy more than 60% of land cover on the relatively flat terrain for the prevailing wind. Areas to the northwest and northeast are forested areas with a small fraction of buildings and small building heights. Within 200 m from the tower to the southwestern residential area, which is the peak location of the turbulent flux footprint, the mean horizontal building area fraction (λ_P) and mean building height (z_H) are approximately 0.35 and ~18.6 m, and the corresponding zero-plane displacement (z_d) and roughness length (z_0) are ~11.3 and ~1.6 m, respectively. The aspect ratio, defined as the ratio of building height to road width (HWR hereafter), is about 1.0, and the sky-view factor at street level is about 0.2. Therefore, the residential area is classified as a compact high-rise (type 1) based on the Local Climate Zone scheme (Stewart and Oke, 2012). It is also noticeable that the residential area around the station is unique compared to other LCZ1 study sites because its building area fraction is relatively smaller (ranging between 0.15 and 0.50) and its vegetation fraction is relatively larger.

The blending height (z_r), the lower boundary of the inertial sublayer (or constant flux layer), is an important criterion of the local-scale flux measurement. For the given values of λ_P in the residential area, turbulent flow around the tower varies between the skimming flow regime (z_r ~1.0035z_H) and the wake regime (z_r ~2z_H) as a
function of the wind direction, and the blending height ranges from 25 and 40 m in unstable condition to 80 m in stable conditions (Grimmond and Oke, 1999b). This suggests that our flux measurements represent the micro-scale to local-scale in unstable and neutral conditions, but the micro-scale in stable conditions.

The peak location of the turbulent flux footprint is approximately 200 m from the tower under unstable conditions. Approximately 86% and 72% of the flux footprint corresponds to the high-rise building complex in unstable and neutral conditions within a 500 m radius of the tower, respectively. Footprint areas extend to larger areas in stable conditions (Figure 3c), thus including more vegetative surfaces in nocturnal fluxes in this prevailing wind direction. However, in general, the night-time flux footprint still corresponds to LCZ1 because strongly unstable to neutral conditions account for about 70% of the flux footprint, even at midnight (Figure 3a). Further, there is no substantial seasonal change in wind direction, and the magnitude of night-time fluxes is smaller than that of daytime fluxes as a result of the effects of high-rise apartment buildings and traffic roads. Our analysis focuses on the urban area from the south to west direction (180–270°); however, our overall conclusions in terms of seasonality and controlling factors of surface fluxes do not change even when the analysis is extended to all wind directions, despite approximately 20% change in the annual CO₂ fluxes.

### 2.3 Climate condition

Korea has four distinct seasons – a hot-humid summer and a cold-dry winter, with mild transition seasons of spring and autumn. The climatological means (1981–2010) for air temperature and precipitation in Seoul are 12.5°C (−2.4°C in January; 25.7°C in August) and 1,450 mm-year⁻¹, respectively. In general, more than 50% of annual precipitation (892 mm in the 30-year average) is concentrated in the summer season, coinciding with the East Asian monsoon. These climate conditions are different from those of other Asian cities with reported surface fluxes, particularly regarding (a) intervals of heavy rain in summer, (b) larger seasonal variations in air temperature, and (c) significant depression of solar radiation in the summer monsoon season.

During the study period (March 2015–February 2016), the climate condition in Korea was abnormally hot and dry. In Seoul, the mean air temperature, accumulated precipitation and solar duration (the period $K_1 > 120$ W·m⁻²) were 13.3°C (+0.8°C from 30-year climatology), 806.7 mm (−643.9 mm), and 2,598.3 hr (+532.4 hr), respectively (Table 1). Unusually, rain fronts remained in the south of the Korean peninsula during Changma, the intensive heavy rainfall period of June 25th to July 29th. The total rainfall recorded in Seoul was only 221.4 mm, which was 145 mm less than the 30-year average.

![Figure 3](image-url)
climatology. The beginning of autumn (September 2015) was warm and dry but gradually became humid with southerly winds from the Philippine Sea. Compared with 30-year statistics, the precipitation (solar duration) increased (decreased) by 52.1 mm (44.2 hr) in November 2015, unlike the summer season. However, the mean air temperature was 1.7°C higher than normal, probably because the reduction of longwave cooling by cloud cover was dominant. During the winter season, relative warmth was maintained until late January 2016, with a strong El Niño and positive Arctic Oscillation. Then, a strong cold wave moved southward from the Arctic region, and cold weather conditions lasted until the end of February 2016.

2.4 Instrumentation and data process

2.4.1 Instrumentation

The 10 m high lattice tower was built on the rooftop of a 20 m building. A three-dimensional sonic anemometer and a closed-path infrared gas analyser (IRGA) (CPEC200, Campbell Scientific Inc., USA) were installed at the top of the tower for turbulent flux measurements with a 10-Hz sampling rate. Closed-path IRGA is advantageous in having a better retrieval rate for monsoon-affected areas, characterized by intense rain during the Asian summer monsoon and dusty conditions in the spring (Choi et al., 2004; Dias et al., 2009; Hong et al., 2014). Q* was computed by the 10-min averages of $K_1$, $K_l$, $L_1$, and $L$ as measured by a net radiometer (CNR4, Kipp & Zonen, Netherlands). All data were recorded using a data-logger (CR3000, Campbell Scientific, USA). At least every 3 months, the IRGA was calibrated with a standard CO₂ gas of 401.1 ppm (CRM No. 112–01-018, Korea Research Institute of Standards and Science) and N₂ gas for zero calibration. The measurement system was checked daily via teleconnection, and data were retrieved every 2 weeks on field trips.

2.4.2 Data processing and flux footprint estimation

Turbulent fluxes were computed by EddyPro software (version 6.2.0, Li-COR, USA) with a 30-min averaging period. The magnetic declination angle (8°15'W on May 1, 2015, changing by 0°03'W year⁻¹) was considered to calculate the wind direction. Double rotation, spike removal (Vickers and Mahrt, 1997), and spectral correction (Moncrieff et al., 2004; Horst and Lenschow, 2009; Fratini et al., 2012) were applied. To filter out the influence of micro-scale anthropogenic activities and to increase turbulent fluxes representativeness of the local-scale source area, an outlier filter was applied to the 10-Hz raw data and post-process (Figure 4; Papale et al., 2006; Hong et al., 2009; Kotthaus and Grimmond, 2012). After quality control, 98.0% of radiative fluxes, 92.4% of $Q_{18}$, 89.5% of $Q_{19}$, and 93.7% of $F_C$ data were available for analysis. For convenience, this study uses LST (local standard time), which is 9 hr ahead of UTC (universal time coordinated).

