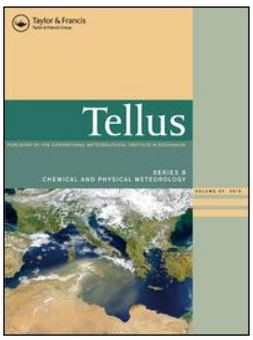


Tellus B: Chemical and Physical Meteorology

ISSN: (Print) 1600-0889 (Online) Journal homepage: http://www.tandfonline.com/loi/zelb20

Short-term variations in atmospheric CO_2 at Ny-Ålesund, Spitsbergen, during spring and summer

Magnuz Engardt, Kim Holmén & Jost Heintzenberg

To cite this article: Magnuz Engardt, Kim Holmén & Jost Heintzenberg (1996) Short-term variations in atmospheric CO_2 at Ny-Ålesund, Spitsbergen, during spring and summer, Tellus B: Chemical and Physical Meteorology, 48:1, 33-43, DOI: 10.3402/tellusb.v48i1.15664

To link to this article: https://doi.org/10.3402/tellusb.v48i1.15664

Published online: 18 Jan 2017.

Tellus (1996), 48B, 33–43
Printed in Belgium – all rights reserved

TELLUS
ISSN 0280–6509

Short-term variations in atmospheric CO_2 at Ny-Ålesund, Spitsbergen, during spring and summer

By MAGNUZ ENGARDT[1,*], KIM HOLMÉN[1] AND JOST HEINTZENBERG[2],
[1] *Department of Meteorology, Stockholm University, S-106 91, Stockholm, Sweden;*
[2] *Institute for Tropospheric Research, Permoserstrasse 15, D-04303 Leipzig, Germany*

(Manuscript received 10 April 1995; in final form 29 August 1995)

ABSTRACT

Results from the continuous measurements of atmospheric carbon dioxide performed at Ny-Ålesund, Spitsbergen are presented. The results are discussed with an emphasis on day-to-day variations during spring and early summer. During all years studied, significant negative anomalies ("dips"), lasting for several days, have been observed in the data from late May through July. The timing coincides with a draw down of dissolved carbon dioxide and nutrients in the surface waters of the North Atlantic observed by others. By using 3-dimensional trajectories, we follow the history of the air arriving in Ny-Ålesund, and show that the air, depleted in CO_2, had been in contact with these waters. Combining the trajectories with a box model yields a simple Lagrangian model, and we demonstrate that the timing and magnitude of the dips are consistent with the degree of CO_2 saturation of the sea during April to mid-June. In late June and July, the model indicates that a significant portion of the dips must have other causes, e.g., CO_2 uptake in the terrestrial biosphere further south.

1. Introduction

Man is increasing the atmospheric content of carbon dioxide, and since the gas is radiatively active in the long-wave part of the spectrum, concern is raised about possible climate changes. A key issue is therefore to predict the future CO_2 content in the atmosphere. Much work has been performed by the scientific community towards understanding the global carbon cycle, yet the budget of atmospheric CO_2 is still uncertain. We can measure the increase of the atmospheric burden with great accuracy and through economic statistics, we are well informed about how much CO_2 is emitted into the atmosphere each year as a result of fossil fuel burning and cement production. By subtracting the observed increase of atmospheric CO_2 from the anthropogenic emissions, we get a net sink for anthropogenic CO_2, which can be split into 2 main terms. A change in the terrestrial biomass, and a flux of CO_2 into the oceans as a result of the increased pCO_2 difference across the air-sea boundary. The 1st term is uncertain even to its sign. While it is believed that deforestation in the tropics is a source for atmospheric CO_2 (Houghton et al., 1987; Dixon et al., 1994), there seems to be an enhanced accumulation of carbon in other terrestrial systems of the world (Sedjo, 1992; Kauppi et al., 1992; Plantinga and Birdsey, 1993; Kolchugina and Vinson, 1993).

Present quantifications of the sink processes are unable to accommodate all of the CO_2 that we know leaves the atmosphere (Watson et al., 1990; Sarmiento and Sundquist, 1992), and we are left with an unidentified sink, with the magnitude of approximately 30% of the fossil fuel emissions. The future fate of the missing carbon is quite different, depending on whether it enters the oceans, the terrestrial biosphere or the organic carbon in the soils. Once in the deep ocean it will stay there

* Corresponding author.

for centuries, while increased amounts of carbon in the terrestrial biosphere is highly susceptible to changes in climate and other environmental or human factors, and can thus rapidly be returned to the atmosphere. Since the magnitude of the sink terms, and consequently also the fraction of the anthropogenic emissions that accumulates in the atmosphere, might change in a future climate, it is important to determine their current size, and also understand the mechanisms controlling their amplitude.

Debate is lively on whether the unidentified sink is to be sought in the sea or on land, but by combining modelling efforts and atmospheric measurements, the general consensus is that a major part of the sink must be located at the high latitude Northern Hemisphere (Keeling et al., 1989; Tans et al., 1990; Enting and Mansbridge, 1991; Sundquist, 1993; Conway et al., 1994). The northern North Atlantic including the Barents, Norwegian, Greenland, Iceland, and Irminger Seas, see Fig. 1, is potentially an area of great importance in regulating the global CO_2 fluxes between the atmosphere and the oceans. The North Atlantic current flows north into the region and, as the water is cooled, its CO_2 partial pressure (pCO_2) is decreased, thereby increasing the potential for the air-sea flux of CO_2. The North Atlantic deep water (NADW) formation is believed to carry large amounts of anthropogenic CO_2 into the abyss (Broecker and Peng, 1992). The waters flowing south across the Greenland-Iceland-Faroe-Scotland sill are a significant contributor to the NADW formation and they acquire some of their CO_2 characteristics within the Greenland, Iceland and Norwegian Seas. Sea-surface concentrations of CO_2 at high latitudes are furthermore strongly modulated by biological processes during phytoplankton bloom episodes. As a result, pCO_2 in the surface waters is highly variable both in time and space (Kempe and Pegler, 1991; Watson et al., 1991), and extremely low CO_2 levels are regularly observed in the waters around Iceland during late May and early June (Takahashi et al., 1993) which must lead to a large local flux of CO_2 into the oceanic mixed layer during this period.

