USING NOCTURNAL COLD AIR DRAINAGE FLOW TO MONITOR ECOYSTEM PROCESSES IN COMPLEX TERRAIN

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Abstract. This paper presents initial investigations of a new approach to monitor ecosystem processes in complex terrain on large scales. Metabolic processes in mountainous ecosystems are poorly represented in current ecosystem monitoring campaigns because the methods used for monitoring metabolism at the ecosystem scale (e.g., eddy covariance) require flat study sites. Our goal was to investigate the potential for using nocturnal down-valley winds (cold air drainage) for monitoring ecosystem processes in mountainous terrain from two perspectives: measurements of the isotopic composition of ecosystem-respired CO2 (813CER) and estimates of fluxes of CO2 transported in the drainage flow. To test if this approach is plausible, we monitored the wind patterns, CO2 concentrations, and the carbon isotopic composition of the air as it exited the base of a young (~40 yr-old) and an old (>450 yr-old) steeply sided Douglas-fir watershed.

Nocturnal cold air drainage within these watersheds was strong, deep, and occurred on more than 80% of summer nights. The depth of cold air drainage rapidly increased to tower height or greater when the net radiation at the top of the tower approached zero. The carbon isotopic composition of CO2 in the drainage system holds promise as an indicator of variation in basin-scale physiological processes. Although there was little vertical variation in CO2 concentration at any point in time, we found that the range of CO2 concentration over a single evening was sufficient to estimate 813CER from Keeling plot analyses. The seasonal variation in 813CER followed expected trends: during the summer dry season 813CER became less negative (more enriched in 12C), but once rain returned in the fall, 813CER decreased. However, we found no correlation between recent weather (e.g., vapor pressure deficit) and 813CER either concurrently or with up to a one-week lag. Preliminary estimates suggest that the nocturnal CO2 flux advecting past the 28-m tower is a rather small fraction (<20%) of the watershed-scale respiration. This study demonstrates that monitoring the isotopic composition and CO2 concentration of cold air drainage at the base of a watershed provides a new tool for quantifying ecosystem metabolism in mountainous ecosystems on the basin scale.

Key words: advection; carbon isotopic discrimination; Cascade Mountains; CO2 flux; Douglas-fir; ecosystem respiration; old growth.

INTRODUCTION

Consider a large area of flat land. A variety of tools and measurement approaches could be used to characterize metabolic processes of the ecosystem(s) associated with this land, including micrometeorology, remote sensing, flask sampling of air above the surface, and process-level modeling (Baldocchi 1997, Canadell et al. 2000). However, many of the earth's ecosystems are regions with complex terrain. Most of the tools used to measure and monitor ecosystem metabolism, most notably conventional micrometeorology, are difficult or impossible to use in this complex terrain because nighttime advection of CO2 in cold air drainages increases the uncertainty in the estimates (e.g., Black et al. 1996, Lavigne et al. 1997, Aubinet et al. 2003, Paw U et al. 2004, Staebler and Fitzjarrald 2004). Many ecosystems in mountainous terrain are carbon rich (Harmon et al. 2004) and may be sensitive to climate change (Paw U et al. 2004), but due to topographic considerations, these ecosystems are poorly represented in current research networks to characterize climate-ecosystem interactions. Understanding how mountain ecosystems function and developing new ways to investigate interactions between ecosystem metabolism and environmental conditions in mountain ecosystems are among today’s most important ecological problems (Schimel et al. 2002).

In mountainous regions, disproportionate heating and cooling between air in the valley slopes and the surrounding area results in up-valley airflow during the day and down-valley flow at night (cold air drainage;
Fleagle 1950, Manins and Sawford 1979, Mahrt and Larsen 1982, Doran et al. 1990, Whiteman 1990). At night, the surface of the valley cools radiatively (long wave radiation is lost to the night sky). This cooling is greatest on cloudless nights because greater losses of long wave radiation occur (Stull 1988, Monteith and Unsworth 1990). Because air becomes denser as it cools, the cooler air "slides" (advects) downhill along gravity-driven pathways, often mimicking the pathways of stream systems (for more details see Whiteman 1990). Thus, the airsheds defined by cold air drainage are often superimposed on watersheds.

If the canopy interacts with the atmosphere in any way, such as by release or absorption of gases, cold air drainage flows will carry "signals" of these interactions. Several studies have shown that the isotopic composition of ecosystem-respired CO₂ (δ¹³Cₑr) changes with respect to environmental conditions (e.g., vapor pressure deficit or temperature; e.g., Ekblad and Högborg 2001, Bowling et al. 2002, Barbour et al. 2005), ecosystem type (Bowling et al. 2002), or stand age (Fessenden and Ehleringer 2002). The isotopic composition of respired CO₂ varies because C₃ plants discriminate against the heavy stable isotope of carbon (¹³C; Farquhar et al. 1989, Ehleringer et al. 1992); thus, CO₂ respired from plants and soils is depleted in ¹³C (Pataki et al. 2003). Thus, monitoring the carbon isotopic composition of CO₂ in cold air drainage flows may provide a tool to estimate δ¹³Cₑr of all or part of a watershed. However, to accurately estimate the δ¹³Cₑr, the characteristics of the cold air drainage must be understood.

