

Variability and budget of CO₂ in Europe: analysis of the CAATER airborne campaigns – Part 2: Comparison of CO₂ vertical variability and fluxes between observations and a modeling framework

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Received: 11 October 2009 – Published in Atmos. Chem. Phys. Discuss.: 12 February 2010

Revised: 30 July 2010 – Accepted: 24 August 2010 – Published: 20 June 2011

Abstract. Our ability to predict future climate change relies on our understanding of current and future CO₂ fluxes, particularly on a regional scale (100–1000 km). CO₂ regional sources and sinks are still poorly understood. Inverse transport modeling, a method often used to quantify these fluxes, relies on atmospheric CO₂ measurements. One of the main challenges for the transport models used in the inversions is to properly reproduce CO₂ vertical gradients between the boundary layer and the free troposphere, as these gradients impact on the partitioning of the calculated fluxes between the different model regions. Vertical CO₂ profiles are very well suited to assess the performances of the models. In this paper, we conduct a comparison between observed and modeled CO₂ profiles recorded during two CAATER campaigns that occurred in May 2001 and October 2002 over Western Europe, as described in a companion paper. We test different combinations between a global transport model (LMDZt), a mesoscale transport model (CHIMERE), and different sets of biospheric fluxes, all chosen with a diurnal cycle (CASA, SiB2 and ORCHIDEE). The vertical profile comparison shows that: 1) in most cases the influence of the biospheric flux is small but sometimes not negligible, ORCHIDEE giving the best results in the present study; 2) LMDZt is most of the time too diffuse, as it simulates a too high boundary layer height; 3) CHIMERE better reproduces the observed gradients between the boundary layer and the free troposphere, but is sometimes too variable and gives rise

to incoherent structures. We conclude there is a need for more vertical profiles to conduct further studies to improve the parameterization of vertical transport in the models used for CO₂ flux inversions.

Furthermore, we use a modeling method to quantify CO₂ fluxes at the regional scale from a chosen observing point, coupling influence functions from the transport model LMDZt (that works quite well at the synoptic scale) with information on the space-time distribution of fluxes. This modeling method is compared to a dual tracer method (the so-called Radon method) for a case study on 25 May 2001 during which simultaneous well-correlated in situ CO₂ and Radon 222 measurements have been collected. Both methods give a similar result: a flux within the Radon 222 method uncertainty (35%), that is an atmospheric CO₂ sink of -4.2 to -4.4 gC m⁻² day⁻¹. We have estimated the uncertainty of the modeling method to be at least 33% on average, and even more for specific individual events. This method allows the determination of the area that contributed to the CO₂ observed concentration. In our case, the observation point located at 1700 m a.s.l. in the north of France, is influenced by an area of 1500×700 km² that covers the Benelux region, part of Germany and western Poland. Furthermore, this method allows deconvolution between the different contributing fluxes. In this case study, the biospheric sink contributes 73% of the total flux, fossil fuel emissions for 27%, the oceanic flux being negligible. However, the uncertainties of the influence function method need to be better assessed. This could be possible by applying it to other cases where the calculated fluxes can be checked independently, for example at tall towers where simultaneous CO₂ and Radon 222



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measurements can be conducted. The use of optimized fluxes (from atmospheric inversions) and of mesoscale models for atmospheric transport may also significantly reduce the uncertainties.

1 Introduction

Predictions of future climate change rely on our ability to understand the present and future distribution of CO₂ fluxes (e.g. Geels et al., 2007). However, the value of CO₂ fluxes is still uncertain, especially at the regional scale of 100–1000 km (e.g. Patra et al., 2008; Law et al., 2007; Gurney et al., 2004, 2002). Several methods to quantify CO₂ fluxes exist, mainly inverse modeling (e.g. Rödenbeck et al., 2003; Gloor et al., 2001; Bousquet et al., 1999), the Radon method (e.g. Schmidt et al., 2003, 2001), the boundary layer budget method (e.g. Gibert et al., 2007), and tower flux measurements (e.g. Haszpra et al., 2005). Inverse modeling is the most used approach to quantify regional fluxes, and relies on measurements of atmospheric CO₂ concentrations. Because of the large area they can span in a short time, airborne facilities are a well suited for measuring CO₂ concentrations at the regional scale. In a recent paper, Stephens et al. (2007) highlighted the need to record more vertical profiles for cross-validation of atmospheric transport models. We rely here on airborne in situ CO₂ measurements recorded during two CAATER airborne campaigns conducted on 23–26 May 2001 and 2–3 October 2002 over Western Europe, and during which in situ CO₂ (for both campaigns), CO (CAATER 2 only) and semi-continuous Radon 222 (CAATER 1 only) measurements were collected. In a companion paper (Xueref-Remy et al., 2011), we described the observed atmospheric CO₂ variability. Here, we compare models with observations for the CAATER campaigns. We first assess how a global model and a mesoscale one reproduce CO₂ vertical variability, and second, we use modeled influence functions to quantify CO₂ fluxes during a case study, and assess these results using ²²²Rn-CO₂ observations in the framework of the so-called “Radon method”.

A major source of uncertainty (bias) in atmospheric transport models used in global inversions is how well they represent the variability of CO₂ with altitude (Stephens et al., 2007). Given the range of vertical transport processes (deep convection, boundary layer thermic and dynamical mixing, frontal uplift...) the transport between the atmospheric boundary layer (ABL) and the free troposphere (FT) remains fairly uncertain. This process can be constrained by using CO₂ as a transport tracer and looking at the vertical gradient between the ABL and the FT (Sarrat et al., 2007; Yi et al., 2004; Gerbig et al., 2003a, b; Ramonet et al., 2002). Indeed, the gradient between ABL and FT has an impact on the determination of CO₂ fluxes. As pointed out in Stephens et al. (2007), not only averaged profiles over large regions

should be compared, but also profiles at individual sites. Here, we test the influence of the model scale (global and mesoscale) on the reproducibility of the observed vertical variability, but also different land flux models, all chosen with a diurnal cycle as they give better results than models using only monthly means or daily average fluxes (Patra et al., 2008).