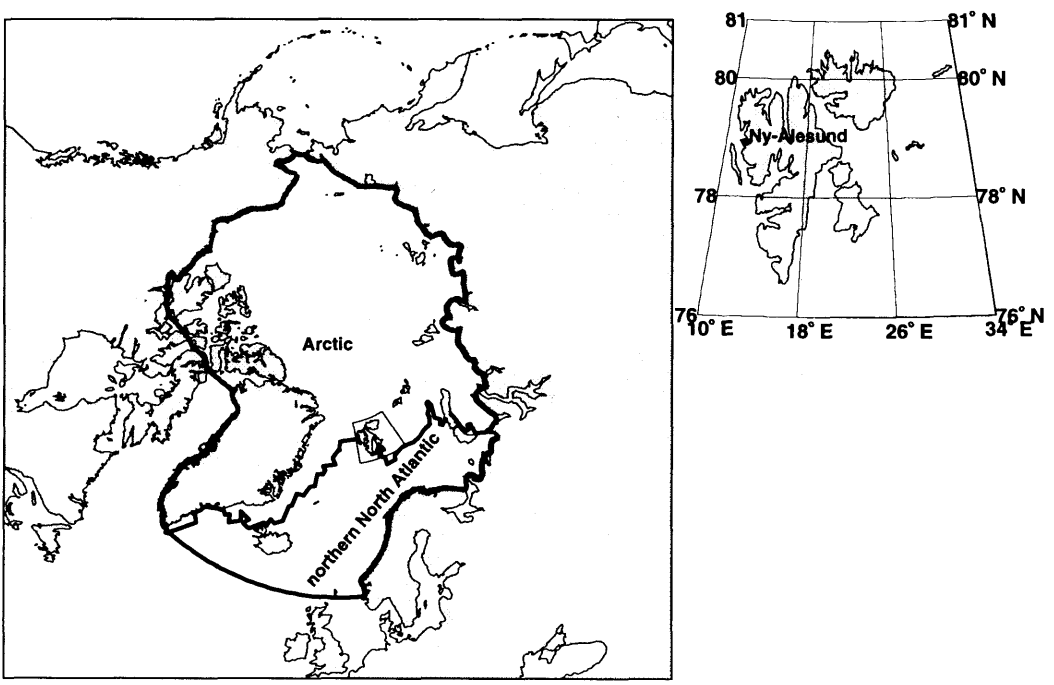


Fig. 1. The geographical position of the measuring site. The location of Ny-Ålesund (78°54′ N, 11°53′ E) is shown in the insert. In this work, the term "northern North Atlantic" is defined as the ice-free Atlantic north of 60° N, as indicated in the overview chart. The area defined as "Arctic" is also shown in the main map.

To construct a refined global budget of carbon in the atmosphere, it is necessary to aim at detailed regional budgets, and to study key aspects of the carbon cycle. Due to its proximity to the North Atlantic and northern Siberia, we believe Spitsbergen is an excellent site to study the cycling of carbon through these high-latitude reservoirs. Below, we present measurements of atmospheric CO_2 at Ny-Ålesund, Spitsbergen, and show that they can provide quantitative estimates of the fluxes into the ocean during specific periods of the year. In Section 4, we discuss future use of the data series and suggest an independent method to determine the fluxes of CO_2 into the North Atlantic.

2. Data

Continuous monitoring of atmospheric CO_2 was commenced on Spitsbergen in October 1988 and has continued ever since. During an initial period, the measurements were performed in the valley of Kongsfjorden, 1 km from the village of Ny-Ålesund, but in March 1990, the instrument was moved to the new baseline station on the mountain Zeppelinfjellet 1.5 km further away from the village. The former location was at about 50 m a.s.l. and susceptible to local contamination from Ny-Ålesund (20–200 inhabitants). The mountain station is placed on a steep ridge ca. 500 m a.s.l. and is surrounded by rocky slopes and glaciers.

The CO_2 analysis in the valley of Kongsfjorden and on Zeppelinfjellet has been performed by the same non-dispersive infrared analyzer (UNOR 4N, H. Maihak AG, Hamburg, Germany). Sample air is dried (Perma Pure model PD, Perma Pure Products, Inc., Farmingdale, NJ, USA) prior to the CO_2-analyzer. In order to check instrumental drift, the system is calibrated every 3 h towards 3 working standards which are carefully calibrated in Stockholm both prior and after usage in Ny-Ålesund. The primary standards used in Stockholm to calibrate the working standards were recently calibrated at the National Oceanic and Atmospheric Administration, Climate Monitoring and Diagnostics Laboratory (NOAA/CMDL, Boulder, Co, USA). All data thus refer to the provisional X93 mole fraction scale. Due to changes in the instrumental set-up and handling of reference gases, the accuracy of the data has varied throughout the years. It has, however, always been better than ca. 0.5 ppmv. For further details about the instrumentation and calibration procedures, we refer to Holmén et al. (1995).