The stability parameter, $\zeta$, was computed to quantify atmospheric stability as follows:

$$\zeta = \frac{(z_m - z_d)}{L} = \left(\frac{z_m - z_d}{-\frac{u^2}{k g w^2} \theta_v}\right),$$

where $L$, $u_\ast$, $w$, $\theta_v$, $k$, and $g$ are the Obukhov length, friction velocity, vertical wind speed, virtual temperature replaced with sonic temperature, von Kármán constant, and gravitational acceleration, respectively.
(= 0.4), and gravitational acceleration (= 9.821 m s\(^{-2}\)), respectively. An overbar denotes the temporal mean, while the prime symbol denotes a perturbation from the mean. The estimated \( \zeta \) is divided into four categories: stable \((\zeta > 0.1)\), near-neutral \((0.1 > \zeta > -0.1)\), unstable \((-0.1 > \zeta > -0.5)\), and strongly unstable \((-0.5 > \zeta)\). The normalized SDs of \( w \) (\( \Phi_w \)) and temperature (\( \Phi_T \)) were tested to check the data quality and the turbulence characteristics:

\[
\Phi_w = \frac{\sigma_w}{T_*} = \sigma_w / (-w' T'_s / u'_s) = C_1 \cdot [1 - C_2 \{ (z_m - z_d) / L \}]^{-1 / 3},
\]

(5)

\[
\Phi_T = \frac{\sigma_w}{u_*} = -C_3 \cdot [C_4 - (z_m - z_d) / L]^{1 / 3}.
\]

(6)

Characterizing the random error (\( \varepsilon \)) of the measured fluxes is essential for model validation, parameter optimization, and estimating statistical confidence in the measured fluxes. This study quantifies the total random uncertainty in the measured turbulent fluxes and its covariant properties with meteorological variations by employing the daily differencing approach (Hollinger and Richardson, 2005; Richardson et al., 2006). If a measurement flux \( (x) \) pair of two successive days \( (x_1 = F + \varepsilon_1, x_2 = F + \varepsilon_2, \) and \( F \) is the true flux) are under equivalent environmental conditions, the SD of random error (\( \sigma(\varepsilon) \)) can be written as:

\[
\sigma(\varepsilon) = \sigma(x_1 - x_2) / \sqrt{2}.
\]

(7)

For this daily differencing method, similar environmental conditions are assumed for 24-hr differences in \( K_z \) within 100 W m\(^{-2}\), air temperature within 2°C, and wind speed within 0.5 m s\(^{-1}\) in the same wind direction sector \((135°-315°)\). More information on the random flux error estimation is found in Hollinger and Richardson (2005).

To quantify representative source areas of turbulent and radiative fluxes, the turbulent flux footprint and radiative flux footprint are calculated using the method of Hsieh et al. (2000) and Lambert’s cosine law (Schm, 1997), respectively. By Lambert’s cosine law, the 99% source area of the net radiometer is a circle with a radius of ~130 m (Schmid, 1997) and a good representation of the surrounding residential environment. The flux footprint model by Hsieh et al. (2000) is an analytical model based on the combination of dimensional analysis and Lagrangian stochastic dispersion model in the inertial sublayer. The simple analytical footprint model is useful to provide a rule-of-thumb estimate of flux source areas but is only valid in the inertial sublayer than the roughness sublayer (RSL) in general. Turbulence in the RSL showed a coherent turbulence structure with vigorous mixing by the strong inflection of horizontal mean wind speed near the canopy top, and the flux footprint probability distribution is contracted in the RSL as compared to that in the inertial sublayer (Raupach et al., 1996; Baldocchi, 1997). Accordingly, it is likely that our footprint climatology includes wider areas around the flux tower than the actual areas.

3 | RESULTS

3.1 | Diurnal variation of wind speed and atmospheric stability

Figure 5 shows the climate conditions observed at the site throughout the study period. The daily mean wind speed is ~1 m s\(^{-1}\), with higher values in spring and summer than in autumn and winter (Figure 5d). The mean diurnal courses of wind direction show the local circulation of the typical warm and cold breezes induced by thermal contrast around the site. During the daytime, the wind comes from the urban area (180°–270°; SSW and WSW), which is warmer than the forested area; the prevailing wind direction gradually changes to the north (315°–45°; NNW and NNE) after sunset. Near-zero wind is observed during these daily transition periods. Throughout the year, this local breeze due to the thermal contrast around the tower dominates, except in rainy conditions. During
rainy conditions, the observed wind direction follows the synoptic weather condition rather than local circulation, probably due to the weakened thermal gradient by the depression of radiation and increased evaporative cooling of vegetation in the summer growing season. The overall seasonal change in the local breeze is small, but the pattern occurs more frequently in summer.

The atmosphere is mainly unstable with a strong southerly wind (Figure 3a) and the fraction of stable conditions is relatively small. In this study, 56.5% of data reflect unstable conditions (27.8% for strong instability, 28.7% for moderate instability), while stable and near-neutral conditions comprise 19.4% and 24.1%, respectively. The fraction of unstable conditions during the night-time (19:00–07:00) is 40.9%, which is larger than the night-time stable condition of 32.5%. This indicates that the nocturnal urban boundary layer remains unstable because of $dQ_S$ and $Q_F$, as reported in previous studies (e.g., Christen and Vogt, 2004).

### 3.2 Surface roughness characteristics

Normalized SDs ($\sigma_i/U$ and $\sigma_i/u_*$ where $i = u, v, w$, $U$ is the mean wind speed) are necessary for various purposes in the micrometeorology, including quality control of the observed data and surface flux estimations. They are useful to investigate the effects of the RSL on turbulent fluxes and flow distortion particularly. The normalized turbulence statistics show directional variation and indicate evidence of the flow distortion in the case of wind from the northeast and northwest (Figure 6). Flow distortion of mean streamline is possible in relatively low measurement height ($z_m/z_D < 2$), making larger vertical azimuth angle (rotation angle) of wind. Our data shows a sinusoidal variation of the mean vertical azimuth angle (rotation angle) with wind direction (Figure 6h). This sinusoidal variation is interpreted as a linear increase of slope from the south to the north (Wilczak et al., 2001; Yuan et al., 2011), which can be removed by the coordinate rotation. The mean vertical azimuth angle (rotation angle) of wind is $6.5 \pm 0.1^\circ$ (mean $\pm SE$) from SSW to WSW. This angle is comparable with previous urban studies (Nemitz et al., 2002; Vesala et al., 2008). However, it is notable that the flow distortion can be a potential error at approximately $220^\circ$ because of its relatively large vertical azimuth angle.

The mixing layer analogy of turbulence in the RSL indicated that $\sigma_u/u_*$ and $\sigma_w/u_*$ are relatively smaller, thus giving a larger value of $-u'^2/\sigma_u\sigma_v$ in the RSL compared to the inertial sublayer (Raupach et al., 1996; Roth, 2000). For the SSW and WSW wind directions, $\sigma_u/u_*$, $\sigma_v/u_*$, and $\sigma_w/u_*$ are $1.91 \pm 0.45$, $1.88 \pm 0.33$, and $1.35 \pm 0.20$, respectively, in near-neutral conditions. These integral turbulence characteristics (ITC) values at the site are comparable with those in the inertial sublayer reported in natural vegetative canopies (Raupach et al., 1996; Hong et al., 2002) and wind tunnel but $\sigma_u/u_*$ and $\sigma_w/u_*$ correspond to upper and lower bound of urban canopies (Roth, 2000). Importantly, the ITC values were relatively smaller in the WSW, thus indicating that our observation could not capture the local-scale source areas in this direction.