Our motivation for the second focus of this paper is that using tools such as influence functions (IF) (e.g. Lauvaux et al., 2009), regional CO₂ fluxes can be constrained. We use here a method to constrain regional to continental fluxes (500–1000 km) directly from observations, coupling influence functions from a transport model and a distribution of fluxes. The model used is LMDZt, which reproduces synoptic transport quite well (Patra et al., 2008). The results of the method are assessed with the independent use of Radon-222, a tracer of known surface fluxes that scales with unknown CO₂ fluxes. We conducted our work on a case study flight of 25 May 2001, during which both atmospheric CO₂ and Radon 222 were simultaneously recorded with no data gap.

In Sect. 2, we provide a comparison between observed and modeled vertical CO₂ profiles for both campaigns. The comparison is done on the CO₂ profile averaged for each campaign, but also for individual profiles. Two transport model (LMDZt, CHIMERE) and three biospheric flux models (CASA, SiB2 and ORCHIDEE) are tested within different combinations. In Sect. 3, we apply the Radon method for inferring CO₂ fluxes, and compare its results with the ones from the modeling method based on influence functions and flux maps. Both methods are compared for the CAATER 1 campaign, during a flight on 25 May 2001.

2 Comparison between observed and modeled CO₂ vertical profiles

We evaluate here vertical transport of CO₂ between the ABL and the FT in the LMDZt and the CHIMERE tracer transport models with vertical profile information from both CAATER campaigns, using land fluxes from CASA, SiB2 and ORCHIDEE for the global model LMDZt and from ORCHIDEE for the mesoscale model CHIMERE. All these models are described here below.

2.1 Model-data comparison set-up

The LMDZt model (Hourdin et al., 2006) is an offline transport model derived from the general atmosphere circulation model of the Laboratoire de Météorologie Dynamique LMDZ (Hourdin and Armengaud, 1999). In this version, LMDZt has a global grid, which is zoomed over Europe at horizontal resolution of 1° by 1°. It is parameterized with a diffusive and thermal turbulence convective boundary-layer scheme, and contains 38 vertical levels up to 3 hPa (between 0 and 4000 m). The transport simulation time-step is 1 h;

horizontal winds are nudged on the ECMWF analyzed fields (Filiberti et al., 2006; Uppala et al., 2005) with a time constant of 3 h, ensuring realistic synoptic CO₂ transport during each campaign (see Peylin et al., 2005; Geels et al., 2007; Patra et al., 2008). For optimal comparison with the CAATER aircraft data, the modeled CO₂ profiles are compared with observation exactly at the same time (± 1 h) and location.

The Eulerian mesoscale chemical transport model MM5-CHIMERE (Schmidt et al., 2001) is a three-dimensional atmospheric transport model primarily designed to make long-term simulations for emission control scenarios on air quality. The model domain used here covers Western Europe at a horizontal resolution of 50 km by 50 km. We use 20 layers in the vertical on terrain following sigma-coordinates, with seven layers in the lowest 300 m and the highest one around mid-troposphere. CHIMERE is an off-line model which requires mass-fluxes for transport calculations. These fluxes are provided by a run of the regional meteorological model MM5 (Grell et al., 1994) with output saved every six hours. MM5 is nudged towards the analyses of the European Centre for Medium Range Weather Forecasting (ECMWF) every six hours. The CHIMERE model is a regional model which consequently requires lateral and top boundary conditions, supplied by a run of the global transport model LMDZ (Law et al., 2008; Hauglustaine et al., 2004) at daily frequency. For further information, see the model server (<http://www.lmd.polytechnique.fr/chimere/>).

Surface fluxes prescribed globally to LMDZt are: 1) annual fossil fuel emissions from Andres et al. (1996), adjusted to the year of the campaigns. The use of annual emissions for fossil fuel may lead to an underestimate of the variability of simulated CO₂ concentrations (Peylin et al., 2009). Some efforts have been made in past years to produce time-varying fossil fuel inventories for Europe (Pregger et al., 2007). The validation of these products for all European countries is still ongoing and we decided not to use them in our work. 2) Monthly air-sea climatologic fluxes from Takahashi et al. (1999, 2002). And 3) Net Ecosystem Exchange CO₂ flux calculated for each campaign interval, with 3 different flux models: ORCHIDEE, SiB2 and CASA. The ORCHIDEE model (Krinner et al., 2005) simulations were forced by 1/2 hourly meteorological fields interpolated from ECMWF 6-hourly analysis at a resolution of $0.35^\circ \times 0.35^\circ$ for 2001 and 2002. Two other alternative Net Ecosystem Exchange (NEE) hourly flux maps have been prepared (although computed on year 2002 only) for the Transcom-continuous experiment (Law et al., 2008). These alternative NEE flux maps at resolution of $1^\circ \times 1^\circ$ each 3 h, are from SiB-2 (Sellers et al., 1996) and CASA (Randerson et al., 1997) models. It is interesting to use the SIB-2 and CASA data-oriented NEE as an alternative to the ORCHIDEE process-based model NEE, because the phenology of SIB-2 and CASA is driven by satellite greenness index observations during CAATER 2, whereas the one of ORCHIDEE is calculated from climate.

For each campaign, the mean profiles simulated by the chosen model combinations (LMDZt-ORCHIDEE, LMDZt-CASA, LMDZt-SiB2 and CHIMERE-ORCHIDEE) have been computed and compared to the observed mean profile at 100 m vertical resolution (Figs. 1 and 4). Note that as in the companion paper, the altitude has been normalized to the corresponding ABL height for all of the observed and modeled profiles before any averaging step. For each profile, the ABL height has been determined with a precision of ± 50 m as the altitude at which the vertical gradient of the potential temperature begins to decrease, and where CO₂ and H₂O present step-changes (Gerbig et al., 2003a). In addition, two typical profiles have been selected among the 14 sampled from each campaign to illustrate the performances of the transport model but also of the flux model on the simulations (Figs. 2 and 5). For each campaign, a correlation plot between the observations and LMDZt or CHIMERE simulations is provided (Figs. 3 and 6) to assess the impact of the transport model scale (global/mesoscale) on the reproduction of the CO₂ gradient between the boundary layer and the free troposphere.