The data from 1988 through 1993 is shown in Fig. 2. In order to estimate an annual amplitude and a secular trend, we performed a least square fitting of the consecutive 10-day mean mixing ratios of CO_2 at Kongsfjorden and Zeppelinfjellet to the linearly increasing harmonic function,

$$q(t) = q_0 + q_1 t + \sum (a_k \sin 2\pi t + b_k \cos 2\pi t), \quad (1)$$

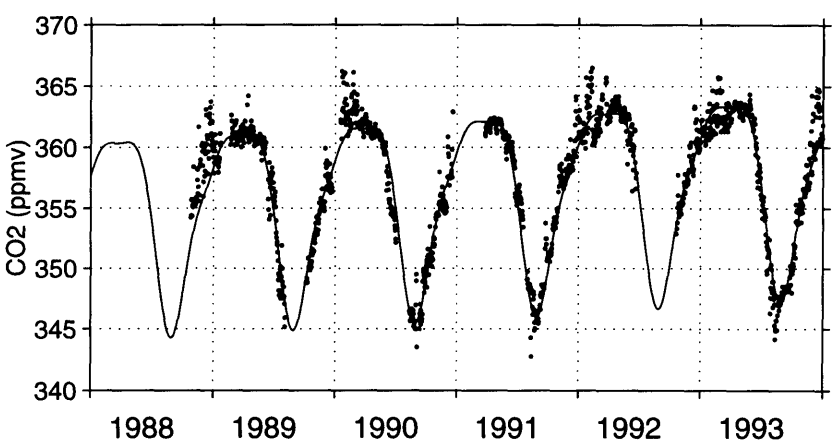
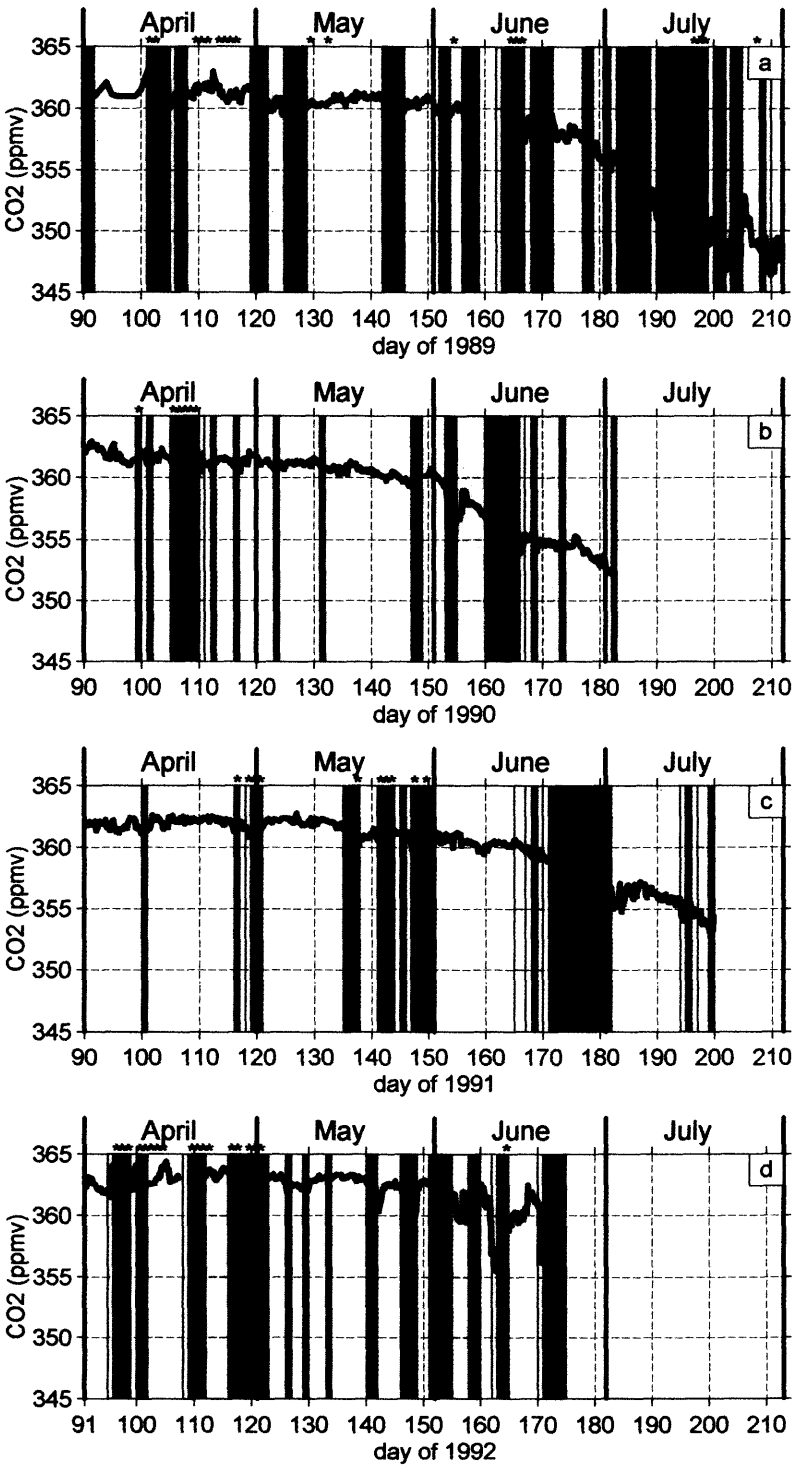


Fig. 2. Mixing ratio of CO_2 at Kongsfjorden and Zeppelinfjellet, Ny-Ålesund, October 1988 through December 1993. Dots are daily-mean mixing ratios. The solid line is the harmonic function (eq. (1)) fitted to the combined data for the years 1988–1993.


