ESTIMATED ANNUAL REGIME OF ENERGY-BALANCE COMPONENTS, EVAPOTRANSPIRATION AND SOIL MOISTURE FOR A DRAINAGE BASIN IN THE CASE OF A CO_2 DOUBLING*

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Abstract. Assuming a doubling of the atmospheric CO_2 concentration, parameters of an empirical formula for calculating the daily net terrestrial radiation under the climatic conditions of Belgium are determined. The developed method takes into account information yielded by climate models about the CO_2 impacts. Annual regimes of the energy-balance components are calculated for a drainage basin in Belgium. A daily step conceptual hydrological model (developed at the Royal Meteorological Institute of Belgium) was run to estimate the effective evapotranspiration and the soil moisture in the $2 \times CO_2$ case; results of this simulation are compared with the present-day conditions.

1. Introduction

It is broadly accepted that a doubling of the atmospheric CO_2 concentration induces an increase in temperature and absolute humidity within the near-surface air layer (Kellogg, 1979; Watts, 1980; Clark, 1982; Manabe, 1983; Flohn and Fantechi, 1984).

It also induces a decrease in net terrestrial and global solar radiations (Chou *et al.*, 1982; Ramanathan, 1981).

These changes have repercussions on the energy balance and thus modify the evapotranspiration and both the latent and sensible heat transfers.

The annual regime of precipitation and the nebulosity also respond to the increase in CO_2 (Manabe *et al.*, 1981; Washington and Meehl, 1983, 1984; Mitchell, 1983).

Knowledge of the rate of the change that affects both potential and effective evapotranspiration in every month of the year is badly needed for assessing the impact of specific climatic changes on either water resources or agricultural production.

It is obvious that an increase in potential evapotranspiration does not have the same significance when river stages are low as when they are high, nor when the unsaturated zone of the soil is normally saturated as when it is drying.

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Our purpose is to evaluate quantitatively the impact of the CO_2 doubling on the annual regime of the effective evapotranspiration and of the soil moisture. This objective justifies the use of a specific hydrological model on a particular drainage basin.

But, before running any hydrological model, the surface energy-balance components that govern surface hydrology must be evaluated.

The net terrestrial radiation $L^\star$ can be expressed empirically (Brunt, 1932; Monteith, 1973) by

$$L^\star = L_u - L_d = \sigma T^4 [1 - (a + b\sqrt{e})(1 + cn^2)] \qquad (1.1)$$

where L_u denotes the upward flux from the surface and L_d the downward flux from the atmosphere; where T and e are respectively the daily absolute air temperature (deg K) and the water vapour pressure (hPa); where n represents the nebulosity as the complement $1 - Ir$ of the percentage of possible sunshine and σ the Stefan – Boltzmann constant (49.01254 10^{-8} J cm^{-2} K^{-4} d^{-1}).

Coefficients a, b, c parameterize the atmospheric emissivity, a function of the amounts of water vapour, CO_2 and nebulosity.

To take into account the differences between seasonal average atmospheric conditions, the parameters a, b, c are estimated for each season. The formula (1.1) is a statistical relationship based on the correlation between atmospheric water vapour content and temperature on the one hand, and their values near the ground on the other hand. The formula does furthermore not discriminate between various cloud types. The average sensitivity of L_d to seasonal nebulosity and cloud types distribution is taken into account through parameter c.

The seasonal parameter values a, b, c have been determined for Belgium on the basis of net radiation daily values observed at Uccle (50°48′N; 04°21′E; 105 m a.m.s.l.) over the ten-year period 1972–1981 (Bultot and Dupriez, 1974; Bultot *et al.*, 1983).

Formula (1.1) allows the calculation of daily approximate values of $L^\star$ for observation stations other than Uccle which is the only site in the country equipped with a pyrradiometer (i.e. where the net radiation is recorded). In this study, the wordings scenario 0 and scenario 1 refer, respectively, to the present-day atmospheric CO_2 concentration and to the $2 \times CO_2$ concentration.

The first object of the research is to predict, as precisely as possible, the values that would be assumed by the parameters a, b and c of formula (1.1) in scenario 1.

For this purpose, account must be taken of:

(1) the properties inherent in the parameters of formula (1.1)

(2) the predictions about modifications of meteorological variables in scenario 1.

These predictions are given by various climate models and are detailed in section 4 of this paper.

In a second part, the formula (1.1) with the modified parameters is used to estimate daily values of the net terrestrial radiation in scenario 1.

In the third part, our IRMB conceptual hydrological model (Bultot and Dupriez, 1976a and b; revised version: Bultot and Dupriez, 1985) is applied, up to the point of effective evapotranspiration estimation, to a particular drainage basin.

The maximum available water content of the soil system and the relative soil moisture are determined. It is then possible to estimate the energy daily consumed by evapotranspiration and to deduce from the energy balance equation the sensible heat transfer and the Bowen ratio.

Our results are next compared with estimations yielded by global circulation model (GCM) simulations.

2. Estimation of the Net Terrestrial Radiation

The equation of the energy balance of a natural surface may be written

$$Q^\star - Q_g = H + LE \tag{2.1}$$

where $Q^\star$ denotes the net radiation, Q_g the heat exchange between soil surface and ground, H and LE the transfers of sensible heat and latent heat of evaporation between the soil surface and adjacent air layers.

The net radiation is expressed by

$$Q^\star = (1 - r)\, K\!\downarrow - L^\star \tag{2.2}$$

where r is the albedo of the surface and $K\!\downarrow$ the global solar radiation.

The exchange between soil surface and ground is estimated by

$$Q_g = \sum_{i=1}^{m} \rho_i c_{pi} p_i \Delta T_i \tag{2.3}$$

where ρ_i and c_{pi} denote the density and specific heat of the soil layer i, while ΔT_i represents the day-to-day temperature difference at level i.