The turbulence intensities (i.e., $\sigma_i/U$, $\sigma_i/\Delta U$, and $\sigma_i/\Delta U$) are $0.54 \pm 0.18$ (mean $\pm SD$), $0.55 \pm 0.18$, and $0.39 \pm 0.12$ in near-neutral conditions, respectively. The ratio of $u_*$ to $U$ is $0.24 \pm 0.01$, corresponding to a drag coefficient of 0.058 (Figure 7a). This value of $u_*/U$ is larger than those

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**Figure 5** Climate conditions for March 2015–February 2016: (a) air temperature ($T_{air}$; unit: °C) with the 30-year (1981–2010) climatology values (solid line: mean, dashed lines: minimum and maximum), (b) 30-min and accumulated precipitation ($Pr$; unit: mm; solid line: study period, dashed line: climatology value), (c) vapour pressure deficit (VPD; unit: hPa), and (d) mean wind speed ($U$; unit: m s$^{-1}$). Grey and black dots indicate 30-min and daily mean values, respectively.
proposed by Roth (2000) (see Figure 1 in Roth, 2000). This discrepancy may arise from the difference in the definition of mean wind speed because the mean wind vector, $U (\sqrt{u^2+v^2+w^2})$, used in this study is generally smaller than the mean wind speed, $(\sqrt{u^2+v^2+w^2})$ (Wilson et al., 1982; Hong et al., 2002). Another possibility arises from the sharp decline in turbulence intensity with height due to a lack of data for more rugged urban surfaces, such as high-rise buildings, like those in this study, and for $z_{mf}/z_{ff}$ of <1.5 in previous studies.

Figure 7 illustrates the drag coefficient, $\Phi_w$, and $\Phi_T$, as functions of $\zeta$. The ruggedness of high-rise urban structures induces significantly higher turbulence intensity ($u_*/U$) and surface drag coefficients relative to those from previous studies in other urban areas. The similarity constants $C_1$, $C_2$, $C_3$, and $C_4$ in Equations (5) and (6) are determined as $C_1 = 1.16$, $C_2 = 4.11$, $C_3 = 1.10$, and $C_4 = 0.12$. Previous studies reported similar values in the ranges of 1.12–1.30, 2.09–3.00, 0.95–1.14, and 0.030–0.085 for $C_1$, $C_2$, $C_3$, and $C_4$, respectively (e.g., Tillman, 1972; Panofsky et al., 1977; de Bruin et al., 1993; Kaimal and Finnigan, 1994; Roth, 2000; Toda and Sugita, 2003; Choi et al., 2004; Hong et al., 2004).

### 3.3 | Random flux error

For the natural ecosystem, random errors in flux measurements have leptokurtic double-exponential (Laplacian) distributions rather than normal (Gaussian) distributions.

**Figure 6** Directional characteristics of (a)–(g) turbulence characteristics and (h) rotation angle ($\psi$): (a) drag coefficient ($C_D^{0.5} = u_*/U$), (b)–(d) turbulent intensity ($\sigma_i$) normalized by the mean wind speed ($U$) and (e)–(g) normalized by the friction velocity ($u_*$).
Previous studies also reported that the random flux error was heteroscedastic, implying that the SD of the random error ($\sigma(\varepsilon)$) increased linearly with increasing measured flux magnitude. Our results suggest that the statistical properties (mean, SD, skewness, and kurtosis) of the random flux error in the urban residential area are similar to those in natural environments (Figure 8 and Table 2). The random flux error shows a peaked distribution rather than a normal distribution (Figures 8a–c, and kurtosis in Table 2). Slopes (0.11, 0.31, and 0.24 for $Q_H$, $Q_E$, and $F_C$) and offsets (15.16 and 8.80 for $Q_H$ and $Q_E$) of the random flux error are similar to those reported for forest canopies (slope: 0.16, 0.23, and 0.63; offset: 19.7, 15.3, 19.7, 15.3).

**FIGURE 7** (a) $u^*$ vs. $U$ as a box plot (box: median, 25th, and 75th; whisker: 10th and 90th; symbol: 5th and 95th percentiles) and normalized SDs of (b) vertical velocity $w$ and (c) temperature as a function of $-\zeta$ ($= -(z_m - z_d)/L$). Only data from the apartment area (180–270°) are used.

**FIGURE 8** The frequency distribution of the random flux error ($\varepsilon$) for (a) $Q_H$, (b) $Q_E$, and (c) $F_C$, and (d)–(f) the scaling of each random flux error ($\sigma(\varepsilon)$) with the flux magnitude. Only data from the south direction (135–315°) are used.
and 0.62) and grassland (slope: 0.07, 0.16, and 0.30; offset: 17.3, 8.1, and 0.38) (Richardson et al., 2006), excepting the offset of $F_C$ (3.97). The seasonality of $\sigma(e)$ is proportional to the magnitude of the measured flux; however, the relative error $(\sigma(e)/F)$ in Table 2 indicates relatively smaller seasonality. The $\sigma(e)/F$ of $Q_H$ (0.28) is lower than those of $Q_E$ (0.83) and $F_C$ (0.80). We speculate that this larger $\sigma(e)/F$ in $Q_E$ and $F_C$ are related to error propagation by the two instruments used (a sonic anemometer and IRGA), while $Q_H$ is measured only using a sonic anemometer. In addition, the source and sink distributions of $Q_H$ are relatively more homogeneous than those for $Q_E$ and $F_C$ (e.g., buildings, soil, and vegetation), which induces greater sensitivity to meteorological conditions.

The relatively larger offset of the random error of $F_C$ (i.e., $\sigma(e)$ at $F_C = 0$) at the site seems to be related to the assumptions applied in the daily differencing approach. Indeed, zero $F_C$ appears at least twice a day in a natural ecosystem (i.e., around sunrise and sunset), but such conditions do not usually occur in high-population urban areas. Instead, irregular changes in anthropogenic emissions, such as traffic volume, occur during equivalent meteorological conditions, violating the assumption in the $\sigma(e)$ estimation. In this respect, our result requires careful interpretation as an upper limit for random flux error $C_22$ ($Q_E$) while the two instruments used (a sonic anemometer and IRGA), which induces greater sensitivity to meteorological conditions. The daily differencing method is practically applicable, but previous studies found that the daily differencing method generally overestimated the $\sigma(e)$ relative to the independent two-tower method and the model residual method, because it is difficult to ensure identical climate conditions (Richardson et al., 2006; Dragoni et al., 2007). Notably, traffic amounts can cause increases in $\sigma(e)$ of $F_C$ in urban areas because the traffic volume is not entirely dependent on meteorological conditions, even at the same time of day.