The characteristics of a cold air drainage flow will vary with vegetation cover. Cold air drainage flows have been observed on slopes as small as 1% (Soler et al. 2002), and this flow is usually laminar (i.e., stratified or poorly mixed) because of the strong stability in the flow. If the land surface is sparsely vegetated, most of the radiative cooling occurs close to the ground, and the flow travels below the canopies of any trees that may be present. The flow may become nonlaminar (i.e., turbulent and mixed) if individual turbulent wakes formed in the lee of each tree branch or stem merge (Mahrt 1986). If the slope is heavily vegetated, the radiating surface to the night sky is the top of the canopy rather than the ground, and a drainage flow may occur above the canopy (Komatsu et al. 2003). In heavily vegetated terrain, the drainage may extend below the canopy, creating regions of high windspeeds above the canopy and near the ground (s-shaped vertical wind profile; e.g., Fons 1940, Allen 1968, Landsberg and James 1971, Shaw 1977, Turnipseed et al. 2003). The vertical gradient in CO₂ concentration will depend on whether the drainage is stratified (laminar) or well mixed (turbulent). Therefore, the characteristics of the cold air drainage must be understood because the precision of the method used to determine δ¹³Cₑr hinges on sampling a sufficient range of CO₂ concentration (Pataki et al. 2003).

To date, virtually all studies of δ¹³Cₑr have been conducted in relatively flat terrain. This is because measurements of δ¹³Cₑr require collecting air samples with a relatively wide range of CO₂ concentrations from a single location; the set of samples is then analyzed using a graphical approach known as a "Keeling plot" (Keeling 1958, 1961, Buchmann and Ehleringer 1998, Flanagan and Ehleringer 1998). In previous studies, the requisite range in CO₂ concentration has been obtained by sampling air at different heights over relatively flat terrain where a vertical stratification of CO₂ develops at night (e.g., Flanagan et al. 1996, Buchmann et al. 1997, Bowling et al. 2002). In mountainous terrain it may not be possible to collect an adequate range of CO₂ concentration because the vertical stratification may be disrupted by turbulence caused by flow around stems and boles (Mahrt 1986). However, we reasoned that in steeply sided watersheds with very well-developed nocturnal air drainage, the concentration of CO₂ may vary sufficiently over the course of an evening, even though it may not vary vertically; thus, it might be possible to obtain a sufficient range of CO₂ concentration for Keeling plot analysis from samples collected over time rather than over a height gradient if the cold air drainage is not stratified. An appeal of using cold air drainage is that the ground area ("footprint") contributing to respired CO₂ of samples collected in the cold air drainage system likely represents a much larger area than that contributing to samples collected over flat terrain, and it would also provide a new way to monitor variations in ecosystem metabolic processes in mountainous ecosystems.

The goal of this study was to investigate the potential for using nocturnal cold air drainage flows for monitoring ecosystem processes in mountainous terrain from two perspectives: measurements of the carbon isotope composition of nocturnal ecosystem respiration and estimates of the fraction of the respired CO₂ flux in the watershed that is advected in the drainage flow. Our specific objectives were to (1) examine nocturnal air flow patterns within two steep-sided watersheds to determine the characteristics of the cold air drainage, (2) determine whether δ¹³Cₑr can be determined by applying the "Keeling plot" approach to air sampled from cold air draining out of the base of the watersheds, and (3) estimate the CO₂ flux advected from one of the watersheds and identify uncertainties and additional measurements necessary for quantifying respired CO₂ fluxes at the watershed scale.

**METHODS**

**Study site**

Towers were constructed near the bases of two adjacent small watersheds located in the H. J. Andrews Experimental Forest in the western Cascades of central Oregon, USA (44.2° N, 122.2° W). One of the
watersheds (WS1) covers 96 ha and was harvested in the late 1960s. It is now populated by a relatively young, closed-canopied forest (~25 m) that consists primarily of Douglas-fir (Pseudotsuga menziesii [Mirb.] Franco) with a smaller component of bigleaf maple (Acer macrophyllum Pursh.), red alder (Alnus rubra Bong.), and western hemlock (Tsuga heterophylla [Raf.] Sarg.; Moore et al. 2004). The other watershed (WS2) covers 60 ha, has not had a major disturbance for more than 450 years, and has a mix Douglas-fir and western hemlock that accounts for ~90% of the sapwood basal area (Moore et al. 2004). This old-growth forest is >60 m tall and has large canopy gaps that are typical of old-growth Douglas-fir/western hemlock forests in the Pacific Northwest (Gray and Spies 1996, Franklin et al. 2002). The site has wet, mild winters and warm, dry summers with a mean annual precipitation of 2300 mm. The soil is described as gravelly clay loam (Swanson and James 1975).

**Tower measurements**

Towers, initially 10 m tall, were constructed in the late spring of 2002 at the base of the young and old-growth watersheds because past research suggested that the air drainage flows may be shallow (Mahrt 1986). However, it quickly became apparent that the drainage was substantially deeper in our ecosystems, and the tower in the young watershed was extended to 28 m (close to the canopy top) and furnished with more instruments in the spring of 2003 (Table 1). The tower in the old-growth watershed was not extended for budgetary reasons. Wind speed and wind direction were measured on the towers using a combination of cup anemometers (010C, Met One Instruments, Grants Pass, Oregon, USA and Ultralight, Thornthwaite, Centerton, New Jersey, USA), wind vanes (020C, Met One Instruments), propeller anemometers (Model 05103, R. M. Young, Traverse City, Michigan, USA), and sonic anemometers (WS425, Vaisala, Helsinki, Finland). Air temperature was measured at several heights using shielded thermistors (107, Campbell-Scientific, Logan, Utah, USA) and net radiation was measured at the top of the tower with a net radiometer (Q7, Campbell-Scientific). Signals were measured by a data logger (CR10x, Campbell-Scientific) every second and were averaged over either 1 minute (sonic anemometers) or 15 minute intervals (all other meteorological instruments). Continuous measurements of these instruments began in May 2003.