If the Bowen ratio $B = H/LE$ is brought in, then

$$H = (Q^\star - Q_g)\,\frac{B}{B+1}, \quad LE = (Q^\star - Q_g)\,\frac{1}{B+1}. \tag{2.4}$$

Based on direct observations recorded at Uccle, the parameters a, b and c of formula (1.1) assume the following seasonal values:

	a	b	c
Winter	0.4117	0.1604	0.1498
Spring	0.4599	0.1006	0.2397
Summer	0.6869	0.0293	0.1741
Autumn	0.5824	0.0718	0.1472.

(2.5)

The value of parameter b is furthermore related with the seasonal mean value of the water vapour pressure $\bar{e}$:

$$b = \alpha[1 - \tanh(\beta\bar{e} - \gamma)] \quad \text{with} \begin{cases} \alpha = 0.1784 \\ \beta = 0.1007 \\ \gamma = 0.4931. \end{cases} \tag{2.6}$$

The statistic R, tantamount to a multiple correlation coefficient used as a goodness of fit criterion, is equal to 0.94, which substantiates the acceptability of the fit (2.6).

For estimating the potential evapotranspiration from a natural surface by the energy balance method, use of the net radiation over the daylight period is preferable to that over the 24 h daylength (Bultot *et al.*, 1983).

The net radiation over the daylight period, $Q_e^\star$, is given by

$$Q_e^\star = (1 - r)\, K{\downarrow} - L_e^\star \tag{2.7}$$

where $L_e^\star$ is the net terrestrial radiation over the day-time hours (as opposed to night-time hours).

More explicitly:

$$L_e^\star = (L_h^\star)_e\, DPE,$$

where $(L_h^\star)_e$ is the hourly mean of the net terrestrial radiation during day-time ($\mathrm{J\,cm^{-2}\,h^{-1}}$) and DPE denotes the astronomically possible duration of sunshine (hours).

For the climatic conditions of Belgium, $(L_h^\star)_e$ is estimated by

$$(L_h^\star)_e = \sigma\, T_e^4\,[1 - (a' + b'\sqrt{e_e})(1 + c'n^2)] \tag{2.8}$$

where T_e and e_e are the means of the absolute temperature (deg K) and of the water vapour pressure (hPa) over the daylight interval and where σ is equal to $2.04219\ 10^{-8}\ \mathrm{J\,cm^{-2}\,K^{-4}\,h^{-1}}$.

The parameters a', b' and c' then assume the following seasonal values

	a'	b'	c'
Winter	0.2856	0.1618	0.3189
Spring	0.3597	0.1012	0.4045
Summer	0.5967	0.0293	0.3275
Autumn	0.4811	0.0726	0.2848.

(2.9)

Again, the parameter b' is related with the seasonal mean value of the water vapour pressure $\bar{e}_e$ by an expression having the same form as (2.6). The parameters α', β' and γ' assume respectively the value 0.1795, 0.1002 and 0.4978. The statistic R is once more equal to 0.94 and confirms the validity of the fit.

3. The IRMB Conceptual Hydrological Model

The IRMB hydrological model is a conceptual daily step model (Bultot and Dupriez, 1976a and b; revised version: Bultot and Dupriez, 1985) developed to

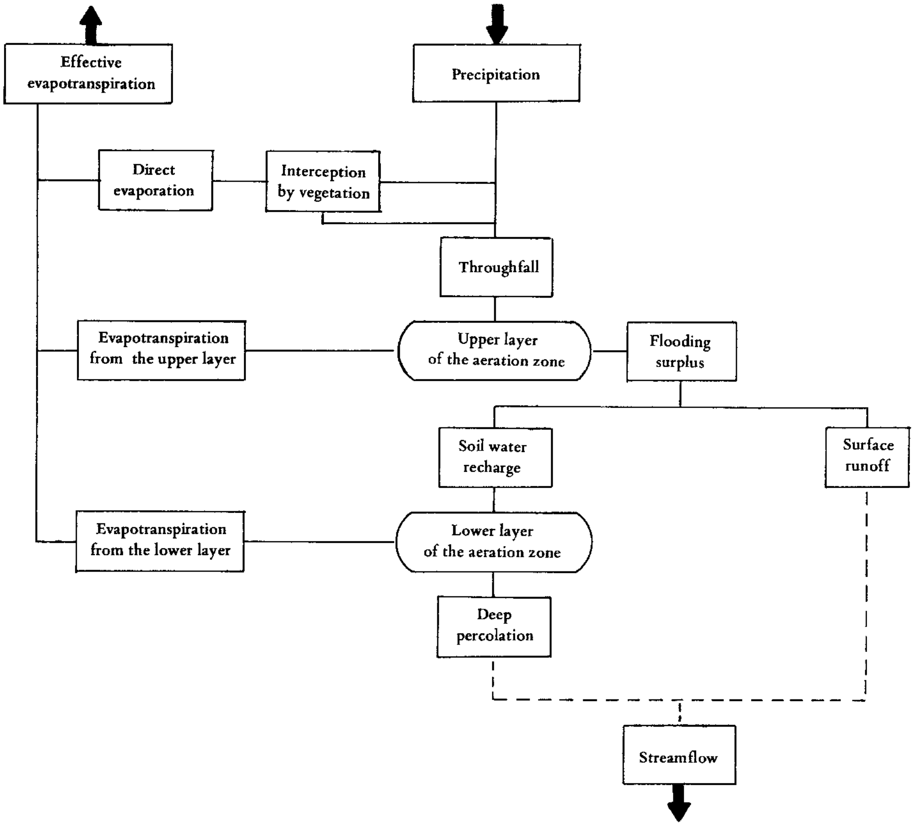

Fig. 1. Flow chart of the IRMB model (limited to its upper part).

simulate the different hydrological variables of a drainage basin, i.e. the water transfers (rainfall interception by vegetation, evapotranspiration, ...), the state variables (water contents of vegetative covers, of the unsaturated and saturated zone of the soil, ...) and the flows at the outlet of the basin (surface runoff, interflow, baseflow, ...).