2.2 Results for CAATER-1

On Fig. 1, we can observe that there are only small differences in the mean CO₂ profile between LMDZt coupled to any of the biospheric fluxes, and CHIMERE coupled to ORCHIDEE. We observe that the variability of CHIMERE-ORCHIDEE (2.5 ppm in the PBL, 2.2 ppm in the FT) is higher than from any LMDZt simulation (about 2 ppm in the PBL, 1.2 ppm in the FT). It is lower than the observed variability in the PBL (4 ppm) but higher than the observed variability in the FT (0.5 ppm) (Xueref-Remy et al., 2011). In addition, the mean value of the observed ABL-FT gradient, J , equals 8.9 ppm. The modelled value of J is 2.2 ppm for LMDZt-ORCHIDEE, 1.5 ppm for LMDZt-SiB2, 1.1 ppm for LMDZt-CASA, and 2.3 ppm for CHIMERE-ORCHIDEE (Fig. 1). Thus, the CHIMERE-ORCHIDEE and LMDZt-ORCHIDEE simulations are closer to observations than when using the two other flux models. However, the shape of the averaged profile is not well simulated. None of the model configurations can represent well the decrease of CO₂ observed in the mid-ABL, all being too diffusive.

To illustrate better the role of the transport model scale and of the fluxes, we selected two typical profiles (Fig. 2). The profile on Fig. 2a was recorded around 14:40 UTC on 26 May 2001 in East Germany, north of Oberpfaffenhofen (OBP). The wind was blowing from the west. CO₂ concentration is quite homogeneous in the boundary layer (~ 364 ppm), with a minimum higher than during the previous days meaning it is likely that the air has travelled over pollution sources and biospheric sinks whose signals have been mixed by convection (see Fig. 4 in Xueref-Remy et al., 2011). The ABL height is at 2500 m a.s.l., and a marked CO₂ gradient between the boundary layer and the free troposphere

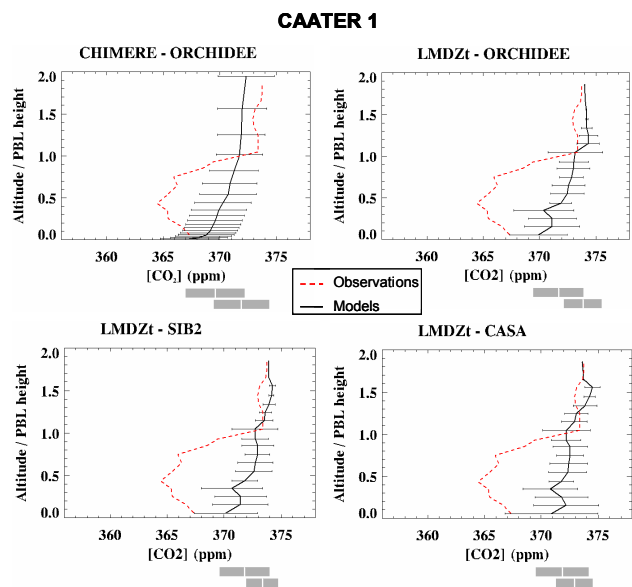


Fig. 1. Comparison of the mean modeled and observed profiles for CAATER 1. Horizontal bars represent the $\pm 1\text{-}\sigma$ standard deviation of the mean, computed every 1/10th of the altitude/ABL height ratio. Below the x-axis, the global mean and the $\pm 1\text{-}\sigma$ variability in the ABL (upper bar) and FT (lower bar) are shown according to the CO₂ concentration scale given by the x-axis.

is observed ($J = 9.5$ ppm). The simulations show that: 1) independently of the fluxes, the simulations with LMDZt give a too-smooth profile with a low boundary layer height (around 700 m a.s.l.) resulting into a jump J comprised between 1.1 ppm (Sib2, CASA) and 2.1 ppm (ORCHIDEE); and 2) the simulation with CHIMERE-ORCHIDEE is better than LMDZt-ORCHIDEE in terms of shape, although not perfect as it produces a decrease in CO₂ below the top of boundary layer as seen on observations, but also in terms of jump (~ 4 ppm). The influence of biospheric fluxes is rather small, indicating that profiles can evaluate transport properties. In addition, the mesoscale model CHIMERE captures the large CO₂ increase across the top of the boundary-layer better than the global model LMDZt.

The profile on Fig. 2b was recorded over OBP around 15:30 UTC on 26 May 2001, with wind blowing from the North. Here as well, the minimum (~ 364 ppm) is not as high as during the previous days (see Fig. 4 in Xueref-Remy et al., 2011). The ABL height is located at 2400 m a.s.l. and the cross ABL vertical CO₂ gradient is quite well marked ($J = 8.2$ ppm). Figure 2b shows that: 1) the LMDZt simulations underestimate the top of the boundary layer (~ 900 m a.s.l.) compared to observations but are quite sensitive this time to the surface fluxes, with $J = 1.1$ ppm, 3.5 ppm and 6.2 ppm for CASA, Sib2 and ORCHIDEE, respectively; and 2) the shape of the profile simulated by CHIMERE-ORCHIDEE is more realistic than with LMDZt-ORCHIDEE, with a

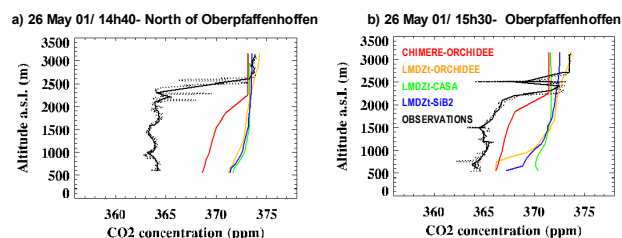


Fig. 2. Comparison between observed and modeled profiles for 2 case studies during the CAATER 1 campaign.

boundary layer height located at ~ 2100 m a.s.l. close to the observed value. The simulated gradient $J = 4.1$ ppm with CHIMERE-ORCHIDEE is lower than the observed one and lower than with LMDZt-ORCHIDEE. The LMDZt-ORCHIDEE combination has best performances among the three sets of biospheric fluxes to simulate the observed gradient. The bias of this simulation seems to come from the fact that the boundary layer height is simulated at too low a level. The CHIMERE-ORCHIDEE simulation is the closest to the observations, as the mesoscale model manages to reproduce (even if not strongly enough) the structure of the profile. Whatever the transport model, ORCHIDEE gives the best results among the three biospheric fluxes tested.

Figure 3 provides a model vs. observed scatter plot of J for the 14 profiles of CAATER-1 for the LMDZt-ORCHIDEE and CHIMERE-ORCHIDEE couples (ORCHIDEE being identified as the best NEE model for CAATER-1). The modeled J value of LMDZt-ORCHIDEE is weakly correlated with the observed value ($R^2=0.07$, slope=0.13) and is also less variable across profiles. Although the sign of the J is correctly modelled for all the profiles, its magnitude is underestimated. This indicates that modeled vertical transport during CAATER-1 is too vigorous in LMDZt. In particular, the ABL height is not marked at all in LMDZt, opposite to the sharp CO₂ discontinuity observed in the aircraft profiles. The CHIMERE-ORCHIDEE simulation of J is not better correlated with the observations ($R^2 = 0.07$, slope = 0.16). Analysis of individual profiles (not shown) reveals that CHIMERE tends to do slightly better than LMDZt while underestimating J .