**Definition of cold air drainage**

For the wind patterns within this valley to be considered nocturnal cold air drainage, we required the flow measured at the tower to meet three criteria: (1) the wind direction must be within 90° and 135° (down valley); (2) the wind-speed profile measured on the 28-m tower must have a local maximum near the ground (Stull 1988); and (3) because we were only interested in nocturnal cold air drainage, the net radiation measured above the tower was required to be negative (indicating a surface cooling; Whiteman 1989). We focused exclusively on nighttime drainage flow because our interest was in nocturnally collected samples of air for Keeling plot analyses. However, it should be noted that under the right conditions, cold air drainage can continue all day below the canopy while uphill flow occurs above the canopy (Hawkes 1947, Whiteman 1986, 1990). The prevailing wind direction measured on the ridge at the head of these valleys was typically up valley, between 240° and 360° (data not shown). Hence, down-valley winds near the surface usually indicated cold air drainage in the valley.

**Measurement of CO₂ concentration**

We measured the CO₂ concentration of the air in the nocturnal cold air drainage in the young and old-growth watersheds on six and three nights, respectively, between June and September 2003 using a vertical profile system. In the younger watershed, a switching system was used to sample air sequentially from six heights on the tower (3, 6, 15, 20, and 28 m). Because the tower was smaller in the old-growth watershed, the profile system only monitored CO₂ concentration at 3, 6, and 10 m. In both watersheds, the air from each height was pumped (pump model UNMP30KNDC, KNF Neuberger, Trenton, New Jersey, USA) at 1.5 L/min for 2 minutes to an infrared gas analyzer (L16262, Li-Cor, Lincoln, Nebraska, USA). The infrared gas analyzer was calibrated in the field using a nitrogen tank (no CO₂) and a second
tank of known CO₂ concentration. To ensure that the measurements only represented the current interval, the last 30 seconds for each 2 minute interval were averaged and stored on a data logger (CR10x, Campbell-Scientific). Each sample height had its own 6.4 mm diameter tubing that connected to a switching mechanism placed near the infrared gas analyzer.

Isotope Sampling and Analysis

Air samples for isotope analysis were collected by hand in the field at multiple heights in the young (3, 6, 10, 15, 20, and 28 m) and old-growth forests (1, 3, and 10 m). Air was pumped at -1.5 L min⁻¹ (Model number UNMP30KNDG, KNF Neuberger) from one of the heights through an infrared gas analyzer (IRGA; LI-6252, Li-Cor) and then discharged out of a tygon tube inserted into a 15-mL septum-sealed vial, which was flushed with gas for approximately one minute. As the sample was collected, the CO₂ concentration from the IRGA was noted and a cap was quickly placed onto the vial as the tygon tube was simultaneously withdrawn. To smooth out higher frequency variations transition in the gas compositions, the air was passed through a small buffer volume (2 L) before it was routed to the IRGA. For each sample period 16-40 sets of vials were collected and analyzed.

For δ¹³C analysis, we used a Finnigan/MAT Delta-Plus XL isotope ratio mass spectrometer (San Jose, California, USA) interfaced to a GasBench II automated headspace sampler (Finnigan) at the College of Oceanic and Atmospheric Sciences, Oregon State University, Corvallis, Oregon, USA. The GasBench-II is a continuous flow interface that allows injections of several aliquots of a single gas sample into a mass spectrometer for automated isotope determinations of small gas samples. A typical analysis consists of three gas standards (tank CO₂-He mixtures), six sample replicates, and an additional two gas standards.

The tank CO₂ was calibrated both on a MAT-252 (dual inlet) mass spectrometer (Finnigan), by comparison to NIST-8544 (also known as NBS-19) limestone prepared using a Kiel-III online acid digestion device (Finnigan), which yields isotopic values of -2.19 ± 0.06‰ for δ¹⁸O and 1.94 ± 0.02‰ for δ¹³C (n = 25), which compares with certified values of -2.20‰ and 1.95‰ for δ¹⁸O and δ¹³C, respectively (no precision estimate; NIST 1992a). The reference gas was also calibrated on the DELTApplus-XL (Finnigan) by analysis of NIST-8541 (also known as USGS-24 Graphite, certified value of δ¹³C = -15.90 ± 0.25‰) and NIST-8542 (also known as ANU Sucores, certified value of δ¹⁸O = -10.47 ± 0.13‰), which were prepared by online combustion using a Carlo-erba NA-1500 elemental analyzer (Milan, Italy), and CONIFLO-III inlet system (Thermo Finnigan, San Jose, California, USA). Repeated analyses of a single gas sample (tank CO₂-He mixtures of 1% CO₂) by the GasBench-II yields external precision for δ¹³C in multiple vials of this continuous flow device to ±0.03‰, which is close to the level typical of dual-inlet mass spectrometry (typically ±0.02‰ on our Finnigan/MAT-252).