### 3.4 | Surface radiative balance

#### 3.4.1 | Radiative fluxes and albedo

Figures 9a–c show seasonal changes in the diurnal variability of surface radiative fluxes. The daily peak times of $K_1$, $K_1$, and $Q^*$ are around noon, but $L_1$ and $L_2$ reach daily maxima at 14:00 local time when the surface temperature reaches its daily maximum. In the daytime, the variability of $Q^*$ is determined by $K_1$. The magnitude of nighttime $Q^*$ depends on the longwave cooling.

The sun elevation angle reaches its annual maximum in summer, but the annual maximum of $K_1$ does not appear in summer, and the longwave cooling rate is minimized during the surface monsoon season because of the lengthy rainy spells (Hong and Kim, 2011; Hong and Hong, 2016). The $K_1$ gradually decreases from spring to winter, and the mean longwave cooling rates are approximately −78, −56, −62, and −65 W·m$^{-2}$ in spring, summer, autumn, and winter, respectively (Table 3).

#### TABLE 2

| Season | $n$ | $F$ | $\bar{e}$ | $\sigma(e)$ | $(\sigma(e)/F)_r$ | Skew | Kurt | Scaling at $F \geq 0$ |
|--------|-----|-----|-----|-----|--------|-----|-----|---------------------|
| $Q_H$  | Total | 882 | 59.7 | 0.9 | 16.4 | 0.28 | 0.3 | 5.8 | $\sigma(e) = 15.16 + 0.11 \times Q_H$ $r^2 = 0.84$ |
|        | Spring | 256 | 91.9 | 1.0 | 20.2 | 0.22 |       |       |        |
|        | Summer | 227 | 45.6 | 0.2 | 14.4 | 0.32 |       |       |        |
|        | Autumn | 269 | 46.2 | 1.9 | 12.8 | 0.28 |       |       |        |
|        | Winter | 130 | 48.7 | 1.0 | 15.9 | 0.33 |       |       |        |
| $Q_E$  | Total | 838 | 21.9 | 1.4 | 18.2 | 0.83 | 0.8 | 7.4 | $\sigma(e) = 8.80 + 0.31 \times Q_E$ $r^2 = 0.92$ |
|        | Spring | 239 | 25.0 | 1.2 | 18.9 | 0.76 |       |       |        |
|        | Summer | 223 | 30.3 | 1.7 | 23.7 | 0.78 |       |       |        |
|        | Autumn | 251 | 18.7 | 1.7 | 16.0 | 0.86 |       |       |        |
|        | Winter | 125 | 7.6  | 0.7 | 5.4  | 0.71 |       |       |        |
| $F_C$  | Total | 928 | 7.9  | 0.02 | 6.3  | 0.80 | −0.2 | 4.4 | $\sigma(e) = 3.97 + 0.24 \times F_C$ $r^2 = 0.69$ |
|        | Spring | 270 | 8.7  | 0.3  | 6.7  | 0.77 |       |       |        |
|        | Summer | 246 | 7.0  | 0.04 | 5.8  | 0.83 |       |       |        |
|        | Autumn | 267 | 7.0  | 0.2  | 6.4  | 0.91 |       |       |        |
|        | Winter | 145 | 9.9  | −0.8 | 6.5  | 0.66 |       |       |        |

Notes: Only data from the south direction (135–315°) are used.
No reports exist regarding the seasonal variation in surface albedo ($\alpha$) in urban areas. Typical $\alpha$ in vegetative areas show its annual maximum in the growing summer season, corresponding to the annual maximum of the leaf area index. This is because increased solar radiation is reflected, except for photosynthetically active radiation
(e.g., Brest and Goward, 1987; Song, 1999; Rechid et al., 2009). However, the observed $\alpha$ in urban areas is lower in late summer and higher in winter (Figure 10a). Urban $\alpha$ is strongly affected by several factors, including the sun elevation angle, surface wetness, vegetation phenology, land cover fraction, urban structure, and materials. Among these factors, the land cover fraction and urban structure do not change with the season. Phenology cannot explain our observed variations, as per other studies in natural forest canopies, and snow is also of secondary importance as there were only 3 days of snow. Alternatively, we note two possible reasons for the observed seasonal changes in $\alpha$: the effects of moisture on building surfaces and soil, and the sun elevation angle on $\alpha$.

Typically, $\alpha$ decreases with an increasing sun elevation angle, and the decline is particularly sharp between zero and $30^\circ$ (White et al., 1978; Aida and Gotoh, 1982; Christen and Vogt, 2004; Grimmond et al., 2004; Balogun et al., 2009; Bergeron and Strachan, 2012; Kotthaus and Grimmond, 2014). Our data show the same dependency of $\alpha$ on a sun elevation angle (Figure 11). The mean sun elevation angle in daytime is smaller than 30$^\circ$ in the fall and winter; consequently, $\alpha$ in these two seasons is larger compared to that in summer.

Furthermore, our scrutiny reveals that seasonal changes in surface moisture conditions with the seasonal progression of the East Asian monsoon are related to $\alpha$ seasonality. Figure 10c shows $\alpha$ with accumulated solar radiation after precipitation events as a measure of surface moisture. This provides information on the surface moisture and it is practically infeasible to retrieve surface moisture in urban areas. Some studies have used the elapsed time after precipitation events (e.g., Kotthaus and Grimmond, 2014). In this study, the accumulated solar radiation is used instead of the elapsed time because the latter depends on the timing of the beginning and end of