April
May
June
July
a
b
c
d
CO2 (ppmv)
365
360
355
350
345
90 100 110 120 130 140 150 160 170 180 190 200 210
91 100 110 120 130 140 150 160 170 180 190 200 210
day of 1989
day of 1990
day of 1991
day of 1992

where $k = 1$ to 4, t is time and q is the mixing ratio of CO_2. q_0, q_1, a_k, and b_k are the coefficients to be fitted. The annual amplitude thus achieved was 16.3 ppmv, the mean increase for the years 1988 through 1993 is 0.54 ppmv/year. Despite some gaps in the record, the high temporal resolution maintained during operation, together with the favorable localization of the monitoring site, allows some observations not available elsewhere. The details of particular interest in this work are the synoptic scale variations (time-scales of hours to a few days) appearing throughout the record. In winter (November to March), we occasionally measure a CO_2-increase of several ppmv in only a few hours, often connected with a dramatic increase in the mass of fine particles (Lejenäs and Holmén, 1996). The rapidity of the excursions is connected to the intensive cyclone activity which affects Spitsbergen at this time of the year. After studies of three-dimensional trajectories, we can attribute this phenomenon to long-range transport from regions affected by anthropogenic activities. Worth noting is that air, enriched in CO_2, arriving in Ny-Ålesund during winter, most often seems to stem from the European part of Siberia, as also found to be the case at Alert, Ellesmere Island (Higuchi and Daggupaty, 1985), and at Barrow, Alaska (Halter et al., 1985). In early spring (April to mid-May), the levels of CO_2 are approximately constant, while negative anomalies ("dips"), lasting from hours to several days, are seen from late May and onwards, cf. Fig. 3. The timing and magnitude of these dips show striking similarities from year to year. The transition to the summer minima in atmospheric CO_2 begins in late May and is often step-wise, i.e., a rapid decrease of a few ppmv not followed by a return back to higher levels as in the case of the previous dips. This is a reflection of how air, depleted in CO_2 due to surface uptake further south, penetrates into the Arctic. The exchange of air takes place as a complete change of air-mass rather than a slow mixing of the Arctic air with the air from further south.

The remaining part of the paper will be devoted to understanding the origin of these synoptic spring/summer variations in atmospheric CO_2. For this purpose, we have employed the Irish Meteorological Service trajectory model (Mc Grath, 1989) operating at the European Centre for Medium Range Weather Forecast (ECMWF), and calculated trajectories arriving in Ny-Ålesund for the years 1989 through 1992. The trajectories are first used to determine the origin of the different air-masses affecting Spitsbergen; later, they are combined with a simple air-sea flux scheme to investigate the plausibility of oceanic uptake as the reason for the anomalies.

3. Results

3.1. Classifying the trajectories

Isentropic trajectories, arriving at 950 hPa and extending 5 days backwards, were calculated and divided into 3 classes. If the trajectory passed over the northern North Atlantic (Fig. 1), near the sea surface, it was classified as "N-type", if it passed over the northern North Atlantic, but above a typical mixed layer it was denoted "n-type". As we did not have any information about the mixed layer, we adopted a constant value of $p = 950$ hPa (≈ 500 m) for the top of this layer, which is probably a high estimate, but reasonable. All air-masses that were considered of Arctic origin were denoted "A-type", this class also included trajectories that passed over sea-ice in the Greenland or Barents Sea. Finally, we also noted if the air originated from the continental regions south of Arctic in Fig. 1.

Fig. 3(a). April through July 1989

During the spring of 1989, when the instrument was operating in the valley of Kongsfjorden, dips

Fig. 3. CO_2 mixing ratio at Ny-Ålesund in April–July. The data is plotted against days since start of the year. (1992 is a leap-year; 00 GMT 1 April 1992 is thus 91.0 instead of 90.0, as for 1989–1991). A classification of the trajectories arriving to Ny-Ålesund is indicated in the figure. Blue shading is "N-type" days (the trajectory passed over the northern North Atlantic, below 950 hPa, during at least 3 of the last 5 days), yellow is "n-type" (the trajectory passed over the northern North Atlantic, above 950 hPa, during at least 3 of the last 5 days), white is "A-type" (the trajectory passed over the Arctic during at least 3 of the last 5 days). Red is when none of the above criterions could be fulfilled. An asterisk (*) above the figure indicates that the air-parcel spent some time over the continents south of the Arctic during the last 5 days. (a) Kongsfjorden, 1 April–31 July 1989. (b) Zeppelinfjellet, 1 April–2 July 1990. (c) Zeppelinfjellet, 1 April–19 July 1991. (d) Zeppelinfjellet, 1 April–22 June 1992.

Tellus B 48 1-4

are seen on day 141–145, 151–154, 164–166, 169–171, 192–198, 201–204, and on day 209–210. Most of the dips coincide with the advection of air from the northern North Atlantic. The trajectory arriving on day 144 which is not classified as "N-type" did in fact pass over the Greenland and Iceland Sea, but as the winds were high during passage, it traversed the region rapidly and could not fulfil the criteria of spending at least 3 days over the area. It should be kept in mind that the flux of CO_2 across the air sea interface, defined in eq. (3) below, is a strong function of wind-speed. The large dip in mid-June (day 164–166) occurred in air that had been in contact with the British Isles and the large dip in mid-July (day 196–199) occurred in air that was in contact with Siberia, north of 65°N, 5 days before arrival in Ny-Ålesund. Worth noting in Fig. 3a is the "N-type" days in April and early May (i.e., prior to the phytoplankton bloom observed around Iceland) which are not associated with dips in atmospheric CO_2, while several of the "N-type" days in July do not have significant CO_2-anomalies.

Fig. 3(b). 1 April to 2 July 1990

In 1990, a dip is seen on day 153–155 which coincide well in time with those during 1989. During the first half of June, the mixing ratios of CO_2 decreased by 5 ppmv. When comparing the data with the trajectories, we conclude that the dip in early June occurred when the air at Ny-Ålesund had been in contact with the surface waters of the northern North Atlantic prior to arrival. The "N-type" days in mid-June (day 160–165) are connected with decreasing CO_2-levels.