We conducted extensive tests of procedures for collecting and storing air samples. We found no measurable change in isotope ratios of air samples stored in 15-mL vials sealed with non-punctured butyl or Kel-F septa for up to one week or moved across an altitude range of >500 m. However, if the septa were punctured, the isotope values changed significantly during transport and storage, and unlike Tu et al. (2001), we were not able to adjust for the change with punctured and non-punctured standards. Therefore, we collected field samples in glass vials, capping them by hand with non-punctured septa. Between early June and late September of 2003, we conducted thirteen measurements of δ¹³CER: seven in WSI and six in W2.

The δ¹³CER was estimated using a two-end member-mixing model or “Keeling plot” (Keeling 1958, 1961, Buchmann and Ehleringer 1998, Flanagan and Ehleringer 1998). The δ¹³C of the air samples in the vials was plotted against 1/CO₂ concentration. Because there are measurement errors in both the x and y axes (Pataki et al. 2003), a geometric mean regression approach (Sokol and Rohlf 1993) was used to estimate the value of the y-intercept, or δ¹³CER. The use of geometric mean regression for calculating Keeling plot intercepts has recently been challenged (Zobitz et al. 2006). However, the use of geometric mean regression instead of ordinary least squares regression will have little effect on conclusions drawn in this paper. Because collecting samples by hand has a high probability of operator error (e.g., recording the CO₂ concentration off the IRGA that represents the CO₂ concentration collected in the file), outliers in each data set were systematically removed. Using a procedure outlined by Bowling (2002), we selected outliers by first performing a geometric mean regression instead of ordinary least squares regression and then running a procedure for removing outliers. Future work with an automated sampler will hopefully negate the need to remove outliers in this manner.

Pre-dawn water potential

Pre-dawn water potential of trees was measured on most of the nights that air samples were taken for isotope analysis. Water potential of small twigs was measured one hour prior to sunrise using a field portable pressure chamber (PMS systems, Corvallis, Oregon, USA). In the young watershed, between 5 and 10
Douglas-fir trees were randomly selected within a 100-m radius of the tower, and one branchlet from each tree was removed to determine water potential. In the old-growth forest, five Douglas-fir and five western hemlock trees were randomly selected and water potential determined on shoots from the lower-canopy.

Pre-dawn twig water potentials for sampling nights where pre-dawn measurements were not made were estimated by linear interpolation from the other data points. In the Pacific Northwest there is a pronounced summer dry season and the soil water continuously decreases as the summer progresses (Parsons et al. 2001). During the summer of 2003 there were only two rainfall events in July and August, totaling <5.75 mm of rain (Pypker et al. 2006). Given that young and old-growth Douglas-fir forests lose approximately 20% of gross precipitation to evaporation during a storm event and have a canopy water storage of 1–2 mm and 3 mm respectively (e.g., Klaassen et al. 1998, Pypker et al. 2005), only a very small fraction of this rainfall would reach the forest floor. Therefore, interpolating pre-dawn water potential between measurements is not unreasonable in this ecosystem.

**RESULTS**

**Characteristics of the cold air drainage**

Nocturnal down-valley winds (20:00–05:00) were observed at all heights to 28 m from May to December 2003 (DOY 121–365; Fig. 1). The proportion of down-valley flow was calculated for each measurement height. The cold air drainage in the old-growth watershed followed a similar pattern for periods when the winds (>0.45 m/s) were sufficient to move the wind vane (data not shown). The drainage was most common at the lower heights and during the summer months when the days were warm and the nights were clear. Cloudless nights allowed for greater losses of long-wave radiation from the surface to the atmosphere at night.

The shapes of the vertical wind profiles during cold air drainage were similar from May to December. Beneath the canopy, the wind speed was maximum at 5 m/s, with increasing drag causing decreased wind speeds near the ground surface (1.5 m/s) and within the canopy air space (18 m/s to 25 m/s; Fig. 2). The drainage flow appeared to extend above the canopy to an unknown depth greater than the tower height (Fig. 2), and its velocity typically decreased from 21:00 to 04:00 (Fig. 3a).

The depth of cold air drainage rapidly increased to tower height or greater when the net radiation at the top of the tower approached zero and the vertical gradient of potential temperature approached zero (Fig. 4). Potential temperature represents the temperature an air mass would attain if it expanded or compressed adiabatically from its current elevation to sea level (for more information see Wallace and Hobbs 1977). Hence, it allows for direct comparison between temperature measurements from different elevations.

At all the heights, the down-valley wind speeds were greatest in the early evening, when the thermally caused pressure gradient (greater air density near the slope) between the air adjacent to the slope and the ambient air at the same elevation over the valley was presumably the greatest (Whiteman 1990). The wind speeds typically decreased as the evening progressed.

The vertical potential temperature profile followed a well-defined pattern in summer. By day, the air...
temperature was greatest near the canopy top where the greatest quantity of solar radiation is absorbed (Fig. 3b). At all heights, the daily maximum air temperatures occurred at 15:00, after which the air temperatures at all heights steadily declined. The decline was greatest within the canopy air space, and by 20:00 the potential temperature profile became nearly isothermal, indicating that the air was well mixed. The air temperatures remained close to isothermal until the following morning (Fig. 3b).