2.3 Results for CAATER-2

Figure 4 shows a comparison between observations and model simulations for CAATER-2. As opposed to CAATER-1 (see Fig. 1), here we can observe that the average LMDZt profile is sensitive to the biospheric fluxes, especially near the surface where there is a depletion of CO₂ due to net plant uptake. CHIMERE-ORCHIDEE does not simulate the mean CO₂ vertical profile better than LMDZt. Variability in CHIMERE-ORCHIDEE is higher than in both LMDZt and the observations, as seen by the high standard deviation of the 100 m resolution profile (reaching 9 ppm in the

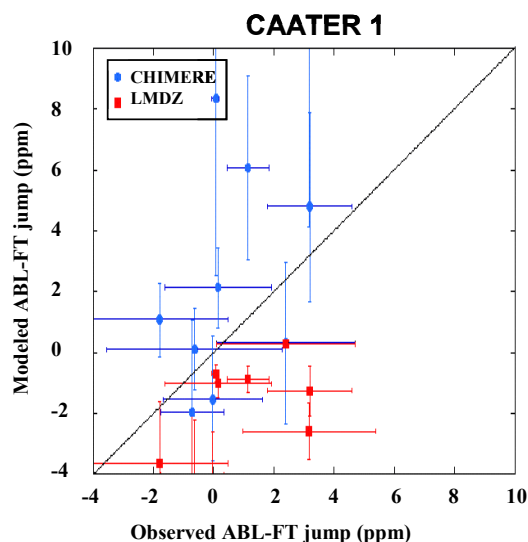


Fig. 3. Comparison of the modelled ABL-FT jumps from LMDZ-ORCHIDEE and CHIMERE-ORCHIDEE to the observed jumps for CAATER 1. Points represent the mean jump for each profile and bars represent the associated $\pm 1\text{-}\sigma$ standard deviation.

lowest levels, compared to an observed value of 4.3 ppm (see Fig. 11 in Xueref-Remy et al., 2011). Furthermore, let us recall the observed ABL-FT gradient, $J=+0.8$ ppm; compared to $J=-0.3$ ppm in LMDZt-ORCHIDEE, -1.8 ppm in LMDZt-SiB2, -1.1 ppm in LMDZt-CASA, and 3.2 ppm in CHIMERE-ORCHIDEE (Fig. 4). Thus, all LMDZt simulations give negative values of J , unlike in the observations. By contrast, the CHIMERE-ORCHIDEE simulated J value is positive as is the observed one. However, LMDZt-ORCHIDEE and LMDZt-CASA combinations are best at reproducing the variability observed in the PBL. Even if having opposite signs, the gradient J is small in both cases, as in the observations. The LMDZt-ORCHIDEE can be selected as the best simulation in terms of jump and profile structure, closely followed by the LMDZt-CASA coupling.

Figure 5 shows two typical profiles to evaluate the effect of transport model scale and of NEE. Figure 5a profile has been recorded above the ORL site, at 11:15 UTC on 2 October 2002. A southerly wind was blowing (see Fig. 5 in Xueref-Remy et al., 2011). The ABL top was observed at 750 m a.s.l., and the gradient J was 1.5 ppm. In the ABL, CO₂ varied between 368 ppm and 377 ppm (likely a mixture of vegetation uptake, respiration transport and anthropogenic sources), while it was 372.5 ppm in the free troposphere. All the LMDZt simulations show an accumulation of CO₂ near the ground that is not observed. The model-data misfit is independent of the underpinning NEE flux model, giving a negative $J\approx 3.5$ ppm. Also, the ABL height is too high in the simulation (around 1450 m a.s.l.). By contrast, the CHIMERE-ORCHIDEE simulation represents quite well the homogeneous vertical profile in the free troposphere. How-

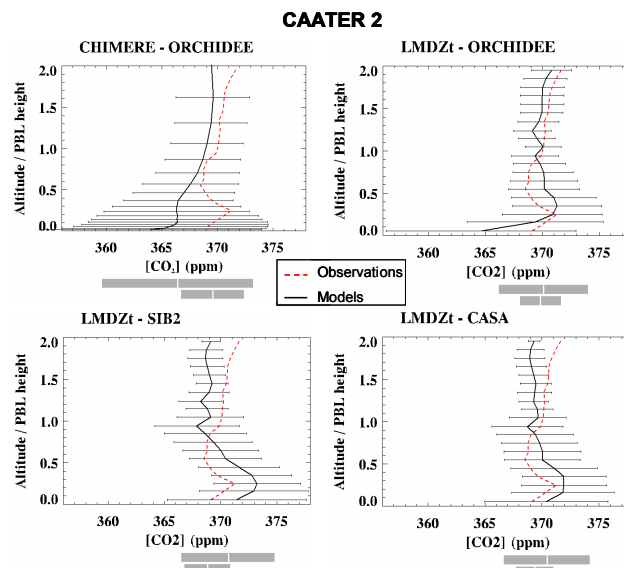


Fig. 4. Comparison of the mean modeled and observed profiles for CAATER 2. Horizontal bars represent the $\pm 1\text{-}\sigma$ standard deviation of the mean, computed every 1/10th of the altitude/ABL height ratio. Below the x-axis, the global mean and the $\pm 1\text{-}\sigma$ variability in the ABL (upper bar) and FT (lower bar) are shown according to the CO₂ concentration scale given by the x-axis.

ever, the CO₂ profile in the ABL has an opposite shape to the observed one, the model giving an accumulation of CO₂ near the ground, followed by an inversion of the CO₂ gradient near the observed boundary layer height.

