Appendix A

Stemflow

The amount of stemflow $S_f$ is directly related to $P$. Table A.1 shows the regression results for the four experimental sites at which stemflow was measured. To investigate a possible relation between $S_f$ and $L_{AI}$ distinction has been made between summer and winter. In summer $S_f$ decreased in the order: oak - poplar - pine - larch. In winter the order is: oak - poplar - larch - pine. At all sites investigated $S_f$ is low, in most cases $S_f < 0.01P$. At the pine site there exists hardly any difference between summer and winter, because the differences in the Vegetation Area Index $V_{AI}$ amounted only to 0.2 m$^2$ m$^{-2}$.

Only for oak and poplar trees the ratio of $S_f/P$ as a function of $V_{AI}$ is significantly different between the foliated and the non-foliated periods (see Fig. A.1). This relationship shows an increase in leaf area reducing the ratio of $S_f/P$. The increase of $S_f$ with decreasing leaf area is a confirmation of the findings of Giacomin and Trucchi (1992). At their beech forest site in Italy $S_f$ was found to be 12% higher in the lesser foliated period for precipitation events $> 5$ mm.

Tree characteristics that may explain the differences in $S_f$ between tree species are differences in the angles of the branches, the capacity of the bark of the trees to absorb and retain water and the foliage. For the tree species of this study the branch angle increases in the order: poplar - oak - larch - pine. As a rule of thumb one may conclude that the smaller the branch angle the greater the stemflow. Additionally the capacity of the bark of the trees to absorb and retain water will influence the

<table>
<thead>
<tr>
<th>Location</th>
<th>Summer</th>
<th>Winter</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$x_0$ (mm)</td>
<td>$a$</td>
</tr>
<tr>
<td>Larch (Bankenbos)</td>
<td>-0.0012</td>
<td>0.0005</td>
</tr>
<tr>
<td>Oak (Edesebos)</td>
<td>-0.0956</td>
<td>0.0148</td>
</tr>
<tr>
<td>Poplar (Fleditebos)</td>
<td>-0.0557</td>
<td>0.0107</td>
</tr>
<tr>
<td>Pine (Loobos)</td>
<td>-0.0039</td>
<td>0.0007</td>
</tr>
</tbody>
</table>

Table A.1: Stemflow parameters at four forest sites assuming a linear relationship with precipitation $P$: $S_f = x_0 + aP$ (mm), where $P$ (mm) is the precipitation based on weekly data.
amount of stemflow. The intercept of the regression line in the left graph of Fig. A.1 gives an indication of this capacity. At this poplar stand during the non-foliated period \( S_f \) only starts at \( P > 3.5 \) mm week\(^{-1}\).

The right hand graph of Fig. A.1 shows the influence of the amount of foliage on the stemflow fraction \( S_f / P \). Foliage influences \( S_f \) in two ways. It prevents rain from wetting branches and trunk and also the weight of the foliage may bend the branches downward, increasing the angle of the branches with the stem and thus reducing \( S_f \). The influence of the foliage is evident as during the foliated period \( S_f \) starts at \( P > 6.0 \) mm week\(^{-1}\).

In view of the low numbers found for \( S_f \) at the sites presented here, a detailed assessment of this quantity seems unnecessary as compared to the uncertainties of the other components of the water balance. Except in the winter periods \( S_f \) will not be taken into account for the oak and poplar sites, when \( S_f > 0.01P \).
Appendix B

Weight fractions of the soil

The tables below show the weight fraction of the soil at the Bankenbos, Fleditebos, Kampina and the Loobos sites. The sand fraction is defined as the fraction of dry matter with particle size between 16 μm - 2000 μm, the silt fraction with particle size 2 μm - 16 μm and the clay fraction with particle size 0 μm - 2 μm.
Table B.1: Weight fractions $x$ of sand, silt and clay and the densities $\rho$ at the Bankenbos site.

| Depth (m) | $x_{\text{sand}}$ (kg m$^{-3}$) | $x_{\text{silt}}$ (kg m$^{-3}$) | $x_{\text{clay}}$ (kg m$^{-3}$) | $\rho_{\text{sample}}$ (kg m$^{-3}$) |
|-----------|-------------------------------|-------------------------------|-------------------------------|--------------------------------|---------|
| 0-0.30    | 0.41                          | 0.01                          | 0.05                          | 1260                           |         |
| 0.55-0.70 | 0.64                          | 0.01                          | 0.02                          | 1650                           |         |
| 0.60-0.90 | 0.65                          | 0.02                          | 0.01                          | 1760                           |         |

Table B.2: Weight fractions $x$ of sand, silt and clay and the densities $\rho$ at the Fleditebos site.