**CO₂ concentration profiles**

The CO₂ vertical concentration profiles measured on the towers in young and old-growth watersheds followed a similar pattern to the temperature profiles (Fig. 5). During the day (07:00–20:00), CO₂ concentration varied with height, with the greatest concentrations occurring near the ground. At about 20:00 the air flow in the canopy became well mixed, and the CO₂ concentrations at all heights converged (Fig. 5) at approximately the same time as the potential temperature profile became nearly isothermal (Fig. 4). The CO₂ concentrations steadily increased throughout the night until sunrise. After sunrise, the drainage remained well mixed for 1–2 h, but the CO₂ concentrations began to decline. At approximately 07:00 when the potential temperature profile became stratified, a vertical gradient of CO₂ concentration was re-established. During the course of a night, the range of CO₂ concentration in the well-mixed flow was typically between 30 and 80 ppm in both the young and old-growth forests. The data presented are primarily from the young forest because it contained the taller tower. However, the pattern of CO₂ concentrations was similar at the 10-m tower in the old-growth stand.

**Isotopic composition**

The range of CO₂ concentration used in the Keeling plots was from 47 to 92 ppm and from 64 to 116 ppm for the young and old-growth forests, respectively (Table 2). There were no discernable differences in the isotopic composition of the air measured at the different heights because the air was well mixed (Fig. 5). Therefore, the samples from the different heights were grouped together to provide an estimate of δ¹³C_E to for each measurement night.

From 2 July to 21 September 2003 (DOY 183–264) the δ¹³C_E was estimated to range between −24.5% and −27% (Table 2). During this same period, the forests received only 6 mm of rainfall (Pypker et al. 2006). As
the seasonal summer dry season progressed and the water potential of the trees decreased (indicating more water stress), the $\delta^{13}$CER became more enriched in both watersheds. The first significant rainfall (58 mm) after the summer dry season occurred on 7–9 September 2003 (DOY 250–252) (Pypker et al. 2006). After this rainfall, the $\delta^{13}$CER decreased from $-25.42 \pm 0.65$ to $-26.43 \pm 0.70$ (Table 2), but the difference is not significant, as the confidence intervals of each mean overlap.

In contrast with Fessenden and Ehleringer (2002), who reported significant enrichment in $\delta^{13}$CER in an old-growth Douglas-fir forest compared with a young forest, we found no consistent difference in $\delta^{13}$CER between the young and old-growth watersheds in our study. The $\delta^{13}$CER was linearly related to the water potential of the upper canopy trees in the young watershed (linear regression slope $P$ value $= 0.05$, $R^2 = 0.68$, $\alpha = 0.05$), but not in the old-growth watershed (linear regression slope $P$ value $= 0.4$, $R^2 = 0.18$, $\alpha = 0.05$; Fig. 6). Incorporating time lags into the analyses did not substantially improve
the correlations. The δ13C_{ER} was not linearly related to concurrent vapor pressure deficit or to time-lagged vapor pressure deficit ranging from 1 to 10 days (all P values > 0.05, α = 0.05).

Estimates of advected CO2 flux

Monitoring the air exiting the base of the watershed may also provide an opportunity to quantify nighttime respiration within the watershed using a mass-balance approach. For example, if the valley is treated as an open cuvette, a rough estimate of respiration occurring inside the cuvette (watershed) can be made:

\[ R_e = F_0 - F_1 + \Delta F_u \]  

where \( R_e \) is the rate of CO2 input from ecosystem respiration (mol/s), \( F_0 \) is the output flux of CO2 exiting from the valley (mol/s), \( F_1 \) is the input flux to the valley of atmospheric CO2 (at background concentration), and \( \Delta F_u \) is the rate of storage of respired CO2 within the watershed (mol/s). The precision with which \( R_e \) can be estimated from Eq. 1 depends on the magnitudes and uncertainties of the other terms on the right-hand side of the equation. If all the CO2 exiting the valley was in the drainage flow measured below the top of the tower, we could estimate \( F_0 \) by treating the cross section of the watershed at the base tower as multiple layers that are perpendicular to the cold air drainage and calculating the product of wind speed and measured CO2 concentration for each layer. With a similar assumption, the value of \( F_1 \) could be estimated as the product of wind

Table 2. The isotopic composition of ecosystem respiration δ13C_{ER} for the young and old forests.