Figure 1 shows the flow chart of the IRMB model limited to the step of effective evapotranspiration. In addition to daily effective evapotranspiration, the model estimates the three different ways in which water is transferred into the atmosphere in the form of vapour:

(1) *From the Vegetative Cover*

Some part of the rainfall intercepted by the vegetation is evaporated directly. The interception storage is followed from day to day. It governs the respective amounts of water that are temporarily stored on the canopy or lost by evaporation.

(2) *From the Upper Layer of the Zone of Aeration*

In the zone of aeration, a shallow near-surface layer is considered distinctly from the lower depth. This upper layer must be brought to saturation before runoff can occur during a rainfall event.

On account of its well developed root system, this layer is the first to lose soil moisture for meeting the evapotranspiration requirements; it is also the first to be replenished when precipitation occurs.

(3) *From the Lower Layer of the Zone of Aeration*

The lower layer of the zone of aeration contributes a large part of the moisture consumed by the evapotranspiration during dry weather spells. The excess of moisture, if any, percolates to the aquifer.

Five series of parameters are needed to characterize a drainage basin:

(1) surface runoff rates, linked to the surface features of the basin (soil nature and average slope); these rates are estimated through the streamflow data;
(2) areas covered by the various types of vegetation;
(3) parameters of the relationships used to estimate rainfall interception by vegetation (e.g. leaf area index);
(4) maximum available water capacities for each vegetation cover (i.e. field capacity minus storage capacity at wilting point);
(5) albedo for each type of vegetation.

A preliminary model calculates the potential evapotranspiration according to the energy balance method (Penman, 1948; Bultot *et al.*, 1983) and taking into account the phytogeographical characteristics of the drainage basin i.e. the distribution and albedo of the various types of vegetation.

The maximum available water capacity WSX of the upper layer of the zone of aeration is estimated in such a manner that simulated and observed runoffs are statistically equal (Bultot and Dupriez, 1985).

The maximum available water capacity WX of the whole zone of aeration is roughly estimated from month to month by considering both the type of soil and the depth of the root zone of the different kinds of vegetation (Thornthwaite and Mather, 1957). It is then fitted in such a manner that long period means of effective annual evapotranspiration (from the energy balance method) and flow deficit (from the water balance equation) are nearly equal.

The maximum available water capacity WIX of the lower layer of the zone of aeration is thus obtained by

$$WIX = WX - WSX.$$

4. Specific Climatic Changes Induced by a Doubling of the Atmospheric CO_2 Concentration

Various types of mathematical models of the climate system have shown that a greater amount of CO_2 in the troposphere can produce large disturbances of the climate (e.g. Watts, 1980; Clark, 1982; Manabe, 1983).

Specific changes of the climate upon which the present research is based are summarized in table I, with references listed in the last column.

Similar results were obtained by other authors (e.g. Ramanathan, 1981; Mitchell, 1983, 1986) but merely references from which values were extracted are quoted.

As regards the monthly modifications in temperature and precipitation, the expected values for $4 \times CO_2$ experiments (Manabe and Wetherald, 1975; Manabe *et al.*, 1981) have been divided by 2 (Washington and Meehl, 1983).

So, in the $2 \times CO_2$ case, temperature would be increased by 2.3 K to 3.4 K according to time of the year, and precipitation would be higher from October through April and somewhat decreased from May through August.

It is as yet ascertained that CO_2 atmospheric enrichment can have physiological effects on plant growth and photosynthesis, on stomatal resistance and on plant albedo.

According to some authors (Aston, 1984; Callaway and Currie, 1985) it seems that the effects on plant growth and stomatal resistance can counterbalance each other. This is still a matter of research and it is therefore assumed in the present study that the physiological properties of plants remain unchanged in scenario 1.

5. Sensitivity of the Parameters of the Net Terrestrial Radiation Estimation Formula

The coefficients a, b and c of formula (1.1) parameterize the present-day atmospheric emissivity. A change in atmospheric CO_2 concentration involves a modification of this emissivity (Manabe, 1983).

To estimate the values of the parameters a, b, c in the $2 \times CO_2$ case, the formula (1.1) is differentiated relatively to each one of the variables and relatively to the parameters a and b.

For lack of information, it is assumed that the distribution of cloud types is the same in scenario 0 as in scenario 1. Thus, the parameter c does not change. So:

$$\begin{aligned} \mathrm{d}L^\star = {} & 4\ \sigma T^3\, \mathrm{d}T[1-(a+b\sqrt{e})(1+cn^2)] \\ & - \ \sigma T^4[(\mathrm{d}a+\sqrt{e}\ \mathrm{d}b+b\ \mathrm{d}\sqrt{e})(1+cn^2) \\ & + \ (a+b\sqrt{e})\ 2\ cn\ \mathrm{d}n]. \end{aligned} \tag{5.1}$$

The differentiation of formulas (2.2) and (2.4) leads to

$$\mathrm{d}Q^\star = (1-r)\,\mathrm{d}K\downarrow - \mathrm{d}L^\star, \tag{5.2}$$

$$\mathrm{d}H = (\mathrm{d}Q^\star - \mathrm{d}Q_g)\,\frac{B}{B+1} + (Q^\star - Q_g)\,\frac{\mathrm{d}B}{(B+1)^2}\,. \tag{5.3}$$