The second case study profile in Fig. 5b was recorded over Thüringen, eastern Germany at 10:00 UTC on 3 October 2002. A west/south-westerly wind was blowing (see Fig. 5 in Xueref-Remy et al., 2011). The ABL height was found at 550 m a.s.l., with a large negative gradient $J=-9.1$ ppm. The data seem to contain an influence by local pollution in the lowest levels, with a maximum of CO₂ reaching 385 ppm. LMDZt prescribed by ORCHIDEE, SiB2 and CASA NEE give distinct profiles. The three NEE flux models produced a negative ABL-FT gradient ($J=-4.5$ ppm in SiB2, $J=-5.1$ ppm in CASA, $J=-10.9$ ppm in ORCHIDEE) as in the observation. The ABL height is simulated too high as in the previous case (around 1300 m a.s.l. for SiB2 and CASA, and 1500 m a.s.l. for ORCHIDEE). But, indeed, if the ABL height was simulated properly in LMDZt-CASA and LMDZt-ORCHIDEE runs, the CO₂ maximum in the ABL would be about 2.3 higher and would match the observed profile quite well (see inset in Fig. 5b). The CHIMERE-ORCHIDEE simulation leads to a CO₂ profile oscillating around the observations, too variable and with a very small J value making the ABL height hard to define from CO₂. In this case, the parameterization of the model is not diffusive enough.

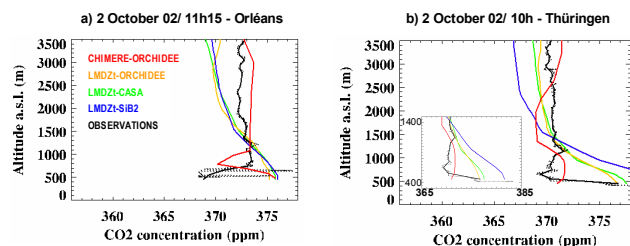


Fig. 5. Comparison between observed and modeled profiles for 2 case studies during the CAATER 2 campaign.

Figure 6 provides a model vs. observed scatter plot of J for LMDZt-ORCHIDEE that we have identified to be the best combination for LMDZt and NEE model in CAATER-2 as for CAATER-1, and for CHIMERE-ORCHIDEE. A similar bias in LMDZt is observed resulting in J values that are too small and not variable enough, as is the case for CAATER-1. However, the model low bias is smaller than for CAATER-1, and the modelled vs. observed linear regression slope of J is better defined ($R^2=0.3$, slope=0.6) indicating that the model captures different J better. The CHIMERE-ORCHIDEE tends to overestimate the J values with too large a sensitivity ($R^2=0.3$, slope=1.3).

2.4 Discussion

One finding of this comparison is that the NEE flux magnitude may occasionally play a role in determining the magnitude and the sign of the ABL-FT gradient J , and the shape of the CO₂ vertical profiles (see for instance the profile on Fig. 2b). Among the three NEE models tested, ORCHIDEE (a climate driven model) gives the best results compared to SiB2 and CASA (process-based models); but let us recall that ORCHIDEE has been prepared for both 2001 and 2002, while CASA and SiB2 only for 2002. However, errors in model transport seem to be the most frequent cause of mismatch with observations. This demonstrates that vertical profiles can be used efficiently as a constraint to falsify model transport. This was shown for instance by Stephens et al. (2007) for monthly profiles at various sites around the globe, and we here confirm the results of this global study using the two CAATER intensive campaigns. The fact that the value of J differs strongly among the profiles during the same campaign is important to outline, because it suggests that the CAATER airplane trajectory sampled a diversity of flux-transport situations that can be used to cross-validate LMDZt and CHIMERE.

When the LMDZt model global results are contrasted with CAATER-1 data, in the lowermost atmosphere between 0 and 4000 m, the model shows *consistently* the bias of simulating too stiff vertical profiles. This points to an overestimation of the mixing rate between the ABL and the FT. It is possible that the entrainment zone and the non-mixing zone

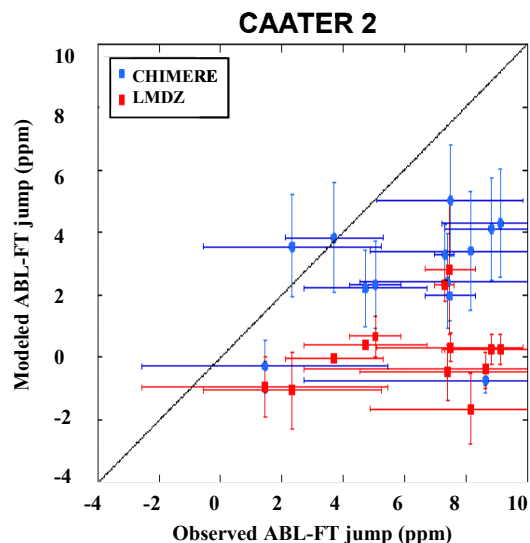


Fig. 6. Comparison of the modelled ABL-FT jumps from LMDZ-ORCHIDEE and CHIMERE-ORCHIDEE to the observed jumps for CAATER 2. Points represent the mean jump for each profile and bars represent the associated $\pm 1-\sigma$ standard deviation from observations and models.

at the top of the ABL are not well resolved by the model parameterization when convection is established (Gibert et al., 2007) (Hourdin et al., 2002). An overestimation of ABL-FT mixing by transport models was already shown by Yi et al. (2004) using CO₂ vertical profiles along the WLEF tall tower in Wisconsin, and by Ramonet et al. (2002) using ABL aircraft vertical profile data during an intensive campaign over a forest in Russia. In CAATER-2, the LMDZt model results show that the boundary layer upper level is systematically too high. It may be noted that the model is unable to reproduce a shallow boundary layer and is too diffusive. At a large scale, the cross-validation analysis of global transport models by Stephens et al. (2007) also pointed to an overestimation of simulated vertical transport in summer, just as we found for the LMDZt model in CAATER-1 and CAATER-2. But the Stephens et al. dataset was more related to evaluation of vertical mixing in the mid troposphere, driven by cloud transport and frontal activity than the CAATER dataset.

The CHIMERE model results including the underlying dynamical fields of MM5 are globally better than those of LMDZt, as the vertical transport simulated by CHIMERE is less diffusive. Even if not perfect, CHIMERE often improves the representation of vertical structure of the profiles (see for example profiles from Fig. 2). However, the modeled CO₂ profile is too variable at times leading to unrealistic behavior such as seen in Fig. 5b.

Although the mesoscale model CHIMERE appears to be better able to reproduce CO₂ vertical variability than the global model LMDZt, this kind of study should be extended to more data, and more synoptic situations. It underlines the

strong need for more aircraft campaigns to generate vertical profiles to help to calibrate vertical transport parameterization in models and to better constrain CO₂ flux calculation through inverse modeling (Stephens et al., 2007).