<table>
<thead>
<tr>
<th>Forest</th>
<th>Date</th>
<th>CO2 range ppm (maximum-minimum)</th>
<th>Max. CO2 ppm</th>
<th>δ13C_{ER} (%)</th>
<th>SE</th>
<th>R²</th>
</tr>
</thead>
<tbody>
<tr>
<td>Young</td>
<td>2 Jul 2003</td>
<td>92.3</td>
<td>460</td>
<td>-26.42</td>
<td>0.18</td>
<td>0.99</td>
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<tr>
<td>Young</td>
<td>15 Jul 2003</td>
<td>75.6</td>
<td>446</td>
<td>-26.20</td>
<td>0.35</td>
<td>0.99</td>
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<tr>
<td>Young</td>
<td>23 Jul 2003</td>
<td>60.7</td>
<td>431</td>
<td>-26.98</td>
<td>0.35</td>
<td>0.99</td>
</tr>
<tr>
<td>Young</td>
<td>30 Jul 2003</td>
<td>69.2</td>
<td>452</td>
<td>-25.99</td>
<td>0.33</td>
<td>0.98</td>
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<tr>
<td>Young</td>
<td>13 Aug 2003</td>
<td>61.8</td>
<td>429</td>
<td>-24.49</td>
<td>0.29</td>
<td>0.99</td>
</tr>
<tr>
<td>Young</td>
<td>26 Aug 2003</td>
<td>47.0</td>
<td>411</td>
<td>-25.42</td>
<td>0.65</td>
<td>0.98</td>
</tr>
<tr>
<td>Young</td>
<td>7 Sep 2003</td>
<td>55.0</td>
<td>415</td>
<td>-26.43</td>
<td>0.70</td>
<td>0.98</td>
</tr>
<tr>
<td>Young</td>
<td>21 Sep 2003</td>
<td>60.4</td>
<td>460</td>
<td>-25.76</td>
<td>0.54</td>
<td>0.98</td>
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<td>Old-growth</td>
<td>7 Aug 2003</td>
<td>107</td>
<td>519</td>
<td>-26.12</td>
<td>0.12</td>
<td>0.99</td>
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<tr>
<td>Old-growth</td>
<td>22 Jul 2003</td>
<td>64.6</td>
<td>490</td>
<td>-25.47</td>
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<td>0.97</td>
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<tr>
<td>Old-growth</td>
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<td>534</td>
<td>-25.54</td>
<td>0.18</td>
<td>0.99</td>
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<td>Old-growth</td>
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<td>454</td>
<td>-25.89</td>
<td>0.72</td>
<td>0.97</td>
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<td>Old-growth</td>
<td>26 Aug 2003</td>
<td>63.3</td>
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<td>-25.45</td>
<td>0.55</td>
<td>0.99</td>
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<td>505</td>
<td>-26.82</td>
<td>0.23</td>
<td>0.99</td>
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</tbody>
</table>

Notes: The values of δ13C_{ER} were estimated using geometric mean regression. The R² values are for linear regression only.
speed in each layer and background CO₂ concentration, i.e., if there were no respiration or storage \( F_i = F_0 \). Thus, watershed-scale respiration \( R_w \) would be given by the net advected flux \( (F_0 - F_i) \) plus the storage term \( \Delta F_{st} \). In practice, drainage flow and other exchange between air in the watershed and the free atmosphere probably also occur above the top of the tower, so the net advection in the valley cross-section below the tower top \( (F_0 - F_{1h}) \) is probably an underestimate of the total net advection, as we discuss later.

We calculated values of \( (F_0 - F_{1h}) \), as described in the Appendix, for data on a typical night of drainage flow in Watershed 1. Fig. 7 shows that \( (F_0 - F_{1h}) \) ranged from 0.3 to 1.3 mol/s between 20:00 to 05:00 on 5-6 August 2003.

**DISCUSSION**

**Characteristics of nocturnal cold air drainage**

During the summer months, nocturnal cold air drainage was frequent in both watersheds (Fig. 1) because summer in western Oregon has little rainfall and mostly clear skies (Taylor and Hannan 1999). These conditions promote cold air drainage, as frequent clear skies increase the occurrence of nocturnal cold air drainage down steep, forested valleys (Whiteman 1990, Whiteman and Doran 1993). Past research has demonstrated that cold air drainages can range from a few meters (e.g., Mahrt and Larsen 1982, Mahrt et al. 2001) to tens or hundreds of meters in depth (e.g., Manins and Sawford 1979, Doran et al. 1990, Whiteman 1990, Smith and Skillingstad 2005). The nocturnal in-canopy drainage flow at this study site was frequent (Fig. 1), deep (>28 m), and usually well mixed (Fig. 5). The depth of the cold air drainage is consistent with past measurements in forested ecosystems where the cold air drainage extended well above the canopy, with a second maximum near the canopy top (e.g., Komatsu et al. 2003). Given that the dominant wind direction is between 240° and 360° (data not shown), downhill flow generally indicated nocturnal cold air drainage.

The cold air drainage persisted throughout most nights (Fig. 1), but its velocity declined as the night progressed (Fig. 3a). The nocturnal wind speeds may decline during the night because warmer air from above becomes entrained within the drainage, thereby decreasing the buoyancy driving the flow (Barr and Orgill 1989). Alternatively, it might indicate the “damming effect” of a pool of cold air accumulating at the base of the watershed that generates hydrostatic pressure which slows the flow of cold air (e.g., Magono et al. 1982, Yoshino 1984, Maki and Harimaya 1988, Mori and Kobayashi 1996, Iijima and Shinoda 2000, Clements et al. 2003). Regardless of why the wind slows as the night progresses, the data do indicate that cold air drainage in these basins is common during the summer.