| TABLE 3 | | | |
|---|---|---|---|
| | Spring | Summer | Autumn | Winter |
| $K_1$ | All-day | 333.0 ± 323.8 | 283.5 ± 293.0 | 215.1 ± 246.2 | 203.2 ± 207.6 |
| | Daytime | 508.7 ± 274.1 | 434.7 ± 261.5 | 379.0 ± 217.9 | 335.1 ± 203.2 |
| | Night-time | 13.7 ± 28.6 | 8.6 ± 19.8 | 8.4 ± 22.4 | 10.6 ± 24.3 |
| $K_1$ | All-day | −62.6 ± 59.8 | −51.4 ± 52.6 | −39.5 ± 44.8 | −39.6 ± 40.2 |
| | Daytime | −95.3 ± 50.2 | −78.6 ± 46.9 | −69.2 ± 39.8 | −65.0 ± 33.4 |
| | Night-time | −3.3 ± 6.2 | −2.0 ± 4.1 | −2.1 ± 5.1 | −2.6 ± 5.7 |
| $L_1$ | All-day | 314.6 ± 49.8 | 406.4 ± 29.2 | 336.6 ± 38.6 | 259.9 ± 39.6 |
| | Daytime | 319.8 ± 45.9 | 407.3 ± 29.2 | 337.8 ± 41.1 | 250.2 ± 38.1 |
| | Night-time | 305.1 ± 55.1 | 404.9 ± 29.2 | 334.9 ± 35.3 | 251.9 ± 41.8 |
| $L_1$ | All-day | −407.3 ± 53.1 | −472.1 ± 31.6 | −406.6 ± 44.0 | −325.4 ± 23.2 |
| | Daytime | −426.0 ± 49.5 | −486.7 ± 28.4 | −421.8 ± 46.3 | −333.6 ± 21.4 |
| | Night-time | −373.4 ± 41.5 | −445.7 ± 16.1 | −387.5 ± 31.9 | −315.3 ± 20.5 |
| $Q^*$ | All-day | 177.6 ± 234.4 | 166.4 ± 212.1 | 105.5 ± 176.8 | 89.1 ± 148.9 |
| | Daytime | 307.2 ± 193.2 | 276.7 ± 187.5 | 225.9 ± 150.7 | 186.8 ± 115.4 |
| | Night-time | −57.9 ± 31.3 | −34.3 ± 21.3 | −46.2 ± 27.4 | −53.6 ± 28.4 |
| $Q_H$ | All-day | 103.8 ± 97.3 | 75.1 ± 79.9 | 50.9 ± 67.4 | 59.3 ± 65.9 |
| | Daytime | 150.1 ± 90.2 | 109.4 ± 79.1 | 89.6 ± 66.2 | 91.0 ± 63.7 |
| | Night-time | 19.7 ± 30.4 | 12.7 ± 23.5 | 2.0 ± 21.8 | 12.9 ± 33.7 |
| $Q_E$ | All-day | 27.5 ± 30.4 | 43.8 ± 45.7 | 21.4 ± 26.8 | 8.5 ± 10.6 |
| | Daytime | 37.6 ± 31.7 | 60.8 ± 46.7 | 32.7 ± 28.5 | 10.7 ± 8.5 |
| | Night-time | 9.0 ± 15.8 | 12.8 ± 20.5 | 7.1 ± 15.5 | 4.7 ± 9.3 |
| Residual | (= $Q^* - Q_H - Q_E$) | All-day | 46.3 ± 146.5 | 47.5 ± 129.7 | 33.2 ± 111.5 | 21.4 ± 102.2 |
| | Daytime | 119.5 ± 131.0 | 106.4 ± 124.6 | 103.5 ± 99.7 | 84.8 ± 78.0 |
| | Night-time | −86.7 ± 43.4 | −59.8 ± 36.8 | −55.4 ± 38.4 | −71.3 ± 48.5 |

Notes: Only data observed from apartment areas (180°-270°) are used.
a precipitation event. $\alpha$ clearly increases with increased accumulated $K_i$ after precipitation events, showing an equilibrium value of 0.185 after one clear day (i.e., 20 MJ m$^{-2}$ of accumulated $K_i$). Precipitation during the summer monsoon season accounts for approximately half of the total annual precipitation at 397.9 mm and 49.3%; this wet summer period promotes smaller $\alpha$ in summer compared to that in other seasons, as mentioned above, consistent with the observed seasonal variation in $\alpha$. Our results highlight that this urban canyon wetting feedback with $\alpha$ significantly affects the surface radiative balance and SEB in Asian cities influenced by monsoon systems.
3.4.2 | Footprint of a radiometer

The footprint mismatch between a hemispheric radiometer and the turbulent flux footprint is important in this analysis. Figure 11 shows the observed \( \alpha \) with sun elevation angles and the modelled dependency of \( \alpha \) on the aspect ratio, HWR. The model simulation in Figure 11 is based on the urban canopy radiative transfer scheme proposed by Harman et al. (2004) and Porson et al. (2010). The observed upward radiative fluxes are viewed differently from actual urban structures when measured by a hemispheric radiometer because of radiometer footprint weighting (Adderley et al., 2015). The footprint of a hemispheric radiometer obeys Lambert’s cosine law (Schmid, 1997), and therefore, a building structure near a radiometer has a relatively greater contribution to the radiative flux measurement per unit area than structures far from the instrument. This means that urban canopy structures, such as HWR, embedded in the observed radiative fluxes may differ from the actual urban structure around the flux tower. The HWR value at the study site is 1.0 based on aerial LiDAR imaging around the flux tower (Hong and Hong, 2016). However, HWR weighted by the radiative flux footprint of the hemispheric radiometer is much smaller than 1 from the aerial LiDAR image, indicating that the observed upward radiative fluxes are reflected more than the actual amount coming from inside of the urban canyon. Consequently, the observed \( \alpha \) could be approximately 10% smaller than the actual \( \alpha \), thereby leading to overestimation of \( Q^* \) and the surface energy imbalance. This bias corresponds to \( \sim 2\% \) of \( K_1 \) or 20 \( \text{W m}^{-2} \) at most around noon in clear conditions in spring and summer. The modelled \( \alpha \) is sensitive to HWR, and the model reproduces the observed \( \alpha \) when using the radiative flux footprint-weighted HWR. This indicates that footprint mismatches in the hemispheric radiometer can contribute to uncertainty in the SEB unless it is properly considered in the data analysis and modelling interpretation.

3.5 | Surface energy balance

3.5.1 | Seasonal variations

Urbanization favours \( Q_H \) over \( Q_E \) by enhancing the fraction of impervious surfaces. Like those in other cities, our study site shows \( Q_H^* \) values larger than \( Q_E^* \) throughout the year (Figures 9d–e and Table 3). However, evapotranspiration is not negligible even in the high-rise, high-population residential area because of vegetation between the buildings and the summer monsoon. In addition, \( Q_H \) and \( Q_E \) show different seasonality as the monsoon progresses. \( Q_H \) gradually decreases from the spring to winter season (Figure 9d and Table 3), whereas \( Q_E \) is maximized in the rainy summer monsoon season (Figure 9e and Table 3) and approaches zero in the cold and dry winter season (Figure 9e and Table 1). Consequently, the Bowen ratio \( (\beta = \sum Q_H/\sum Q_E) \) is lower in summer and higher in winter (Figure 12). The seasonal mean daytime \( \beta \) values are approximately 4.0, 1.8, 2.7, and 8.2 for spring, summer, autumn, and winter, respectively. Other Asian cities have not reported exact seasonal means of \( \beta \) but have reported values of approximately 2–4 in Tokyo, Japan (Moriwaki and Kanda, 2004), 2–4 in Shanghai, China (Ao et al., 2016), and 1.5–3 in Osaka, Japan (Ando and Ueyama, 2017). In comparison with other Asian cities, the \( \beta \) at our site demonstrates similar changing trends but a larger seasonal range, attributable to the greater seasonal difference in precipitation at our site. In particular, the wintertime \( \beta \) value is higher than those in other studies because January 2016 was drier and colder than normal. The opposite situation occurred in summer, as summer 2015 was drier than the climate average (Table 1).

3.5.2 | Diurnal variation

Figure 9 shows that the peak times of both turbulent heat fluxes (\( Q_H \) and \( Q_E \)) are delayed (13:30–14:00) relative to the peak of \( Q^* \) around noon. This phase-shift between turbulent heat fluxes and \( Q^* \) was also reported for several other cities in various climate zones (Grimmond and Oke, 1995; Oke et al., 1999; Nemitz et al., 2002; Christen and Vogt, 2004; Grimmond et al., 2004; Lemonsu et al., 2004; Moriwaki and Kanda, 2004; Offerle et al., 2005; Coutts et al., 2007; Balogun et al., 2009; Ramamurthy and Pardyjak, 2011; Kotthaus and Grimmond, 2014). However, the behaviours of this time lag were inconsistent in previous studies. For example, the phase-shift was reported only in \( Q_H \) (Grimmond and Oke, 1999a; Newton et al., 2007; Masson et al., 2008; Ao et al., 2016), only in \( Q_E \) (Grimmond and Oke, 1995; Ando and Ueyama, 2017), or lacking in both \( Q_H \) and \( Q_E \) (Grimmond and Oke, 1995; Christen and Vogt, 2004; Coutts et al., 2007; Frey et al., 2011; Velasco et al., 2011; Goldbach and Kutttler, 2013; Ward et al., 2013).