3 Regional CO₂ flux calculation

We estimate here regional CO₂ fluxes using two independent methods. The first method, based on collocated CO₂ and Radon 222 observations, is called “the Radon method”. The second method combines influence function of the measurement to surface fluxes calculated by LMDZt with an a priori distribution of surface fluxes. Both methods are applied to a case-study during the CAATER-1 campaign and the results are compared to each other.

3.1 Flux calculation with the Radon method

Simultaneous ²²²Rn and CO₂ concentration observations allow the inference of unknown CO₂ surface fluxes, assuming known ²²²Rn fluxes within the hypothesis on the ²²²Rn flux distribution. This dual tracer method, where the concentration change of tracer is scaled to the other in proportion of their surface fluxes, has been applied to ground-based observatories time series (Biraud et al., 2000, 2002; Levin, 1984, 1999; Schmidt et al., 1996, 2001; Wilson et al., 1997). Radon 222 is a radioactive noble gas with a half-time of 3.8 days, that is emitted at relatively constant rates by soils, while the flux from the ocean surfaces is negligible. ²²²Rn emitted by soils is transported by winds and reduced by radioactive decay. We make the (reasonable) hypothesis that the surface flux of ²²²Rn is uniform and constant (but we take into account its variability for error calculation, see below) in order to infer less well-known continental emissions of other compounds. According to Schmidt et al. (2001), CO₂ fluxes can be calculated using Eq. (1) below, on condition that the correlation factor between Radon 222 and CO₂ data is better than 0.5 (Levin et al., 1999). The calculated CO₂ fluxes is thus expressed as:

$$j_{\text{CO}_2} = j_{\text{Rn}} \Delta C_{\text{CO}_2} / \Delta C_{\text{Rn}} [1 + \lambda_{\text{Rn}} C_{\text{Rn}} / (\Delta C_{\text{Rn}} \Delta t)]^{-1} \quad (1)$$

In Eq. (1), ΔC_{CO_2} and ΔC_{Rn} are the species spatial gradients between the concentration at the measurement location and the marine boundary layer (MBL) concentration, this latter being taken as a baseline, on the day and at the latitude of the measurement; λ_{Rn} is the radioactive half-time of ²²²Rn (3.814 days), j_{Rn} is the surface ²²²Rn flux influencing the airborne observation and Δt defines a transit time of air parcel from emission to the observation site. Note that in other papers (e.g. Biraud et al., 2000, 2002; Schmidt et al., 2001, 2003) Δ stands for temporal, not spatial, gradients.

In the following, we first explain the Radon instrumentation and then apply the Radon method to a case study.

3.1.1 Semi-continuous Radon-222 instrumentation

Radon has been measured with the AVIRAD instrument (Filippi, 2000), which consists of an isokinetic probe fixed to the fuselage and of a filtration unit located inside, with limited and straight tubing between the probe and the filter. The isokinetic probe was built by Sextant Avionique Corporation using the same design as for the NASA C-141 (Kritz et al., 1998). The radon measurements are made with alpha spectrometry of the radon progeny products deposited on the collected aerosols. The Paper Filtering Unit is provided with 4 Si detectors in order to measure the alpha activity of each sample at 4 successive decay times. The Data Acquisition System runs 4 alpha spectrometers allowing implementation of different methodologies to validate the radon concentration data. The probe includes a stagnation reservoir, the null-type air inlet and a flow line sensor. It was mounted on the cargo window, under the airplane. At the location of the probe, the boundary layer of the plane was expected to be less than 10 cm. Thus in order to keep the air inlet beyond the boundary layer of the plane, it was fixed at the tip of a 29-cm long mast. The null-type nozzle operates by measuring static pressure on the outside of the probe nozzle, and static pressure inside the inlet opening of the nozzle. When zero pressure differential is developed between the inside and outside pressure taps, the isokinetic velocity is obtained. This null differential pressure is automatically adjusted in real-time. The null-type nozzle was calibrated in a wind tunnel for the Mach number range of the aircraft, between Mach number 0.6 and 0.7. At the rear part of the air inlet, a flow-line sensor made of 4 dynamic pressure taps is used during test flights to make sure that the angle of attack of the isokinetic nozzle was as close as possible to zero. The flow line sensor was also calibrated in a wind tunnel for the Mach number range of the aircraft. Except during turns and turbulences, the angle of attack of the probe was less than 2 degrees, even during ascents and descents. The Filter Unit used a continuous paper filter strip which is advanced by a motorised take-up spool at programmable time intervals. To preserve the collected samples, a blank strip is rolled up along with the filtering strip on the take-up spool. Different paper filter media can be used depending on subsequent analytical procedure. The number of samples per spool is about 70. The active area of each sample is 1700 mm². The precision of the measurement is 30%.

3.1.2 Case study

The uncertainties associated with Eq. (1) are the following. First, we suppose a constant and uniform ²²²Rn flux. In reality this flux depends on soil bedrock type, total pore space, tortuosity, soil moisture and precipitation. Its mean variability in Western Europe soils is of the order of 30% (Nazaroff, 1992; Jutzi, 2001; Ielsch et al., 2002; Szegvary et al., 2007). Secondly, the ²²²Rn measurement precision itself is ~30%

which translates into a relative error of the same magnitude in the inferred CO₂ surface flux. Thirdly, the error on the transit time is of the order of 6 h, that is a 5% error on the inferred CO₂ flux. In total, we estimate the error on the CO₂ flux of Eq. (1) to be 35%. This is similar to the uncertainty estimated in Schmidt et al. (2003).