The heat transfer between soil surface and ground, Q_g, given by (2.3), depends only on day-to-day soil temperature differences. Its differential

$$\mathrm{d}Q_g = \sum \rho_i c_{pi} p_i\, \mathrm{d}(\Delta T_i)$$

is related to the infinitesimal increment $\mathrm{d}(\Delta T_i)$ of the day-to-day soil temperature differences. The soil temperature is expected to be increasing from scenario 0 to scenario 1 but failing information about its variability, it is assumed that the day-to-day differences ΔT_i are the same in both scenario 0 and scenario 1. So,

$$\mathrm{d}(\Delta T_i) = 0 \text{ and } \mathrm{d}Q_g = 0 \text{ and, from (5.3):}$$

$$\mathrm{d}H = \mathrm{d}Q^\star\,\frac{B}{B+1} + (Q^\star - Q_g)\,\frac{\mathrm{d}B}{(B+1)^2}. \tag{5.4}$$

By introducing (5.2), expression (5.4) becomes:

$$\mathrm{d}H = \frac{B}{B+1}\,[(1-r)\,\mathrm{d}K\downarrow - \mathrm{d}L^\star] + (Q^\star - Q_g)\,\frac{\mathrm{d}B}{(B+1)^2}\,,$$

or still

$$H\,\frac{B+1}{B}\,\frac{\mathrm{d}H}{H} = (1-r)\,\mathrm{d}K\downarrow - \mathrm{d}L^\star + \frac{Q^\star - Q_g}{B+1}\,\frac{\mathrm{d}B}{B}\,.$$

Taking expression (5.1) into account, this relation expands as follows

$$\begin{aligned} H\,\frac{B+1}{B}\,\frac{\mathrm{d}H}{H} &= (1-r)\,\mathrm{d}K\downarrow - 4\,\sigma T^3\,[1-(a+b\sqrt{e})(1+cn^2)]\,\mathrm{d}T \\ &+ \sigma T^4\,(1+cn^2)\,\mathrm{d}a \\ &+ \sigma T^4\,(1+cn^2)\sqrt{e}\,\mathrm{d}b + \sigma T^4\,(1+cn^2)\,b\,\mathrm{d}\sqrt{e} \\ &+ 2\,\sigma T^4\,(a+b\sqrt{e})cn^2\,\frac{\mathrm{d}n}{n} + \frac{Q^\star - Q_g}{B+1}\,\frac{\mathrm{d}B}{B}\,. \end{aligned} \tag{5.5}$$

Under the assumption that the relative humidity is the same in the two scenarios (Manabe and Wetherald, 1975), the increase in air temperature involves an increase in water vapour pressure Δe which, in comparison with the usual values of e, cannot be considered as an infinitesimal element. Hence, for $\mathrm{d}\sqrt{e}$ in formula (5.5), the difference $\Delta\sqrt{e}$ must be used instead of its differential which may possibly be imprecise.

Although formula (2.6) is but a statistical relationship, it is used to estimate

Δb from Δe since the $e + \Delta e$ values are still in its range of validity. Just as for $d\sqrt{e}$, it is obvious that Δb must be used instead of db.

When solved for da, the relation (5.5) becomes

$$\begin{aligned} k_1 \, da &= k_2 \, dT + k_3 \, dK{\downarrow} + k_4 \frac{dn}{n} + k_5 \frac{dH}{H} + k_6 \frac{dB}{B} \\ &\quad + k_7 \, \Delta\sqrt{e} + k_8 \, \Delta b \end{aligned} \tag{5.6}$$

with

$$\begin{aligned} k_1 &= \sigma T^4 (1 + cn^2) \\ k_2 &= 4\,\sigma T^3 [1 - (a + b\sqrt{e})(1 + cn^2)] \\ k_3 &= -\,(1 - r) \\ k_4 &= -\,2\,\sigma T^4 (a + b\sqrt{e})\, cn^2 \\ k_5 &= H\,\frac{B+1}{B} = Q^\star - Q_g \\ k_6 &= -\,\frac{Q^\star - Q_g}{B+1} = -\,LE \\ k_7 &= -\,\sigma T^4\,(1 + cn^2)\, b \\ k_8 &= -\,\sigma T^4\,(1 + cn^2)\sqrt{e}\,. \end{aligned}$$

All the terms, and particularly k_5 and k_6, must be estimated for the true conditions of an entity 'natural surface – substratum'. For the ten-year period 1970–1979, formula (5.6) has been applied by using daily data from the Semois drainage basin at Membre (49°52′N; 04°54′E; 176 m a.m.s.l.; area 1235 sq.km).

The effective evapotranspiration E has been estimated by using the IRMB conceptual hydrological model described in section 3 of this paper. For the differentials dT, $dK{\downarrow}$ and dn/n, dH/H and dB/B, the values are extracted from table I. The increment $dK{\downarrow}$ has been distributed from day to day in proportion to the possible duration of sunshine.

Hence, daily values of the increases of the parameters a and b are obtained, together with their seasonal means.

Under the hypothesis of scenario 1, it follows that the parameters a and b of formula (1.1) assume the following values:

	a	b
Winter	0.4480	0.1327
Spring	0.5224	0.0752
Summer	0.7606	0.0122
Autumn	0.6432	0.0503

(5.7)

For a more specific determination of the impact of increased atmospheric CO_2 on potential evapotranspiration from a natural surface, the changes of net radiation and of net terrestrial radiation considered over the daylight interval rather than the entire daylength should be assessed.

For the purpose, the values of the parameters a' and b' of formula (2.8) are determined for the $2 \times CO_2$ case.