For the time lag of both turbulent fluxes, the daily peaks of both the vapour pressure deficit (VPD) and the thermal difference between the urban surface and the overlying air, which drives \( Q_H \) and \( Q_E \), appears in the afternoon, about 1–2 hr after the \( Q^* \) maximum. Furthermore, the major sources of \( Q_E \) (e.g., trees, grasses, and soil) are affected by building shadows in the daytime because they are located at road edges under buildings within the urban canyon, while the major sources of \( Q_H \) are mainly distributed on roofs, walls, and roads. The
shadow effect on $Q_E$ may contribute to the observed peak-time delay in $Q_E$. In this study, the two tallest buildings on the site are located at the eastern end, so shadows emerge in the morning inside the urban canopy (not shown here).

The ratios of $Q_H$ and $Q_E$ to $Q^*$ ($Q_H/Q^*$ and $Q_E/Q^*$) show similar diurnal variations with monotonic increases and similar slopes from morning to afternoon (Figure 12b–c), such that $\beta$ is constant and $dQ_S/Q^*$ decreases monotonically throughout the daytime (Figure 12a). This result suggests typical variations in $dQ_S/Q^*$ during the day. In the morning, most $Q^*$ is used to heat buildings and roads, and, in the afternoon, the urban surface is hot enough, so $dQ_S/Q^*$ decreases accordingly. The relationship between the residual and $Q^*$ shows a clockwise hysteresis pattern (Figure 12e), which is typically observed in urban areas (Grimmond and Oke, 1999a; Rigo and Parlow, 2007; Velasco et al., 2011; Ramamurthy et al., 2014; Hong and Hong, 2016). During the night-time, near-zero values of nocturnal turbulent heat fluxes (i.e., unity of $dQ_S/Q^*$) indicates a balance between the long wave radiation and $dQ_S$.

### 3.6 CO$_2$ concentration

The observed diurnal CO$_2$ concentration shows a morning peak and then decreases to the minimum in late afternoon hours (Figure 13a). The peak in the morning is well-matched with the traffic volume around the tower; the smaller values in the afternoon are associated with the higher planetary boundary layer height. Such diurnal patterns have been reported in other cities (Reid and Steyn, 1997; Grimmond et al., 2002; Velasco et al., 2005; Kumar and Nagendra, 2015; Crawford et al., 2016; Schmutz et al., 2016; Roth et al., 2017). During the study period, the mean concentration is 414.8 ppm. The daily amplitude, or the difference between the maximum and minimum, is 18.4 ppm (4.4% of the mean concentration). The background CO$_2$ concentration at Mauna Loa is 401.4 ppm throughout the study period (NOAA, 2018). In the mid-latitudes of the northern hemisphere, CO$_2$ concentrations are generally higher in winter than in other seasons because more fossil fuel combustion is used for heating, while vegetation absorbs less CO$_2$. Similarly, our data show that the seasonal mean CO$_2$ concentration is the highest in winter at 424.3 ppm (with a daily maximum of 524.7 and a 15-day mean of 486.3 ppm) but does not substantially decrease during the summer season (Figure 14). The CO$_2$ concentration increases during the summer monsoon season. We speculate that this summer pattern is associated with the reduced photosynthesis during the depression of the solar radiation concurrent with the summer monsoon season (Kwon et al., 2010; Hong and Kim, 2011). This indicates that the observed CO$_2$ concentration is seasonally controlled by the
FIGURE 13  Seasonal diurnal pattern of (a) the CO₂ concentration (unit: ppm), and that of the CO₂ flux ($F_C$; unit: μmol m$^{-2}$ s$^{-1}$) during (b) all days, (c) working days, and (d) non-working days. Figures show the median, interquartile (box), 5th and 95th percentiles (whisker), and mean values (black dot). Only data from the apartment area (180–270°) are used.

FIGURE 14  Seasonal diurnal pattern of (a) the CO₂ concentration (unit: ppm), and that of the CO₂ flux ($F_C$; unit: μmol m$^{-2}$ s$^{-1}$) during (b) all days, (c) working days, and (d) non-working days. Figures show the median, interquartile (box), 5th and 95th percentiles (whisker), and mean values (black dot). Only data from the apartment area (180–270°) are used.
TABLE 4 Daily (24-hr), daytime (\(Q^* > 0 \text{ W m}^{-2}\)), and night-time mean CO\(_2\) flux (\(F_{C}\); unit: \(\mu\text{mol m}^{-2} \text{s}^{-1}\)) and SD values for all, working, and non-working days

|                  | Spring  | Summer | Autumn | Winter |
|------------------|---------|--------|--------|--------|
|                  | All-day | All-day| All-day| All-day|
|                  | 10.3 ± 7.4 | 7.5 ± 6.1 | 8.9 ± 6.5 | 10.9 ± 7.1 |
|                  | Daytime | Daytime| Daytime| Daytime|
|                  | 11.3 ± 7.2 | 7.3 ± 6.3 | 10.4 ± 6.3 | 11.9 ± 7.2 |
|                  | Night-time | Night-time| Night-time| Night-time|
|                  | 9.0 ± 7.5 | 7.8 ± 5.9 | 7.2 ± 6.4 | 10.0 ± 6.9 |
| Working days     | All-day | All-day| All-day| All-day|
|                  | 11.1 ± 7.6 | 8.1 ± 6.4 | 9.8 ± 6.8 | 11.6 ± 7.3 |
|                  | Daytime | Daytime| Daytime| Daytime|
|                  | 12.4 ± 7.5 | 8.0 ± 6.6 | 11.8 ± 6.3 | 12.8 ± 7.6 |
|                  | Night-time | Night-time| Night-time| Night-time|
|                  | 9.5 ± 7.3 | 8.1 ± 6.0 | 7.7 ± 6.7 | 10.4 ± 6.9 |
| Non-working days | All-day | All-day| All-day| All-day|
|                  | 8.4 ± 6.8 | 6.1 ± 5.3 | 7.1 ± 5.6 | 9.6 ± 6.5 |
|                  | Daytime | Daytime| Daytime| Daytime|
|                  | 9.0 ± 6.0 | 5.6 ± 5.2 | 7.8 ± 5.4 | 10.1 ± 6.1 |
|                  | Night-time | Night-time| Night-time| Night-time|
|                  | 7.5 ± 7.8 | 6.9 ± 5.4 | 6.4 ± 5.6 | 9.2 ± 6.8 |

Notes: Only data observed from apartment areas (180°–270°) are used.