| Depth (m) | $x_{\text{sand}}$ (kg m$^{-3}$) | $x_{\text{silt}}$ (kg m$^{-3}$) | $x_{\text{clay}}$ (kg m$^{-3}$) | $\rho_{\text{sample}}$ (kg m$^{-3}$) |
|-----------|-------------------------------|-------------------------------|-------------------------------|--------------------------------|---------|
| 0.07-0.34 | 0.61                          | 0.12                          | 0.18                          | 880                             |         |
| 0.36-0.70 | 0.42                          | 0.15                          | 0.32                          | 980                             |         |
| 0.90-1.10 | 0.25                          | 0.32                          | 0.29                          | 1100                            |         |

Table B.3: Weight fractions $x$ of sand, silt and clay and the densities $\rho$ at the Kampina site.

| Depth (m) | $x_{\text{sand}}$ (kg m$^{-3}$) | $x_{\text{silt}}$ (kg m$^{-3}$) | $x_{\text{clay}}$ (kg m$^{-3}$) | $\rho_{\text{sample}}$ (kg m$^{-3}$) |
|-----------|-------------------------------|-------------------------------|-------------------------------|--------------------------------|---------|
| 0-0.20    | 0.52                          | 0.01                          | 0.05                          | 1320                           |         |
| 0.20-0.50 | 0.53                          | 0.01                          | 0.01                          | 1450                           |         |

Table B.4: Weight fractions $x$ of sand, silt and clay and the densities $\rho$ at the Loobos site.

| Depth (m) | $x_{\text{sand}}$ (kg m$^{-3}$) | $x_{\text{silt}}$ (kg m$^{-3}$) | $x_{\text{clay}}$ (kg m$^{-3}$) | $\rho_{\text{sample}}$ (kg m$^{-3}$) |
|-----------|-------------------------------|-------------------------------|-------------------------------|--------------------------------|---------|
| 0-0.20    | 0.96                          | 0.01                          | 0.02                          | 1560                           |         |
| 0.40-0.60 | 0.98                          | 0.00                          | 0.02                          | 1630                           |         |
Appendix C

Data processing for the eddy correlation technique

One of the main objectives of the experimental work of the forest hydrology project was to obtain long term (about three years) of continuous measurements. All sites were located in remote areas where no main power was available. This was solved by using batteries powered by solar and wind energy and by minimizing the power consumption. Due to this some concessions had to made to the quality of the flux data. Reduction of the power consumption was achieved among others by using palmtop PC’s and not storing the raw data continuously, but instead storing the means, variances and covariances. However, the disadvantage of the use of these palmtops is, besides the limited data storage capacity, the processor speed. This lead to the second concession, namely not calculating the covariances of the wind speed components in $x$- and $y$-direction $u$ and $v$ with temperature $T$ and scalar $\kappa$ and the covariance of $T$ and $\kappa$. To check if the mentioned concessions caused unacceptable data quality loss, raw data were collected for short periods (maximum a couple of weeks at each site).

For a complete description of the eddy-correlation technique in combination with infra red gas analyser and the processing software for the raw data the reader is referred to Moncrieff et al. (1996) and Aubinet et al. (2000). Here the data processing is described as used for a 3D sonic and an open path hygrometer applying eddy correlation technique. This was the basis for the long term data collection of this study.

The equipment used for the measurement of the momentum, sensible and latent heat flux is a 3-D sonic anemometer (SOLENT 1012R2, Gill) and a Krypton hygrometer (KH$_2$O, Campbell). The three wind components are derived from the calibrated relation between the wind speed and the transit time for a sound pulse travelling over the known distance between one of the three transmitters to the corresponding receiver. The temperature fluctuations needed for the sensible heat flux density are derived from the known relation between the speed of sound in air $c_s$ (m s$^{-1}$) and
the absolute air temperature $T$ in K (Kaimal and Finnigan, 1994):

$$c_s^2 = \gamma \frac{R}{M_d} T (1 + 0.32 \frac{e}{p})$$  \hspace{1cm} (C.1)

where $M_d$ the molar mass of dry air (0.028965 kg mol$^{-1}$), $R$ is the universal gas constant (8.314 J mol$^{-1}$ K$^{-1}$), $\gamma_c = c_p/c_v$ is the ratio of specific heat, with $c_p$ and $c_v$ the specific heat (J kg$^{-1}$ K$^{-1}$) of dry air at constant pressure and constant volume respectively ($\gamma_c R/M_d = 403$ m$^2$ s$^{-2}$ K$^{-1}$), $e$ denotes the vapour pressure of water (hPa) and $p$ the atmospheric pressure (hPa).