3.2 Flux calculation using influence functions and map fluxes

Although back-trajectories are useful tools to trace the origin of air masses, they do not allow a quantitative determination of the influence of surface flux on the atmospheric CO₂ concentration. Here, we use influence functions (IF) calculated by backward transport in the LMDZt model (Hourdin et al., 2006) to quantitatively link surface fluxes and aircraft-measured concentrations. Briefly, a mass of inert tracer is emitted at each aircraft measurement location and transported backward in time using the LMDZt 3-D dynamical fields, backward transport being an analog of the adjoint of the transport. This resulting influence function (IF) to surface fluxes quantifies the contribution of each surface grid point to a given measurement. The IF is the potential sensitivity of the measured concentration to surface fluxes (e.g. Lauvaux et al., 2009; Krol et al., 2003; Stohl et al., 1998a, b, c). Indeed, even if the vertical transport parameterizations have shown some weaknesses in global transport models such as LMDZ, synoptic transport in LMDZt is proven to be quite well-represented (Patra et al., 2008). We combined IF, as calculated by LMDZt, with surface flux maps in order to estimate CO₂ fluxes influencing the airborne observations, assuming that the flux model is already a realistic image of the flux. IF are computed for five days backwards (corresponding to the Δt from Sect. 3.1) at the observing point corresponding to event D in Fig. 7a, starting at noon, which is the time when the depletion of CO₂ started to occur. Note that on Fig. 8a, IF are only shown for days 1–3 backwards, as for days 4–5, the signals are less than 1% of the maximum sensitivity. The surface flux (Fig. 8b) is the sum of a priori fluxes described in Sect. 2: air-sea flux from Takahashi et al. (1999, 2002), fossil fuel emissions from Andres et al. (1996) and NEE from ORCHIDEE (Krinner et al., 2005) as this was the model that gave the best results among the 3 biospheric models tested in Sect. 2. By multiplying the IF by this a priori flux map, we infer the CO₂ flux (Fig. 8c) which influences the airborne observations to be $-4.39 \text{ gC m}^{-2} \text{ day}^{-1}$. NEE contributes dominantly for 73.2% of the total flux as a sink ($-6.91 \text{ gC m}^{-2} \text{ day}^{-1}$), fossil fuel emissions for 26.8% as a source ($+2.53 \text{ gC m}^{-2} \text{ day}^{-1}$). The ocean contribution is a sink less than 0.01% of the total flux.

Although LMDZt has been proven to model transport at the synoptic scale quite well, the dynamics are not perfect (Patra et al., 2008; Geels et al., 2007). Furthermore, the method here relies on a flux set that has not been optimized. Thus, there are two sources of errors in the method: transport

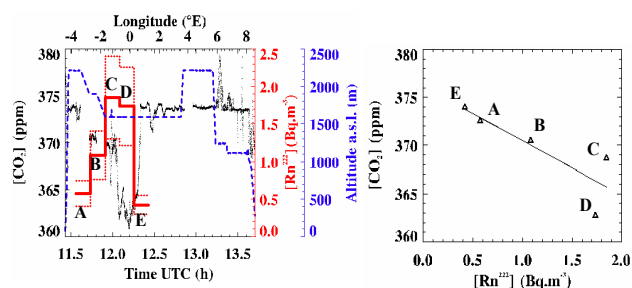


Fig. 7. Left panel: Radon concentration (red plain line) with measurement precision (red dashed line), CO₂ concentration (black plain line) and altitude (blue dashed line) concentrations vs. time measured during CAATER 1 on 25 May 2001. The longitude is indicated on top. Jumps in Radon and CO₂ concentrations are defined as events A to E. Right panel: CO₂ vs. Radon 222 concentration for events A to E.

uncertainties and flux veracity. Indeed, the method allows an estimation of the flux distribution that influences the observations, rather than a real optimization of the flux. Assessing the differences between observed and modeled CO₂ concentrations, we can attempt a rough estimate of the error in the method. Figure 9 represents a comparison between the time-series of the observed CO₂ concentration along the aircraft path, and altitude/time cross sections of the LMDZt/flux set simulation. Three main points can be highlighted:

1. The simulation represents mainly two air masses in the PBL: one during the first half of the flight, with concentrations in the range of 368–370 ppm typically representative of a mixture of oceanic and biospheric air lower than the marine boundary layer (MBL) background value (374.5 ppm: see Xueref-Remy et al., 2011). And a second one during the second half of the flight, with higher concentrations of about 373 ppm close to the MBL background concentration typical of oceanic air-masses, with some peaks at 380 ppm indicating an enrichment of air masses in CO₂ due to fossil fuel emissions from the Benelux and the Ruhr regions. This is in agreement with the back-trajectories analysis conducted in the companion paper (Xueref-Remy et al., 2011).
2. The match between the median amplitudes (and not the mean ones!) of the observed concentrations and the modeled ones is quite good, with an observed median amplitude of about 6 ppm versus a simulated median amplitude of about 4 ppm. This leads to an underestimation of about 33% by the model framework.
3. The model framework does not reproduce concentration extremes such as the observation depletion during the Radon episode D around 12:10 UTC, and the peak of CO₂ of 380 ppm observed over the Ruhr area around 13:15 UTC. The amplitude between extremes is

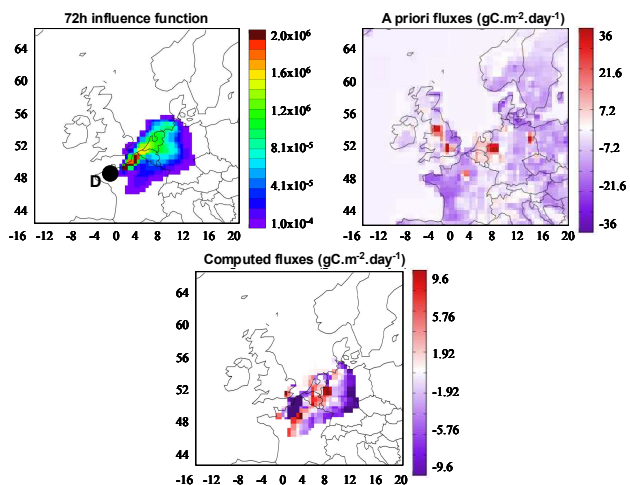


Fig. 8. Top left panel: LMDZt influence function (IF) calculated for 3 days backwards from the observing point (dark circle) corresponding to event D in Fig. 7a. Top right panel: a priori fluxes (anthropogenic, biospheric and oceanic), summed and averaged for the period from 21 May 2001, 12:00 UTC to 25 May 2001, 12:00 UTC. Low panel: Weighted fluxes contributing to the CO₂ signal recorded at the observing point. Flux units are in $\text{gC m}^{-2} \text{day}^{-1}$.

3.5 ppm from the model versus 11.5 ppm from observations, therefore the model framework underestimates extremes by a factor of 3 roughly.

So we conclude that the error on this method can be large for individual events and is more than 33% when considering averages.