FIGURE 15 (a) Vehicular traffic counts (vehicles per 30-min interval), (b) relationship between the CO\(_2\) flux (\(F_{C}\); unit: \(\mu\text{mol m}^{-2} \text{s}^{-1}\)) and traffic, and (c) the temperature response of nocturnal \(F_{C}\) (00:00–05:00) in bins of 50 data points. The error bars in (b) indicate SE, and the shaded area in (c) indicates the interquartile range.

...intensity and duration of the summer monsoon, as well as by anthropogenic emission (i.e., vehicular traffic) and the evolution of the planetary boundary layer (i.e., dilution and accumulation).

3.7 CO\(_2\) flux

Figure 13b shows the seasonally averaged diurnal variation in CO\(_2\) fluxes (\(F_{C}\)). The seasonal mean \(F_{C}\) is 10.3, 7.5, 8.9, and 10.9 \(\mu\text{mol m}^{-2} \text{s}^{-1}\) (39.1, 28.4, 33.7, and 41.6 g CO\(_2\) m\(^{-2}\) year\(^{-1}\)) for spring, summer, autumn, and winter, respectively (Table 4). The total CO\(_2\) emission in summer (approximately 2.6 kg CO\(_2\) m\(^{-2}\)) is 1.2 kg CO\(_2\) m\(^{-2}\) less than that in winter (3.8 kg CO\(_2\) m\(^{-2}\)). This difference comprises approximately 71% of the annual net ecosystem production in East Asian temperate forests (1,640 ± 630 g CO\(_2\) m\(^{-2}\) year\(^{-1}\); seven sites with annual mean air temperatures of 10–16°C as reported in Kato and Tang, 2008). We speculate that the seasonality of \(F_{C}\) is mainly associated with vegetation phenology, which covers approximately 32% of the land surface around the flux tower, rather than with local heating systems. The district heating system for the apartment complex utilizes hot water transport from a power plant far outside the flux footprint, and thus does not contribute to the observed seasonality of \(F_{C}\).

The summertime \(F_{C}\) of 7.5 \(\mu\text{mol m}^{-2} \text{s}^{-1}\) is comparable with the values obtained from the relationship between the summertime \(F_{C}\) and the plan area fraction of buildings (\(\lambda_{p}\)) reported in other urban areas by Grimmond and Christen (2012) and Christen (2014). The annual CO\(_2\) emission rate at the site is 13.1 kg CO\(_2\) m\(^{-2}\) year\(^{-1}\), which is also comparable with those in other cities reported (Ward et al., 2015). The \(F_{C}\) difference between working and non-working days is similar to the seasonal variation (Figures 13c–d, Table 4). On working days, the mean \(F_{C}\) is 10.1 \(\mu\text{mol m}^{-2} \text{s}^{-1}\) (38.5 g CO\(_2\) m\(^{-2}\) day\(^{-1}\)), which is 29% higher than that on non-working days (7.8 \(\mu\text{mol m}^{-2} \text{s}^{-1}\) = 29.7 g CO\(_2\) m\(^{-2}\) day\(^{-1}\)). The nocturnal \(F_{C}\) is similar on working and non-working days; most of the difference is explained by the reduction in daytime anthropogenic activity from lower traffic volumes on non-working days (Figure 15a).

Figure 15a shows the mean diurnal patterns for vehicular traffic volume (vehicles per 30-min) during the traffic survey period of November 3–9, 2014. In the morning, the...
traffic volume on working days is almost twice that of non-working days. This is consistent with the observed diurnal pattern of $F_C$, indicating a strong dependency of $F_C$ on the traffic volume. Similarly to previous studies from several other cities (Nemitz et al., 2002; Soegaard and Møller-Jensen, 2003; Vesala et al., 2008; Velasco et al., 2009; Järvi et al., 2012), our site shows a linear relationship between the traffic load ($Tr$) and $F_C$ (Figure 15b):

$$F_C = 0.023 \times Tr + 3.5 \text{ (unit: μmol·m}^{-2}·\text{s}^{-1}; r^2 = 0.83)\text{.} \quad (8)$$

The slope of Equation (8) indicates that the observed $F_C$ is increased by approximately 2.3 μmol·m$^{-2}$·s$^{-1}$ for every traffic volume increase of 100 vehicles in 30-min. This slope is consistent with the emission inventory data for road vehicles. In the study area, the vehicular speed limit is 30 km·hr$^{-1}$ as all roads are assigned as children protection zones. Based on the inventory data, the CO2 emission per vehicle per kilometre of travel at 10–30 km·h$^{-1}$ is approximately 300 g CO2 km$^{-1}$ (= 6.82 mol·km$^{-1}$; Kim et al., 2011), which is approximately twice the emission rate at high speeds (>40 km·hr$^{-1}$). Therefore, for an increase in traffic by 100 vehicles, the increase of $F_C$ on the road is 253 μmol·m$^{-2}$·s$^{-1}$ (= 300 g CO2·km$^{-1}$ vehicle$^{-1}$ × 1/44 g·mol$^{-1}$ × 10$^{-3}$ m·km$^{-1}$ × 10$^6$ mol·mol$^{-1}$ × 100/1800 vehicle s$^{-1}$). Considering the fraction of main roads (~10%) and the mean road width (~10 m) in the study site, these inventory data estimate 2.53 μmol·m$^{-2}$·s$^{-1}$ per 100 vehicles, consistent with the observed slope in Equation (8).

The magnitude of the intercept in Equation (8) (i.e., 3.5 μmol·m$^{-2}$·s$^{-1}$) is smaller than that of the city centre of Edinburgh in the UK (11.7 μmol·m$^{-2}$·s$^{-1}$; Nemitz et al., 2002) and larger than that of Helsinki in Finland (0.3–1.1 μmol·m$^{-2}$·s$^{-1}$; Vesala et al., 2008; Järvi et al., 2012). The intercept of the regression represents non-traffic CO2 emissions, such as ecosystem respiration ($RE$) of vegetation, soil, and humans in urban areas. Vegetative ($RE_V$) and human ($RE_H$) respiration are functions of air temperature (Davidson et al., 2006) and population density (Moriwaki and Kanda, 2004; Velasco and Roth, 2010; Ward et al., 2013, 2015), respectively. To quantify the contributions of both respiration sources, we add the constant term ($y_0$) for $RE_H$ to a van’t Hoff-type $RE_V$ function (van’t Hoff, 1898; Davidson et al., 2006; Equation 9a):

$$RE = RE_H + RE_V = y_0 + a \times e^{b \cdot T_{air}}. \quad (9a)$$

The observed $F_C$ from 00:00–05:00 LST is selected to avoid the contribution of traffic emissions because the traffic amount is less than 100 vehicles during this period (Figure 15a). Using this nocturnal $F_C$, the temperature response curve is fitted to Equation (9a) (Figure 15c):

$$RE = 2.33 + 0.025 \times e^{0.18 \cdot T_{air}} \quad (r^2 = 0.61). \quad (9b)$$

The coefficients of $y_0$, $a$, $b$, and the corresponding $Q_{10}$ ($= e^{b \times 10}$) are 2.33, 0.025, 0.18, and 6.0, respectively. The exponential respiration curve reproduces the nocturnal urban respiration well ($r^2 = 0.61$). Notably, the fitted $RE_H$ of 2.33 is comparable with the $RE_H$ value of 2.1 μmol·m$^{-2}$·s$^{-1}$ estimated from the population density (~15,000 people km$^{-2}$) and human metabolism data (2.1 μmol·m$^{-2}$·s$^{-1}$ = 12.4 mol·day$^{-1}$ = 11 times·min$^{-1}$×1,440 min·day$^{-1}$×0.5 l·time$^{-1}$×3.5%×1/22.4 mol·l$^{-1}$) (Prairie and Duarte, 2007; West et al., 2009). This result suggests that the estimation of $RE_H$ is a good rule-of-thumb for CO2 source partitioning in urban areas, despite the simplistic approach and uncertainties (Velasco and Roth, 2010).