Fig. 3(c). 1 April to 19 July 1991

A clear dip is evident in late May (day 146–150), as in previous years. Also in this case (and on days 118–120 and 137–138 when weaker dips occur), the anomaly is associated with air stemming from the atmospheric mixed layer of the northern North Atlantic. In late April through May, there are several days which encounter air that has been in contact with continental regions 5 days prior to arrival; most of these days are associated with negative anomalies in CO_2. Between 15 and 27 June, the mixing ratio decreased by 5 ppmv, most of this taking place between day 174 and 175 when the CO_2 mixing ratios decreased by ca. 3 ppmv in 24 h. This precipitous drop occurred during a period of advection of air that had been in contact with the surface of the northern North Atlantic.

Fig. 3(d). 1 April to 22 June 1992

During 1992, we had problems with the instrument and the sampling errors during this period are larger than for previous years. Nevertheless, clear dips are visible around day 142, 146–148, 153–154, and day 162–163. Apart from the dip at days 153–154, all of these are associated with "N-type" trajectories (note however, that the "N-type" days sometimes leads or lags the measured dips by 1 or 2 days during this year). The large drop of atmospheric CO_2 (4 ppmv in 12 h) occurring on the "*n*-day" 19 June (day 170), is similar, though greater in magnitude than in June 1991. We can not rule out that this drop (and the dips occurring in late May and early June) is a result of the lower precision in the measurements during this time, but the similarity with previous years and the correspondence with the calculated trajectories are clear.

From this analysis, we conclude that 3-dimensional trajectories calculated only 5 days backwards do give valuable information about the characteristics of the air arriving in Ny-Ålesund. In April through mid-May, the changes in CO_2 are small, even during periods of advection of air that has been in contact with the waters of the North Atlantic, or the continents further south. From mid-May through June, changes in the origin of the 5-day trajectory are often also seen in the data. Air which has been in contact, or sometimes only passed over, the North Atlantic frequently shows lower CO_2-values. However, the negative anomalies are also occasionally associated with air which has been in contact with Europe or Siberia. Due to different technical problems at Ny-Ålesund, data are missing in July for several of the years, which hinders us from drawing general conclusions for this month. Throughout all years, the changes in CO_2 concentrations are small in "A-type" air.

3.2. *The Lagrangian model*

In order to get a more quantitative picture of the situation during spring and summer, we combined the trajectories with an air-sea flux scheme, see Fig. 4. If the parcel of air travels above the mixed layer, or over land, or ice, it keeps its previous CO_2 concentration. When the air-parcel finds itself in

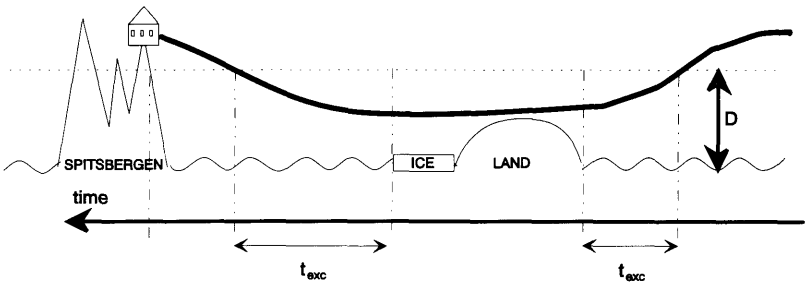


Fig. 4. Schematic picture of the Lagrangian model used in the calculations.

the maritime mixed layer at one of the discrete time-steps of the trajectory model, the change in CO_2 that occurs at the bottom, due to air-sea exchange, affects the whole mixed layer and consequently also the air which eventually ends up at Ny-Ålesund. The calculated change in CO_2 concentration, Δc, due to oceanic uptake is thus:

$$\Delta c = \sum \frac{F}{D} t_{\text{exc}} \qquad \text{moles/m}^3, \tag{2}$$

where D is the depth of the maritime mixed layer and t_{exc} the time the air-parcel finds itself in the mixed layer, over open water. The change in concentration is converted to mixing ratio using the density of air at 288 K and sea-surface pressure. The flux of CO_2, F, across the air sea boundary is proportional to the square of the instantaneous 10 m wind-speed (Wanninkhof, 1992) and also slightly dependent on the temperature and salinity of the sea-water. We have chosen the following relation (which varies within $\pm 5\%$ in the temperature- and salinity-range of $-1°\text{C} \leqslant T \leqslant 25°\text{C}$ and $30 \leqslant \sigma \leqslant 37$):

$$F = 3.0 \cdot 10^{-11} (u_{10})^2 \, \Delta\text{pCO}_2 \qquad \text{moles/(m}^2\text{s)}. \tag{3}$$

u_{10} is the instantaneous wind 10 m above the sea-surface (in m/s), ΔpCO_2 is the difference in partial pressure between the sea-surface and atmosphere (in μatm). The trajectory model only gives information about the wind at air-parcel height. A first attempt to reduce these winds to 10 m by using a standard Ekman profile yielded unreasonably low wind-speeds, probably because the temperature gradient was not included. Instead we reduced the winds with a constant factor. The factor was chosen in order that the mean wind in the atmospheric mixed layer, when all winds above 10 m in the mixed layer were multiplied by the factor, become approximately the same as the mean wind at Bear Island (74.5°N, 19°E) during the months of April–July ($\sim$6.5 m/s, Vowinckel and Orvig, 1970). The factor 0.9 gave a reduced mean wind of 6.1 m/s for April–July 1989–1992. This is of course a gross oversimplification, but the critical parameters in the model are the depth of the mixed layer, and ΔpCO_2, none of which are available to us and furthermore highly variable in time and space. As in Subsection 3.1. above, we have chosen $D = 500$ m (constant in time and space) as a reasonable value to use in the following calculations. For ΔpCO_2, we are somewhat better off. Systematic monitoring of the carbonate chemistry in the Northern Atlantic has been performed around Iceland since the beginning of the 1980s (Takahashi et al., 1985; Takahashi et al., 1993).