3.3 Discussion

Both methods give a comparable CO₂ flux of the same order of magnitude within the Radon method uncertainty (35%), the error on the modeling method estimated to be at least 33%. This flux is the one seen at 1700 m, representative of the area over which the air masses travelled before reaching the observing point. The 5-days backward IF covers the North of France, The Benelux, The Netherlands, Germany, Western Poland and the Czech Republic. But most of the surface grid elements are concentrated over Northern France, Germany, The Benelux and Western Poland. The catchment area is of the order of 1500 km (longitudinally) per 700 km (latitudinally). This estimates the flux at this moment and for this region of Europe to be a net negative flux of -4.2 to $-4.4 \text{ gC m}^{-2} \text{day}^{-1}$. The concordance between the ²²²Rn method and the IF method relying on an explicit description of transport dynamics is encouraging. It suggests that, even if not optimized, the fluxes prescribed to LMDZt are rather realistic. In fact, the IF method can be applied to any observation point to estimate the footprint of the air mass before it reaches the measurement location. The flux scale is a function of the dynamical fields, of the time backwards and of

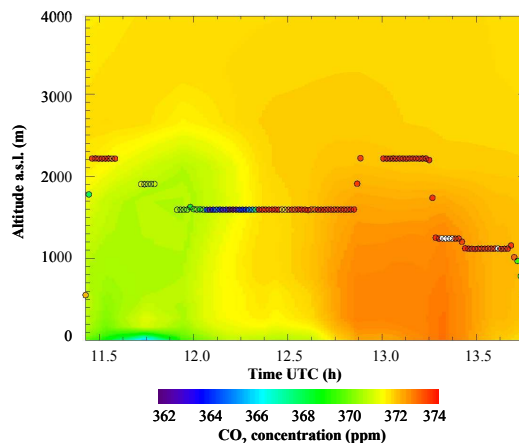


Fig. 9. Comparison of CO₂ concentration along the aircraft path on 25 May 2001 according to altitude and time (circles), with simulated CO₂ concentration fields from the LMDZt modeling framework. Note that the white color indicates high concentrations that are out of scale (~ 380 ppm).

the altitude of the observation point. The catchment area increases with altitude and can be defined using the IF maps (for example, for the profile done in the flat region of Brest, we get a fetch of $50 \times 50 \text{ km}^2$ at 70 m a.s.l., $500 \times 500 \text{ km}^2$ at 900 m of altitude and $1200 \times 700 \text{ km}^2$ at 1500 m). Thus, the knowledge of IF should help to fill the gap between local and continental scale for carbon flux calculations on the continents. Of course, this paper only shows one case study and deeper studies must be conducted to better characterize the errors on the method, here estimated to be at least 33%. The use of optimized fluxes (from atmospheric inversions) and of mesoscale models for atmospheric transport may significantly reduce the uncertainties.

4 Conclusions

In this paper we have conducted a comparison between observations and modeling atmospheric CO₂ studies for the airborne CAATER campaigns that occurred over Europe in May 2001 and October 2002, as described in the companion paper (Xueref-Remy et al., 2011).

Firstly, we compared CO₂ modeled and observed vertical profiles using different combinations of transport models (the global model LMDZt and the mesoscale model CHIMERE) and biospheric flux models with a CO₂ diurnal cycle (CASA, SiB2 and ORCHIDEE). For the CAATER-1 campaign, the observed mid-ABL gradient of CO₂ is not reproduced by any of the tested model combinations (LMDZt-CASA, LMDZt-SiB, LMDZt-ORCHIDEE and CHIMERE-ORCHIDEE), all being too diffusive. For the CAATER-2 campaign, CHIMERE-ORCHIDEE is closer to observations concerning the jump in CO₂ concentration between

the ABL and the FT, but the profile structure is better reproduced by LMDZt-CASA. However LMDZt is still too diffusive and the ABL height is not well located. On the whole, we can conclude that: (1) NEE fluxes sometimes play a role in the gradient magnitude and shape, the ORCHIDEE model (a climate driven model) giving the best results compared to SiB2 and CASA (process-based models); (2) however, mismatches between observed and modeled profiles mainly come from errors in the transport models. In fact, LMDZt systematically simulates vertical profiles that are too stiff, overestimating the mixing rate of the ABL into the FT. This conclusion was also given by Stephens et al. (2007) who reported a systematic overestimation of simulated vertical transport in Summer by 12 models from the TRANSCOM 3 Level 2 study (Gurney et al., 2004). During CAATER 2, the ABL height is generally too high in that model. The CHIMERE mesoscale model gives better results on average, as the vertical transport is less diffusive and the jumps are better reproduced. However CHIMERE is sometimes too variable, leading to incoherent structures. On the whole, the mesoscale model in this case study seems better than the global one in reproducing vertical profiles. Finally, this work highlights the fact that more intensive and regular vertical profiles are needed in the future to conduct further comparisons between observations and models, and thus to make important progress in the parameterization of the models.

Further, we coupled influence functions (IF) and CO₂ map fluxes to compute the CO₂ flux observed at a given point. We compared the results of this modeling method to CO₂ flux calculated with the Radon method from simultaneous CO₂ and Radon 222 measurements. Both methods were applied to a CAATER-1 campaign case study, during which a good correlation between in situ CO₂ measurements and semi-continuous Radon 222 observations was observed. Using IF from LMDZt (for which synoptic transport is known to be quite reliable: see Patra et al., 2008) we estimated the catchment area of the observation point (located at 1700 m a.s.l.) to be 1500 km (longitudinally) by 700 km (latitudinally) over Northern France, Benelux, Germany and Western Poland. Both methods give a CO₂ flux of the same order of magnitude (−4.2 to −4.4 gC m^{−2} day), within the uncertainty of the Radon method (35%), the uncertainty of the modeling method being estimated at more than 33%. The agreement between the results of both methods is very promising for future application of the modeling method on any observation point. However, errors in this second method need to be better assessed, for example at tall towers where simultaneous CO₂ and Radon 222 measurements can be conducted. Uncertainties may also be significantly reduced by using optimized fluxes (from atmospheric inversions) and mesoscale models for atmospheric transport.

Acknowledgements. This work was funded by the Institut National des Sciences de l'Univers, France, and by the European Commission in the framework of the AEROCARB project. We thank Dominique Filippi for providing the Radon data. We acknowledge Ingeborg Levin and Martina Schmidt for helpful discussions on the Radon calculations. The authors are very grateful to Nicolas Viovy for providing ORCHIDEE fluxes. We thank very much Mary Minnock for text corrections.

Edited by: J. Rinne



The publication of this article is financed by CNRS-INSU.

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