A similar reasoning as above leads to

$$\begin{aligned} \mathrm{d}(L_h^\star)_e = {} & 4\,\sigma T_e^3\,\mathrm{d}T_e\,[1-(a'+b'\sqrt{e_e})(1+c'n^2)] \\ & -\ \sigma T_e^4\,[(\mathrm{d}a'+\sqrt{e_e}\,\mathrm{d}b'+b'\mathrm{d}\sqrt{e_e})(1+c'n^2) \\ & +\ (a'+b'\sqrt{e_e})\,2\,c'n\,\mathrm{d}n]. \end{aligned}$$

Since it can be granted that the net terrestrial radiations during the daylight hours and during the complete day (24 hr) are changed in proportional manner, i.e. that

$$\frac{\mathrm{d}L_e^\star}{L_e^\star} = \frac{\mathrm{d}L^\star}{L^\star} \tag{5.8}$$

the expressions (5.2) and (5.4) become

$$\begin{aligned} \mathrm{d}Q^\star &= (1-r)\,\mathrm{d}K{\downarrow} - L^\star\frac{\mathrm{d}L_e^\star}{L_e^\star} \\ \mathrm{d}H &= \mathrm{d}Q^\star\ \frac{B}{B+1} + (Q^\star - Q_g)\frac{\mathrm{d}B}{(B+1)^2} \\ &= \frac{B}{B+1}\left[(1-r)\,\mathrm{d}K{\downarrow} - \frac{L^\star}{L_e^\star}\mathrm{d}L_e^\star\right] \\ &\quad + (Q^\star - Q_g)\frac{\mathrm{d}B}{(B+1)^2} \end{aligned}$$

so that

$$\begin{aligned} \frac{L_e^\star}{L^\star}\,H\,\frac{B+1}{B}\,\frac{\mathrm{d}H}{H} = {} & \frac{L_e^\star}{L^\star}\,(1-r)\,\mathrm{d}K{\downarrow} \\ & -4\,\sigma T_e^3\,[1-(a'+b'\sqrt{e_e})(1+c'n^2)]\,\mathrm{d}T_e \\ & +\ \sigma T_e^4\,(1+c'n^2)\,\mathrm{d}a' + \sigma T_e^4\,(1+c'n^2)\sqrt{e_e}\,\mathrm{d}b' \\ & +\ \sigma T_e^4\,(1+c'n^2)\,b'\Delta\sqrt{e_e} + 2\,\sigma T_e^4\,(a'+b'\sqrt{e_e})\,c'n^2\,\frac{\mathrm{d}n}{n} \\ & +\ \frac{L_e^\star}{L^\star}\,\frac{Q^\star - Q_g}{B+1}\,\frac{\mathrm{d}B}{B}\,. \end{aligned}$$

When solved for $\mathrm{d}a'$ with the assumption that $\mathrm{d}T_e = \mathrm{d}T$, and after having calculated b' through a formula of the same form as (2.6), the above relation becomes

$$\begin{aligned} k_1'\,\mathrm{d}a' = {} & k_2'\,\mathrm{d}T + k_3'\,\mathrm{d}K{\downarrow} + k_4'\frac{\mathrm{d}n}{n} + k_5'\frac{\mathrm{d}H}{H} + k_6'\frac{\mathrm{d}B}{B} \\ & +\ k_7'\,\Delta\sqrt{e_e} + k_8'\,\Delta b' \end{aligned} \tag{5.9}$$

with

$$\begin{aligned} k_1' &= \sigma T_e^4\,(1+c'n^2) \\ k_2' &= 4\,\sigma T_e^3\,[1-(a'+b'\sqrt{e_e})(1+c'n^2)] \end{aligned}$$

$$k_3' = -\frac{L_e^\star}{L^\star}(1-r)$$

$$k_4' = -2\,\sigma T_e^4 (a' + b'\sqrt{e_e})\,c'n^2$$

$$k_5' = \frac{L_e^\star}{L^\star} H \frac{B+1}{B} = \frac{L_e^\star}{L^\star}(Q^\star - Q_g)$$

$$k_6' = -\frac{L_e^\star}{L^\star}\frac{Q^\star - Q_g}{\mathrm{B}+1} = -\frac{L_e^\star}{L^\star} LE$$

$$k_7' = -\sigma T_e^4 (1 + c'n^2)\,b'$$

$$k_8' = -\sigma T_e^4 (1 + c'n^2)\sqrt{e_e}\,.$$

By applying (5.9) to the daily data of the Semois drainage basin at Membre, the following mean values are derived for the parameters a' and b' of formula (2.8)

	a'	b'
Winter	0.3260	0.1340
Spring	0.4285	0.0756
Summer	0.6759	0.0123
Autumn	0.5468	0.0511

(5.10)

6. Annual Regime of the Energy-Balance Components, of the Potential and Effective Evapotranspirations, and of the Soil Moisture

The daily climatological observations collected in the Semois drainage basin (outlet Membre) over the ten-year period 1970–1979 have been amended to reflect the changes listed in Table I; the amended values are used to calculate successively for scenario 1:

- the net terrestrial radiation over the 24 hr ($L^\star$) and the daylight ($L_e^\star$) intervals by means of formulas (1.1) and (2.8) and using the parameter values given in (5.7) and (5.10),
- the corresponding net radiations ($Q^\star$ and $Q_e^\star$),
- the potential evapotranspiration (ETP) by the energy balance method.

The IRMB conceptual hydrological model is then run up to the point of effective evapotranspiration estimation. As for the precipitation, it is assumed that the frequency of rainy days does not change with increased CO_2 (Mitchell, 1986) and that the monthly increments (positive or negative) are distributed between all rainy days according to the proportional rule

$$(P_{ij})_1 = (P_{ij})_0 \times [(\overline{P}_j)_0 + \Delta P_j]/(\overline{P}_j)_0$$

where the subscripts 0 and 1 refer to the scenarios 0 and 1 respectively, where P_{ij} denotes the precipitation on day i of month j, $\overline{P}_j$ the mean precipitation of month j, and ΔP_j the monthly increment induced by the doubled CO_2 concentration (as given in table I).