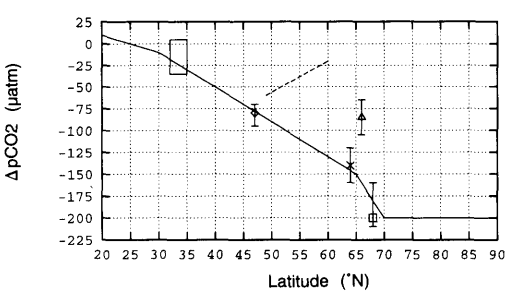


Fig. 5. The solid line is the latitudinal gradient of ΔpCO_2 used as lower boundary in the Lagrangian model. The line is based on data from: square and cross, north and south of Iceland (Takahashi et al., 1993); triangle is Ocean station M (Gislefoss, 1994); diamond is JGOFS/NABE site (Takahashi et al., 1993); hatched line is from Watson et al., (1991); the large box is the approximate ΔpCO_2 and positions of TTO/NAS leg 3, 16 May–14 June 1981 (Takahashi and Brewer, 1986).

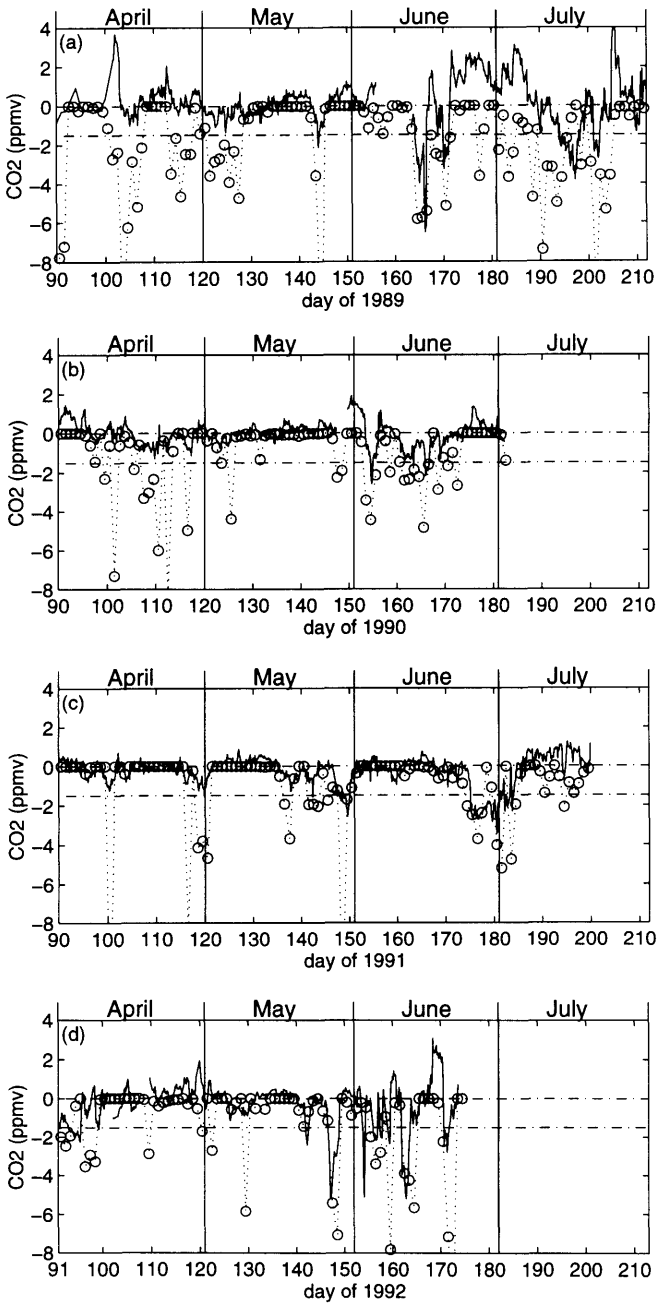


Fig. 6. CO_2 anomalies at Ny-Ålesund in April–July. The data is plotted against days since start of the year. The solid line is the measured deviation from the harmonic function, eq. (1), fitted to the CO_2 data each year. The circles connected by a dotted line are the decrease in CO_2 calculated every 24 h. A dash-dotted line indicating negative anomalies greater then 1.5 ppmv is also shown in each graph. (a) Kongsfjorden, 1 April–31 July 1989. (b) Zeppelinfjellet, 1 April–2 July 1990. (c) Zeppelinfjellet, 1 April–19 July 1991. (d) Zeppelinfjellet, 1 April–22 June 1992.

The measurements show a substantial decrease of dissolved inorganic carbon, carbon dioxide, and nutrients in late May and early June, caused by phytoplankton blooms. The onset of withdrawal reveals striking similarities from year to year. While the oceanic pCO_2 is in approximate equilibrium with respect to the atmosphere in mid-May, the situation in late May only a few weeks later, is dramatically changed. The air-sea difference is now $\sim 200\,\mu$atm in the waters north of Iceland and $\sim 130\,\mu$atm at the south. The same strong seasonality, albeit with a smaller amplitude, is observed at other North Atlantic sites (e.g., at JGOFS/NABE site $\sim 47°$N, 20°W (Takahashi et al., 1993), and at Station M (66°N, 2°E), (Gislefoss, 1994). In order to achieve consistent boundary data for our calculations, we gathered several measurements performed in the region at the end of May, and constructed the latitudinal gradient of ΔpCO_2 depicted in Fig. 5 which is assumed to be constant during the course of the 4 months. When sea-ice is present (see Fig. 1), ΔpCO_2 is set to zero in the model.

Additional minor sources of error in our simple Lagrangian model are possible imperfections in the trajectory model, and the fact that the ice-edge is a climatological value for May–June.

In Figs. 6, we have plotted the deviation of the measured CO_2 from a deseasonalised and detrended curve, together with the results of the modeling exercise. As can be seen, the correlation in time between the calculated dips and the measured ones is often excellent, while the success in capturing the magnitude of the dips is rather poor throughout the record, as expected due to the simplified assumptions made.