**FIG. 7.** The estimated flux of CO₂ passing the 28-m tower located at the base of the watershed on 5-6 August 2003. The fluxes were estimated using the method outlined in the Appendix.
Feasibility of quantifying δ¹³C_{ER} of the air in the cold air drainage using the “Keeling-plot” approach

Three considerations are involved in evaluating whether air samples from the cold air drainage can be used to determine δ¹³C_{ER} using Keeling-plot analyses. First, there must be a sufficient range of CO₂ concentration for accurate estimation of the intercept in the Keeling plot (Pataki et al. 2003). Second, the isotopic composition of the “source” must be stable over the measurement period (Keeling 1958, 1961, Ehleringer et al. 1992). In other words, the flux from the land surface that contributes to respired CO₂ in the samples (i.e., the “footprint”) must be stable, and there must not be a significant shift in the components of respiration (i.e., aboveground vs. belowground) or their isotopic characteristics. Third, the observed variations in δ¹³C_{ER} should be consistent with environmental variations (Farquhar et al. 1989, Madhavan et al. 1991, Ekblad and Högberg 2001, Bowling et al. 2002).

Is the range of CO₂ concentration adequate for Keeling-plot analyses?—After analyzing Keeling plots from 26 locations in the Americas, Pataki et al. (2003) reported that a minimum value of 75 ppm CO₂ is required to maintain a standard error in δ¹³C_{ER} of less than 1%. Of the 14 sampling nights for the young and old-growth forests in our study, 12 had a range in CO₂ concentration that was greater than 60 ppm (Table 2). While 60 ppm is lower than the desired range suggested by Pataki et al. (2003), the standard errors for estimating the intercept (δ¹³C_{ER}) were between 0.12% and 0.7%, well below the desired 1% (Table 2). However, our precision, along with at least some of data presented in Pataki et al. (2003; e.g., Bowling et al. 2002), was improved because we removed data points that were two standard deviations away from the predicted regression line. As reported in Pataki et al. (2003), the standard error rapidly increases when the range of CO₂ concentration drops below 75 ppm (Table 2). For the two sampling periods when the range in CO₂ concentration was less than 60 ppm, the SE was greater than 0.65%. Overall, we conclude that the nocturnal variation in CO₂ concentration in the two basins we investigated is more than adequate for accurate determination of the Keeling intercept.

Is the isotopic composition of respired CO₂ consistent over the measurement period?—In this study, we were not able to evaluate whether the “footprint,” or the aboveground/belowground components, of respiration varied over the course of nocturnal measurement periods. However, if there were a significant change in the isotopic composition of the source over the time that samples were taken, it should be apparent as a “kink” or curve in the Keeling plot line (because the ordinate is plotted as 1/CO₂ concentration, and CO₂ concentration tended to increase over time for each set of samples; Fig. 5). We found no evidence of kinks or curves in any of the data we collected, so we conclude there is no evidence for variation in the isotopic composition of the source during measurements. However, these questions require closer investigation in the future.

Relationship of δ¹³C_{ER} to environmental variables.—The technology involved in collecting samples for determination of δ¹³C_{ER} is rapidly evolving. Although the laboratory precision of δ¹³C determination in our samples was excellent, our methods for collecting air samples and determining CO₂ concentration in this study were somewhat primitive and labor intensive. We have subsequently developed a more precise, automated system for collecting air samples and have improved methods for measuring CO₂ concentration in the laboratory. Nevertheless, the methods used in this study reveal that considerable variation in δ¹³C_{ER} is likely due to short-term variations in canopy physiology, demonstrating that the approach is well worth pursuing further.

The δ¹³C_{ER} values measured in this study follow an expected pattern, for a region with a seasonal summer dry season (Bowling et al. 2002), ranging between −24.5 and −27% (Table 2). Past research reported δ¹³C_{ER} for Douglas-fir forests in the Pacific Northwest to range between −24% and −33% (Bowling et al. 2002, Fessenden and Ehleringer 2002, Ponton et al. 2006). In this study, the seasonal trend for δ¹³C_{ER} was to gradually increase as the summer dry season progressed (Table 2). Only after a significant rainfall in September did the δ¹³C_{ER} appear to decrease (Table 2). Bowling et al. (2002) monitored δ¹³C_{ER} at six forested sites in Oregon that stretched along a precipitation gradient. Two of these sites were young Douglas-fir forests (−13 and 30 years old) and, as at the study site here, Bowling et al. (2002) reported that from June to September the δ¹³C_{ER} for the 13 and 30 year old Douglas-fir forests increased from approximately −29.5% to −25.5% and from −27% to −25.5%, respectively. Furthermore, Ponton et al. (2006), when monitoring seasonal changes in δ¹³C_{ER} during the seasonal summer dry season at a 50 yr-old Douglas-fir forest on Vancouver Island, British Columbia, Canada, found that between May and September the δ¹³C_{ER} increased from approximately −27.5% to −24.8%. To prevent xylem cavitation and embolism, Douglas-fir trees use their stomata to regulate water potential in the tree (Phillips et al. 2002, Domec et al. 2004). By closing their stomata, the isotopic signature of the recently fixed photosynthate is thus expected to increase (Farquhar et al. 1989, Madhavan et al. 1991). Hence, past researchers have shown δ¹³C_{ER} to be sensitive to changes in vapor pressure deficit or soil moisture (Ekblad and Högberg 2001, Bowling et al. 2002). However, despite following seasonal changes in water availability, the variation in the δ¹³C_{ER} was not closely correlated with any one of the environmental variables we investigated.