TABLE I: Specific changes of climate under the hypothesis of a doubling of the atmospheric CO_2 concentration.

	Decrease	Increase	Sources
Net terrestrial radiation; $L^\star$	3.1 Wm^{-2}		Chou *et al.*, 1982
Global solar radiation; $K\downarrow$	2.5 Wm^{-2}		Chou *et al.*, 1982
Flux of sensible heat; dH/H	8%		Manabe and Wetherald, 1975
Flux of latent heat of evaporation; dE/E		7%	Manabe and Wetherald, 1975
Bowen ratio; dB/B	0.06 or 18%		Manabe and Wetherald, 1975
Cloudiness; dn/n		1.5%	Washington and Meehl, 1983
Air temperature; T		see Table I.1	Manabe and Stouffer, 1980
Water vapour pressures; e and e_e		linked to T and T_e, the relative humidities being assumed invariant	Manabe and Wetherald, 1975 Washington and Meehl, 1983
Precipitation; P	see Table I.1		Manabe *et al.*, 1981 Washington and Meehl, 1983

TABLE I.1

	J	F	M	A	M	J	J	A	S	O	N	D	Year
Air temperature (deg K)	+3.1	+3.4	+3.4	+3.1	+2.8	+2.7	+2.5	+2.3	+2.3	+2.7	+2.8	+3.2	+2.86
Precipitation (mm/month)	+9.3	+10.5	+9.9	+10.2	−1.2	−2.7	−1.6	−2.2	0.0	+5.3	+8.1	+8.7	+54.3

For the air temperature, the monthly increments are added to the present-day values.

The computation gives the daily values of:

- the effective evapotranspiration E, sum of the evapotranspirations from the water storage in the vegetation cover and from the two layers of the aeration zone,
- the water content in the soil and the relative soil moisture,

and, consequently, the daily values of:

- the latent heat transfer LE,
- the sensible heat transfer H, through the energy balance equation (2.1),
- the Bowen ratio.

The seasonal means of the energy-balance components over the 1970–1979 period are given in Table II.

The monthly means of the main climatic and hydrological variables are given in Table III.

TABLE II: Compared regimes of the energy-balance components (J cm^{-2} d^{-1}) under the present climate conditions (scenario 0) and assuming a doubling of the atmospheric CO_2 concentration (scenario 1).

	Winter	Spring	Summer	Autumn	Year
Net terrestrial radiation over the 24 hr daylength					
$L_0^\star$	284.4	502.6	510.8	396.7	424.4
$L_1^\star$	270.9	470.5	469.3	378.9	398.1
$\Delta L^\star$	−13.5	−32.1	−41.5	−17.8	−26.3
$\Delta L^\star / L_0^\star$	−5%	−6%	−8%	−4%	−6%
Global solar radiation					
$K{\downarrow}_0$	299.0	1293.2	1789.5	721.3	1030.2
$K{\downarrow}_1$	283.3	1269.0	1761.9	702.1	1008.5
$\Delta K{\downarrow}$	−15.7	−24.2	−27.6	−19.2	−21.7
Net radiation over the 24 hr daylength					
$Q_0^\star - Q_g$	28.6	570.8	956.6	224.8	447.9
$Q_1^\star - Q_g$	12.8	586.3	975.4	220.6	451.6
$\Delta Q^\star$	−15.8	15.5	18.8	−4.2	3.7
Net terrestrial radiation over the daylight period					
$L_{e0}^\star$	152.7	381.3	446.7	249.9	308.6
$L_{e1}^\star$	143.0	353.5	405.7	236.7	285.6
$\Delta L_e^\star$	−9.7	−27.8	−41.0	−13.2	−23.0
$\Delta L_e^\star / L_{e0}^\star$	−6%	−7%	−9%	−5%	−7%
Net radiation over the daylight period					
$Q_{e0}^\star - Q_g$	160.3	692.1	1020.7	371.6	563.7
$Q_{e1}^\star - Q_g$	140.7	703.3	1039.0	362.8	564.1
$\Delta Q_e^\star$	−19.6	11.2	18.3	−8.8	0.4
Transfer of latent heat of evaporation					
LE_0	55.8	420.1	640.2	222.2	336.3
LE_1	80.3	475.6	670.2	240.6	368.4
ΔLE	24.5	55.5	30.0	18.4	32.1
$\Delta LE/LE_0$	44%	13%	5%	8%	10%
Transfer of sensible heat					
H_0	−27.2	150.6	316.4	2.5	111.6
H_1	−67.5	110.7	305.2	−20.0	83.1
ΔH	−40.3	−39.9	−11.2	−22.5	−28.5
Bowen ratio					
B_0	−0.069	0.270	0.384	0.092	0.228
B_1	−0.191	0.184	0.344	0.016	0.155
ΔB	−0.122	−0.086	−0.040	−0.076	−0.073

7. Discussion

It is first necessary to verify whether the simulated mean perturbation value of the net terrestrial radiation over 24 hr, $L^\star$, coincides with the value expected from energy balance model simulations (Chou *et al.*, 1982) and given in Table I.

TABLE III: Compared regimes of the evapotranspiration and of other climate and hydrological variables under the present climate conditions (scenario 0) and assuming a doubling of the atmospheric CO_2 concentration (scenario 1).