Fig. 6(a). April through July 1989

During 1989, the measured data show negative anomalies greater than 1.5 ppmv at day 143–144, 164–166, 168, 170–171, 193–198, and on day 201–202. The large dip around day 197 is not seen by the model, while all the others are caught well in time. The magnitude of the dips on day 144 and 201 is greatly overestimated. Apart from the dip on day 201, all deviations postulated by the model in July, and April through early May are absent, or small, in the data.

Fig. 6(b). 1 April to 2 July 1990

There are significant dips in the data at day 154, 163, and on day 165–166. The timing is predicted well by the model while the magnitude is overestimated by a factor of two. As in the previous year, the model has calculated fairly large dips in April to early May which are not seen in the real data.

Fig. 6(c). 1 April to 19 July 1991

During this year, the data indicates dips on day 142, 148–150, and on day 175–184. The model has caught all these, albeit with up to a factor of two overestimate in magnitude. The large dips predicted by the model in April through mid-May are connected with small dips in the data.

Fig. 6(d). 1 April to 22 June 1992

In 1992, there were dips in the data on day 91, 93–95, 99, 141–142, 146–148, 153–154, 155–157, 159, 161–163, and on day 170–171. The anomalies in early April and late May are well correlated, both in time and magnitude with the modelled dips. In June, the data is very scattered, possibly due to instrumental errors. Nevertheless, the deviations seen on day 161–163 and 170–171, do correlate with the calculated ones, the latter is, however, greatly overestimated by the model. On a few occasions, in late April and early May, the model calculates dips which are not seen in the data.

During all years, the model frequently overestimates the amplitude of the dips in April and early May (or postulates dips which are not seen in the data). In late May and June, however, the calculated dips most often correlate with significant measured dips. This is probably caused by the assumed ΔpCO_2 used in our calculations. We know through measurements that pCO_2 in the mixed layer of the North Atlantic is in approximate balance with the overlying atmosphere in April and early May. In our calculations, we have used a, constant in time, latitudinal gradient of ΔpCO_2, which is obviously unrealistic during the first part of our time-series. A second reason for the poor agreement might be due to the fact that the ice-coverage in April and early May is larger than the constant value used throughout all our calculations.

While the prescribed ΔpCO_2 is more realistic in late May through June, the cause for the occasional mismatch in magnitude between the calculated and measured dips, during this period is most likely a reflection of the simple treatment of the mixed layer in our model. The mixed layer

depth, and thus the concentration changes, can vary by at least a factor of 2 during the course of a single 5-day trajectory. Our simple model can therefore not be used to calculate the expected concentration change during a single event; rather, it gives us semi-quantitative information about the timing and magnitude of the anomalies, given the assumed boundary conditions.

The positive excursions seen in the data, are obviously not captured by the model, since it includes only sinks. As can be seen in Fig. 3, however, we have not found any correlation between these anomalies and the horizontal origin of the 5-day trajectories. It is possible that the anomalies are the result of vertical movements of the air-parcels prior to arrival in Ny-Ålesund.

The model sinks only come into work over open water, so it is interesting to note that almost all the dips seen in the measured record during May and June are caught by the model. This indicates that oceanic uptake is the major cause for the anomalies in the Ny-Ålesund CO_2-record during this period.

During 1989 when we have data for July, the measured record displays a dip not captured by the model. This episode is most likely the result of air transported to Ny-Ålesund from the continents (cf. Fig. 3d), where the photosynthesis of the terrestrial plants has reduced the atmospheric loading of CO_2. At some occasions during July 1989 and 1991, the model indicates oceanic uptake which is not seen in the record, again this can be the result of an imperfect prescription of ΔpCO_2. The pCO_2 in the surface of the North Atlantic is increasing in summer due to the seasonal warming of the surface water.

4. Conclusions and future plans

Both near sea-surface, and on a mountain top, on Spitsbergen we have measured dips in atmospheric CO_2 of 2–4 ppmv during spring and early summer. These dips have occurred during all years and the timing is identical to a measured draw down in North Atlantic pCO_2. We have shown that during the 4 spring-periods so far studied, air arriving in Ny-Ålesund after crossing the North Atlantic is depleted in CO_2. We have also demonstrated by simple calculations that these dips are in accordance with the CO_2 saturation in the sea at those times. We therefore conclude that most of the rapid CO_2 decreases seen in the early summer at Ny-Ålesund are caused by oceanic uptake. One problem in this analysis is to distinguish the CO_2 uptake in the North Atlantic from that which is taking place in the terrestrial biosphere further south. A future goal is to utilize a high-resolution 3-dimensional atmospheric transport model and thereby calculate the fluxes of CO_2 into the different regions. The input to such a model should be archived weather, together with high resolution sea surface pCO_2, and ice coverage. Such work is ongoing under the auspices of a collaborative Nordic program. By monitoring a species with only marine sources (or solely continental sources) at Ny-Ålesund, we would not only have a method to verify the transport model, but also a handy way of classifying the air-masses affecting Spitsbergen. Candidate species include the oceanic species dimethyl sulphide (DMS), the isotopes ^{13}C and ^{14}C in CO_2, and the truly terrestrial Radon-222. Another way of verifying the model would be to obtain CO_2 data with high temporal resolution at other locations around the North Atlantic.

5. Acknowledgments

Thanks are due to the Norsk Polar Institutt technical staff at Ny-Ålesund for help with the instruments. Ray Mc Grath also deserves great thanks for providing access to his trajectory model and help with solving practical matters when calculating the trajectories. This work has been supported by the Swedish Environmental Protection Agency and the Nordic Council of Ministers.

REFERENCES

Broecker, W. S. and Peng, T.-H. 1992. Interhemispheric transport of carbon dioxide by ocean circulation. *Nature* **356**, 587–589.