In contrast with studies of Ekblad and Högberg (2001) and Bowling et al. (2002), the variation of δ¹³C_{ER} was not significantly related to vapor pressure deficit using any time lag when compared using linear regression ($R^2$=...
0.01 with no time lag). There was a significant correlation between $\delta^{13}$C$_{ER}$ and pre-dawn water potential of dominant trees in the young forest, but not the old-growth forest (Fig. 6). It is not surprising that $\delta^{13}$C$_{ER}$ cannot be explained by such simple environmental parameters because the isotope signal integrates a large number of variables. As McDowell et al. (2002) noted, environmental variations that affect canopy physiology (e.g., temperature, humidity, soil moisture) are also likely to affect the relative contribution of different respiratory sources (e.g., foliage, litter, roots, bulk soil), thereby affecting $\delta^{13}$C$_{ER}$ in more complex ways.

The larger sampling area increases the heterogeneity of the components contributing to the $\delta^{13}$C$_{ER}$ measured at the tower. One of the benefits of monitoring the isotopic signal contained within the cold air drainage is that it provides an integrated signal of the $\delta^{13}$C$_{ER}$ for all or part of a watershed. However, the larger sampling area incorporates respiration from riparian and upslope positions. Given that the microclimates at these locations are likely very different, it is possible that the lack of correlation between vapor pressure deficit and $\delta^{13}$C$_{ER}$ may result, in part, because of this heterogeneity.

**Advected CO$_2$ flux estimates**

Estimates of the net advected flux of carbon dioxide ($F_0 - F_t$), through the valley cross-section below the top of the tower range from 0.3 mol/s to 1.3 mol/s (Fig. 7). Past work in similar forests of this region suggests that the respiratory flux per unit ground area is approximately 7 µmol CO$_2$ m$^{-2}$ s$^{-1}$ (Harmon et al. 2004, Campbell and Law 2005). Given that the floor area of the valley is about 96 ha, this corresponds to a watershed-scale respiratory flux $R_0$ of approximately 7 mol/s. Therefore, this method estimates that the advected flux passing the 28-m tower accounts only for approximately 5-18% of the total ecosystem respiration within the 96-ha watershed. If our advection calculations and assumptions are accurate, the remaining respired CO$_2$ must have either been advected in drainage flow above the tower or have exited the flow by becoming entrained in the overlying air or have been stored in the sub-canopy airspace up valley from the tower. An alternative explanation is that the source area of the CO$_2$ measured in the drainage flow (the footprint of the tower) may not have included the entire watershed, so that the CO$_2$ in the cold air drainage that advected past the tower was only a small fraction of the watershed respiration.

Quantifying advective fluxes using a single vertical profile measured at the base of the watershed is uncertain, as windspeeds and CO$_2$ concentrations are assumed to be horizontally homogeneous (King 1989). In the present case, the drainage flow was well mixed, so homogeneity of CO$_2$ concentration is a reasonable assumption (Fig. 5). However, wind speeds were not likely to be horizontally homogeneous. For example, wind speeds measured above the canopy are unlikely to represent wind speeds on the adjacent vegetated slope at the same elevation because of frictional drag exerted by the foliage (e.g., Fons 1940, Allen 1968, Landsberg 1971, Shaw 1977, Turnipseed et al. 2003). King (1989) concluded that calculations based on a single vertical profile overestimated the advected volume flux of air out of a watershed in Colorado by a factor of two.

Although CO$_2$ storage in canopy airspace is unlikely to be a significant term in the mass balance over long sampling times, it is probably not negligible on the time scale of Fig. 7. The largest cause of the difference between $(F_0 - F_t)$ and $R_0$ is probably the footprint of the tower being smaller than the area of the watershed. To more accurately estimate fluxes the depth of the cold air drainage and the footprint of the tower must be determined. We plan future studies with a taller tower, an inert tracer gas, and improved spatial sampling of temperature gradients in the watershed to determine whether CO$_2$ mass balance studies can be refined to provide better estimates of ecosystem respiration in steep terrain.

**Conclusion**

Nocturnal air drainage within the canopies of two steep, forested watersheds was strong and deep. The flow became well mixed at roughly the same time as net radiation above the canopy became negative and a temperature inversion formed at the canopy top. The well-mixed regime persisted through the night and into the early morning. These deep, well-mixed drainage systems offer opportunities to monitor and quantify ecosystem function on the basin or sub-basin scale in complex, mountainous terrain. Nocturnal drainage carries a large quantity of ecosystem-respired CO$_2$. The carbon isotope composition of CO$_2$ in the drainage system holds promise as an indicator of variation in basin-scale physiological processes. We found that the range of CO$_2$ concentration over a single evening was sufficient to estimate $\delta^{13}$C$_{ER}$ from Keeling plot analyses and that variation in $\delta^{13}$C$_{ER}$ over a summer dry season and early fall rains generally followed expected trends. Estimates of the advected CO$_2$ flux in the cold air drainage were much smaller than the likely respiration from the entire watershed. Eventually, it may be possible to quantify variations in ecosystem respiration rates from mass balance measurements in these flows. However, to quantify the mass flux of carbon and to use $\delta^{13}$C$_{ER}$ to monitor intra- and interannual variations in ecosystem metabolism on the basin scale requires further work, particularly better knowledge of vertical and lateral advective transfer within the ecosystem and a deeper understanding of the interactions between environmental drivers and the short-term dynamics of carbon cycling.

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APPENDIX

Estimation of the net flux ($F_0 - F_1$) of CO2 in the valley cross-section below the tower top (Ecological Archives A017-029-A1).