	J	F	M	A	M	J	J	A	S	O	N	D	Year
Potential evapotranspiration (mm d^{-1})													
ETP_0	0.149	0.373	0.805	1.768	2.744	3.130	3.132	2.721	1.881	0.890	0.258	0.167	1.507
ETP_1	0.211	0.551	1.061	2.069	2.950	3.377	3.337	2.997	2.015	0.995	0.359	0.233	1.685
ΔETP	0.062	0.178	0.256	0.301	0.206	0.247	0.204	0.276	0.134	0.105	0.101	0.067	0.178
$\Delta ETP/ETP_0$	42%	48%	32%	17%	8%	8%	7%	10%	7%	12%	39%	40%	12%
Effective evapotranspiration (mm d^{-1})													
ET_0	0.149	0.370	0.793	1.690	2.608	2.778	2.766	2.266	1.611	0.834	0.255	0.167	1.362
ET_1	0.211	0.546	1.042	1.965	2.773	2.936	2.868	2.391	1.655	0.920	0.353	0.233	1.496
ΔET	0.062	0.176	0.249	0.275	0.165	0.158	0.102	0.125	0.044	0.086	0.098	0.066	0.134
$\Delta ET/ET_0$	42%	48%	31%	16%	6%	6%	4%	6%	3%	10%	38%	40%	10%
ET_0/ETP_0	100%	99%	99%	96%	95%	89%	88%	83%	86%	94%	99%	100%	90%
ET_1/ETP_1	100%	99%	98%	95%	94%	87%	86%	80%	82%	92%	98%	100%	89%
Water vapour pressure (hPa)													
e_0	5.96	5.98	6.58	7.46	10.14	12.62	14.09	13.98	11.89	9.55	7.46	6.38	9.36
e_1	7.42	7.60	8.32	9.21	12.19	15.00	16.51	16.18	13.82	11.44	9.05	7.99	11.25
Δe	1.46	1.62	1.74	1.75	2.05	2.38	2.42	2.20	1.93	1.89	1.59	1.61	1.89
$\Delta e/e_0$	24%	27%	26%	23%	20%	19%	17%	16%	16%	20%	21%	25%	20%
Temperature (°C)													
t_0	0.7	1.4	3.4	6.1	10.9	14.0	15.7	15.3	11.8	8.0	3.9	1.6	7.8
t_1	3.8	4.8	6.8	9.2	13.7	16.7	18.2	17.6	14.1	10.7	6.7	4.3	10.6
Δt	3.1	3.4	3.4	3.1	2.8	2.7	2.5	2.3	2.3	2.7	2.8	3.2	2.8
Precipitation (mm d^{-1})													
P_0	3.47	3.42	3.35	2.37	3.07	2.62	2.74	2.12	2.71	2.31	4.77	3.77	3.06
P_1	3.77	3.79	3.67	2.71	3.03	2.53	2.69	2.05	2.71	2.48	5.04	4.05	3.21
ΔP	0.30	0.37	0.32	0.34	−0.04	−0.09	−0.05	−0.07	0.00	0.17	0.27	0.28	0.15
$\Delta P/P_0$	9%	11%	10%	14%	−1%	−3%	−2%	−3%	0%	7%	6%	7%	5%
ET_0/P_0	4.3%	10.8%	23.7%	71.2%	84.9%	106.0%	100.9%	106.9%	59.4%	36.1%	5.3%	4.4%	44.6%
ET_1/P_1	5.6%	14.4%	28.4%	72.5%	91.5%	116.0%	106.6%	116.7%	61.1%	37.1%	7.0%	5.8%	46.7%
Water content of the upper layer of the unsaturated zone (percentage of values at saturation) – Maximum water capacity: WSX = 23 mm													
WS_0/WSX	100%	98%	91%	71%	66%	54%	54%	54%	59%	82%	98%	100%	77%
WS_1/WSX	100%	96%	88%	69%	63%	50%	51%	49%	56%	80%	97%	99%	75%
Water content of the lower layer of the unsaturated zone (mm and percentage of values at saturation)													
WIX	194.1	194.1	195.1	198.7	203.6	208.5	211.1	210.0	205.3	201.0	196.8	193.9	201.0
WI_0/WIX	100%	100%	100%	97%	91%	83%	78%	74%	72%	80%	93%	100%	89%
WI_1/WIX	100%	100%	99%	96%	89%	79%	73%	68%	65%	74%	90%	100%	86%

Over a ten-year period the simulated decrease of $L^\star$ induced by a CO_2 doubling reaches 26.3 J cm^{-2} d^{-1} i.e. 3.04 W m^{-2}. This is very close to the value of 3.1 W m^{-2} though this basic hypothesis has not been used in Equation (5.6). The coherence and validity of the increments da and db are thus well-founded.

In consequence, it is now possible to estimate quantitatively the CO_2 induced perturbation of the annual regime of $L^\star$. In particular, $L^\star$ decreases by 41.5 J cm^{-2} d^{-1} in summer but only by 13.5 J cm^{-2} d^{-1} in winter. And, in consideration of Equation (5.8), the net terrestrial radiation over the daylight period $L_e^\star$ is found to decrease in almost the same ratios. The latter component is used for estimating the potential evapotranspiration.

All other simulated components of the energy balance are also checked against the basic hypotheses given in Table I. The latent heat transfer *LE* increases by 10% on the average, a value close to the hypothesis of 7%. Furthermore, the mean modification of the Bowen ratio is very close to its expected value: –0.07 instead of –0.06.

At first sight it might seem odd that the potential evapotranspiration (*ETP*) increases in all seasons, despite the decrease in available energy $Q^\star - Q_g$ during autumn and winter. But *ETP* also depends on the saturation deficit of the atmosphere, saturation deficit that increases markedly in consequence of the rise in temperature and of the hypothesis about the invariance of the relative humidity (see Table I).