Conway, T. J., Tans, P. P., Waterman, L. S., Thoning, K. W., Kitzis, D. R., Masarie, K. A. and Zhang, N. 1994. Evidence for interannual variability of the

carbon cycle from the National and Oceanic Atmospheric Administration/Climate Monitoring and Diagnostics Laboratory Global Air Sampling Network. *J. Geophys. Res.* **99D**, 22831–22855.

Dixon, R. K., Brown, S., Houghton, R. A., Solomon, A. M., Trexler, M. C. and Wisniewski, J. 1994. Carbon pools and flux of global forest ecosystems. *Science* **263**, 185–190.

Enting, I. G. and Mansbridge, J. V. 1991. Latitudinal distribution of sources and sinks of CO_2. Results of an inversion study. *Tellus* **43B**, 156–170.

Gislefoss, J. S. 1994. *Carbon profiles in the Nordic Seas.* Doctoral thesis, Radiological Dating Laboratory, Norwegian Institute of Technology, Trondheim, Norway.

Halter, B. C., Harris, J. M. and Rahn, K. A. 1985. A Study of winter variability in carbon dioxide and arctic haze aerosols at Barrow, Alaska. *Atmos. Environ.* **19**, 2033–2037.

Higuchi, K. and Daggupaty, S. M. 1985. On variability of atmospheric CO_2 at station Alert. *Atmos. Environ.* **19**, 2039–2044.

Holmén, K., Engardt, M. and Odh, S.-Å. 1995. *The carbon dioxide measurement program at the Department of Meteorology at Stockholm University.* International Meteorological Institute in Stockholm, Report CM-84, 38 pp.

Houghton, R. A., Boone, R. D., Fruci, J. R., Hobbie, J. E., Melillo, J. M., Palm, C. A., Peterson, B. J., Shaver, G. R., Woodwell, G. M., Moore, B., Skole, D. L. and Myers, N. 1987. The flux of carbon from terrestrial ecosystems to the atmosphere in 1980 due to changes in land use: geographic distribution of the global flux. *Tellus* **39B**, 122–139.

Kauppi, P. E., Mielikäinen, K. and Kuusela, K. 1992. Biomass and carbon budget of European Forests, 1971 to 1990. *Science* **256**, 70–74.

Keeling, C. D., Piper, S. C. and Heimann, M. 1989. A three-dimensional model of atmospheric CO_2 transport based on observed winds. (4). Mean annual gradients and interannual variations. In: *Aspects of climate variability in the Pacific and the Western Americas* (ed. D. H. Peterson). American Geophysical Union, Washington, DC, pp. 305–363.

Kempe, S. and Pegler, K. 1991. Sinks and sources of CO_2 in coastal seas: the North Sea. *Tellus* **43B**, 224–235.

Kolchugina, T. P. and Vinson, T. S. 1993. Carbon sources and sinks in forest biomes of the former Soviet Union. *Global Biogeochem. Cycles* **7**, 291–304.

Lejenäs, H. and Holmén, K. 1996. Characteristics of the large-scale circulation during episodes with high and low concentrations of carbon dioxide and air pollutants at an Arctic monitoring site in winter. *Atmos. Environ.*, in press.

Mc Grath, R. 1989. *Trajectory models and their use in the Irish Meteorological Service.* Irish Meteorological Service, Glasnevin Hill, Dublin, Internal Memorandum No. 112/89, 12 pp.

Plantinga, A. J. and Birdsey, R. A. 1993. Carbon fluxes resulting from U.S. private timberland management. *Clim. Change* **23**, 37–53.

Sarmiento, J. L. and Sundquist, E. T. 1992. Revised budget for the oceanic uptake of anthropogenic carbon dioxide. *Nature* **356**, 589–593.

Sedjo, R. A. 1992. Temperate forest ecosystems in the global carbon cycle. *Ambio* **21**, 274–277.

Sundquist, E. T. 1993. The global carbon dioxide budget. *Science* **259**, 934–941.

Takahashi, T. and Brewer, P. G. 1986. Hydrographic and chemistry data for the TTO/NAS expedition, April–October 1981: Revised carbon chemistry data: In: *Transient tracers in the oceans (TT). Hydrographic data and carbon dioxide systems with revised carbon chemistry data.* Carbon Dioxide Information Center, Tennessee.

Takahashi, T., Ólafsson, J., Broecker, W. S., Goddard, J., Chipman, D. W. and White, J. 1985. Seasonal variability of the carbon-nutrient chemistry in the ocean areas west and north of Iceland. *J. Mar. Inst., Reykjavik* **9**, 20–36.

Takahashi, T., Ólafsson, J., Goddard, J. G., Chipman, D. W. and Sutherland, S. C. 1993. Seasonal variation of CO_2 and nutrients in the high-latitude surface oceans. A comparative study. *Global Biogeochem. Cycles* **7**, 843–878.

Tans, P. P., Fung, I. Y. and Takahashi, T. 1990. Observational constraints on the global atmospheric CO_2 budget. *Science* **247**, 1431–1438.

Vowinckel, E. and Orvig, S. 1970. The Climate of the north polar basin. In: *Climates of the polar regions* (ed. S. Orvig). Elsevier Science B. V., Amsterdam, 129–252.

Wanninkhof, R. 1992. Relationship between wind speed and gas exchange over the ocean. *J. Geophys. Res.* **97C**, 7373–7382.

Watson, A. J., Robinson, C., Robinson, J. E., Williams, P. J. le B. and Fasham, M. J. R. 1991. Spatial variability in the sink for atmospheric carbon dioxide in the North Atlantic. *Nature* **350**, 50–53.

Watson, R. T., Rodhe, H., Oeschger, H. and Siegenthaler, U. 1990. Greenhouse gases and aerosols. In: *Climate Change, the IPCC Scientific Assessment*, (ed. J. T. Houghton et al.). Cambridge University Press, Cambridge, pp. 1–40.