Our findings also indicate that $RE$ is significant for $F_C$ from midnight to dawn (00:00–05:00) and that vehicular traffic is the dominant source of CO2 for the rest of the day, at approximately +2.3 μmol·m$^{-2}$·s$^{-1}$ per 100 vehicles. $RE_H$ is likely to remain constant throughout the year; by contrast, $RE_V$ is larger than $RE_H$ in summer (mean $T_{air} > 20^\circ$C) but smaller in winter because of its strong dependency on $T_{air}$. In this study, our statistical and observational data were limited, so additional analysis should be performed using longer collection periods and data with greater detail.

## 4 | SUMMARY AND CONCLUSIONS

Using the eddy covariance method, this study reported the one-year surface fluxes of energy and CO2 to examine the features of urban-atmosphere exchange and the controlling factors over a high-rise high-population residential area in Seoul, Korea. This area is affected by the East Asian monsoon system. The study region consists of multiple identical buildings and has been recently redeveloped to an apartment complex from the compact midrise building with bare soil, and the population has been increasing dramatically following the redevelopment. Approximately 500 m to the southwest, buildings have the same structure and spatial arrays of 24 m building height. Building morphology data analysis shows that the residential areas can be classified as compact high-rise with relatively larger fractions of vegetation and road, and turbulent fluxes range from the micro-scale to the local-scale. To eliminate abrupt contribution of anthropogenic heat and CO2 emission to the corresponding turbulent fluxes, median-based filter is
applied. Random flux error \( (\sigma(\varepsilon)) \) is also quantified using the daily differencing approach, and our analysis suggests that the statistical properties and seasonality of \( \sigma(\varepsilon) \) at urban sites show properties similar to leptokurtic double-exponential distributions, as do previous results from natural ecosystems.

The urban SEB is strongly influenced by the summer monsoon, causing substantial changes in \( \alpha \) and \( \beta \). The observed \( \alpha \) shows the annual minimum (~0.17) and maximum (~0.20) in late summer and winter, respectively. Our investigation strongly suggests that this seasonal variation in \( \alpha \) is associated with changes in the sun elevation angle and soil moisture with the progress of the summer monsoon. Furthermore, the \( \beta \) shows seasonality similar to that of \( \alpha \), ranging from 1.7 in summer to 7.0 in winter. This seasonality of \( \alpha \) and \( \beta \) in urban areas is unique to East Asian urban areas affected by the summer monsoon. Annual mean \( \alpha \) is generally assigned in urban canopy models. This prescription will eventually underestimate (overestimate) \( Q* \) in summer (winter) in urban areas in the Asian monsoon region through the bias of the \( \alpha \) in the models unless the monsoon influence is considered.

Turbulent heat flux maxima are delayed to 13:30–14:00 compared to that of \( Q* \) around noon in their diurnal courses. Our investigation indicates that the VPD and thermal contrast at the urban-atmosphere interface mainly control this observed diurnal variation in turbulent heat fluxes. Notably, the VPD and thermal contrast change as the Asian monsoon progresses. Furthermore, our findings indicate that the heterogeneity between \( Q_{hi} \) and \( Q_E \) contributes to the observed peak-time delay in \( Q_E \). \( dQ_5 \) is important in regulating the partitioning of \( Q* \) into turbulent energy fluxes. In the morning, most of the \( Q* \) is partitioned into \( dQ_5 \), but this fraction steadily decreases during the daytime. Consequently, the peak times of \( Q_{hi} \) and \( Q_E \) are in the afternoon (13:30–14:00), while the peak time of \( dQ_5 \) appears in the morning. Overnight, \( dQ_5 \) is balanced with net longwave radiation, while other components are small.

The observed CO2 concentration shows a bimodal diurnal pattern with rush-hour car emissions and the evolution of the planetary boundary layer. The summer monsoon also clearly affects the seasonal variation in the CO2 concentration and fluxes. The mean annual CO2 concentration is \( \pm 414.8 \pm 9.2 \) ppm. The annual CO2 emission rate is ~13.1 kg CO2 m\(^{-2}\)·year\(^{-1}\). The total seasonal emission shows the distinct role of urban vegetation; the mid-summer depression of CO2 uptake in the monsoon season controls the temporal variability in CO2 emissions. In the summer, CO2 emission is 2.6 kg CO2 m\(^{-2}\), 1.2 kg CO2 m\(^{-2}\) smaller than winter emissions (~3.8 kg CO2 m\(^{-2}\)). The observed \( F_C \) is strongly dependent on vehicular traffic (+2.3 \mu mol m\(^{-2}\)·s\(^{-1}\) per 100 vehicles), while \( RE \) is significant during night-time. \( RE_H \) is estimated as ~2.1 (based on inventory data) to 2.33 \mu mol m\(^{-2}\)·s\(^{-1}\) (based on observed \( F_C \)), and a temperature-dependent equation reproduces the observed nocturnal \( RE_V \) well. The \( RE_V \) shows strong seasonal variation, approaching zero in winter and reaching 4 \mu mol m\(^{-2}\)·s\(^{-1}\) in summer. In the study area, a district heating system is used to heat the buildings, and, therefore, CO2 emissions from local heating are negligible. Also, the difference between working and non-working days is comparable with the seasonal variation in \( F_C \) via the reduction of traffic volume.

Despite a few limitations of the measurement, our study outlines urban-climate feedback in monsoon-affected Asia by reporting several key findings regarding energy and CO2 exchange at the interface between high-population residential areas and the atmosphere. The collected observation data help to fill a gap in the current understanding of the urban surface processes in megalopolis located in East Asian monsoon-affected areas. Our observational study indicates that surface energy partitioning and carbon exchanges in high-rise residential areas in the Seoul metropolitan area are generally similar to those in other mid-latitude urban areas. However, our findings emphasize the effects of the summer monsoon on the urban SEB in the mid-latitude regions, which has not been seriously investigated in previous studies. However, caution must be used in modelling these unique properties of SEB and carbon exchanges by urban canopy models in monsoon-affected Asia.

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