As for the reduction of the sensible heat transfer *H*, the result is markedly larger (–26%) than the expected value (–8%). This discrepancy can be explained easily. The effective evapotranspiration is indeed an output of the hydrological model while the sensible heat transfer *H* is directly calculated as the residual term of the energy balance equation (2.1). Besides, the hypothesis about *H* and *LE* perturbations (Table I) applies to the entire Earth's surface, continents and oceans together. It is not surprising therefore that the values obtained for a particular drainage basin, with its specific physiographic, phytogeographic and climatic particulars, are to some extent different.

Also, the values of the sensible heat transfer (see Table II) point to a strengthening of the heat advection in winter and to a moderate heat advection in autumn in the $2 \times CO_2$ case. This result fully agrees with Mitchell's prediction (Mitchell, 1986) about a reinforcement of the maritime westerly flow over west Europe.

Quantitative information about the annual regime of the energy-balance components and of some hydrometeorological variables is given in Tables II and III.

The diminutions of both global solar radiation and net terrestrial radiation do not imply a decrease of the net radiation in all seasons. In spring and summer the decrease of $L^\star$ and $L_e^\star$ resulting from the joint effect of increased contents of atmospheric CO_2 and water vapour, is such that it induces a slight increase of the net radiation. During these two seasons this effect will contribute to the rise of potential evapotranspiration.

In all seasons, the potential evapotranspiration (*ETP*) increases, with a maximum during April (0.3 mm d^{-1}) and a minimum during January (0.06 mm d^{-1}). However, in relative values the rise is greatest during the winter months, the explanation being the smallness of the corresponding present-day *ETP* values. In response to the increased potential evapotranspiration, the effective evapotranspiration also augments, though not in the same ratio. This is linked to the fact that, in scenario 1, the rainfall is lowered during late spring and summer. During this period of the year the water supply from the soil cannot meet fully the evapotranspiration requirements.

The evolution of soil water resources is an essential information for agricultural practices. During winter, the lower layer of the aeration zone is permanently saturated. It is nearly the same as regards the upper layer. In other seasons, the soil moisture depletion is appreciably greater in scenario 1 than in scenario 0, a larger fraction of the available soil water being indeed consumed by the evapotranspiration. This clearly produces a greater loss of soil moisture; growing larger and larger from spring to autumn it passes through a maximum during September (7%).

With an experiment increasing fourfold the CO_2 concentration, Manabe *et al.* (1981) demonstrated the possible advent of summer dryness, due firstly to the reduction of precipitation in summer and secondly to the rise in temperature leading to the disappearance of snowcovers in winter and, hence, an earlier start to the summer drying season. More recently Mitchell (1983) predicted a similar perturbation for a doubled CO_2 concentration. The results obtained by Washington and Meehl (1984) suggest a general increase in soil moisture during all seasons with a minimum during summer; though these authors predict a general increase of the precipitation, their discrepant conclusion is perhaps due only to a crudely modelized soil moisture through GCM-simulation (Washington and Meehl, 1984).

In contrast with the GCM-simulations, the present research does not point to any winter increase of the soil moisture. During this season the soil remains permanently near saturation, and any increase in precipitation can contribute only to an increase of either surface runoff, percolation or effective evapotranspiration. The other seasons are characterized by a reduction of the soil moisture with an extremum just after the summer as predicted by Manabe *et al.* (1981).

8. Summary and conclusion

The present research attempts

(1) to determine, for $2 \times CO_2$ case and for the Belgian climate conditions, the seasonal parameter values of an empirical formula for calculating the daily values of the net terrestrial radiation $L^\star$ (over 24 hours and over the daylight period)

(2) to estimate quantitatively, for the $2 \times CO_2$ case and on the basis of a ten-year period, daily data for the Semois drainage basin, the annual regimes of net terrestrial radiation, net radiation and potential evapotranspiration (computed by the energy balance method).

From Tables II and III, it appears that the CO_2 doubling has the following consequences:

(1) a decrease in net terrestrial radiation in all seasons, with very different magnitudes according to season, the maximum decrease being found in summer

(2) a decrease in net radiation in autumn and winter, and an increase (rather small in relative values) in spring and summer

(3) an increase in potential evapotranspiration during all months, somewhat greater from March through August (the decrease in net radiation in autumn and winter is overbalanced by the increase in atmospheric saturation deficit).

Daily values of effective evapotranspiration and soil moisture are then simulated by using a conceptual hydrological model. Mean monthly values of these variables are given in Table III for the scenarios 0 and 1.

The conclusions are:

(1) an increase in effective evapotranspiration during all seasons, in a slightly lower ratio than in potential evapotranspiration despite the decrease in precipitation during late spring and summer; a larger contribution of soil water in response to potential evapotranspiration

(2) a decrease in soil moisture during all seasons (except in winter) to a maximum of 7% in September

(3) a strong decrease of the sensible heat transfer in all seasons, a moderate heat advection in autumn and a strengthening of the heat advection in winter, in concordance with a reinforcement of the maritime westerly flow over west Europe.

The results summarized above appear to be in general agreement with the GCM-simulations but provide, in addition, a quantitative description of the annual regime of various hydrometeorological variables.

It is obvious that the indirect climate perturbations induced by a doubling of the atmospheric CO_2 concentration should hold the attention of specialists dealing with agricultural and water resources management strategies (White, 1985; Schnell, 1986).

The present research is but a preliminary step to a more detailed impact analysis dealing with three Belgian river catchments with very different features. This analysis uses the hydrological model in its whole and attempts to estimate, on the basis of daily data over an 80-yr period, the CO_2 impact on both surface – and ground-water resources, and more particularly the expected specific responses of the catchments according to their own characteristics.

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