The sonic temperature is defined as $T_{son} = c_s^2/403$ and is close to the virtual temperature $T_v$ (K):

$$T_v = T (1 + 0.38 \frac{e}{p})$$  \hspace{1cm} (C.2)

The absolute humidity $\kappa_{abs}$ measured by the Krypton hygrometer relates to the vapour pressure by:

$$\kappa_{abs} = \frac{e M_w}{R T}$$  \hspace{1cm} (C.3)

where $M_w$ is the molar mass of water (0.0180153 kg mol$^{-1}$). In the case of high vapour pressure the difference between the sonic temperature and the virtual temperature becomes more pronounced, but are still small enough ($\pm 0.01$ K) for most micrometeorological purposes (Kaimal and Kristensen, 1991). Corrections for the effect of humidity on the sensible heat flux are applied following Schotanus et al. (1983). The three wind speed components, the speed of sound (all in m s$^{-1}$) and the humidity concentration (in mV) are measured at 20.825 Hz. Each 30 minute interval the sound of speed is converted into sonic temperature using Eq. (C.1), where after the fluctuations of all quantities are calculated using a non-centralized running mean algorithm with a 200 s time constant. Means and variances are calculated and stored of $u$, $v$, $w$, $T_{son}$ and $\kappa$, as well as the covariances $u'w'$, $u'v'$, $v'w'$, $w'T_{son}'$ and $w'\kappa'$. 

After collecting the data in the field the mean absolute humidity (kg m$^{-3}$) measured by the Krypton hygrometer is calculated using the calibration parameters supplied by the manufacturer. The same parameter of the humidity calibration span is used for the covariance of vertical wind speed and humidity. The wind speed and direction are calculated from the means of $u$ and $v$. The wind direction is also corrected for the non alignment with the magnetic North.

C.1 Oxygen absorption

The Krypton H$_2$O hygrometer emits ultraviolet light between 123.58 and 116.47 nm. This implies that the sensor is besides to hydrogen also sensitive to ozone and
C.2. Axis rotation

Due to the high relative concentration of oxygen a correction is needed. Differences in the density of oxygen depend on temperature and air pressure. For the absolute humidity measured by the Krypton this correction is assumed constant in the calibration coefficients (i.e. constant temperature and pressure) and implicit in the calibrated offset supplied by the manufacturer. To incorporate a site dependent oxygen correction in the absolute humidity measurement by the Krypton hygrometer, the offset caused by the oxygen concentration during calibration should be subtracted. This gives the following equation

$$\kappa_{\text{abs}} = \ln V_0 - \ln \frac{V_q}{x_p k_v} + \frac{k_{O_2}}{k_v} (\rho_{O_20} - \rho_{O_2})$$  \hspace{1cm} (C.4)

where $V_0$ is the calibration offset (V), $V_q$ the measured voltage (V), $x_p$ the path length (m), $k_{O_2}$ and $k_v$ are the absorption coefficients for oxygen and water vapour (0.0085 and 0.143 m$^3$ g$^{-1}$ cm$^{-1}$ respectively), $\rho_{O_20}$ the oxygen density during calibration (kg m$^{-3}$), $\rho_v$ the density during the measurement (kg m$^{-3}$). The oxygen density is given by

$$\rho_{O_2} = \frac{C_{O_2} M_{O_2} p}{RT}$$  \hspace{1cm} (C.5)

where $C_{O_2}$ is the relative concentration of oxygen (0.21), $M_{O_2}$ is the molecular weight of oxygen (32 kg mol$^{-1}$), $p$ the air pressure (Pa), $R$ the universal gas constant (8.314 J mol$^{-1}$K$^{-1}$) and $T$ the air temperature (K).

The covariance of the absolute humidity and the vertical wind speed is corrected for the oxygen sensitivity by

$$\overline{w' \kappa_{\text{abs}}} = \frac{-w' V_q}{x_p k_v V_q} + \frac{C_{O_2} M_{O_2} k_{O_2} k_v}{RT^2} \frac{w' T'}{k_v}$$  \hspace{1cm} (C.6)

This correction is in the order of 10% of the sensible heat flux. For this study no correction is made for the ozone sensitivity.

C.2 Axis rotation

Although the sites are relatively flat and some effort was made to install the sonics in the mean wind direction perpendicular to the surface the data are processed using stream line coordinates.

To align the $u$-component to the mean horizontal wind speed and to align the mean wind vector parallel to the mean stream lines of the wind speed, a two axis coordinate rotation is applied to the means, variances and covariances of the wind
speed components following McMillen (1988) and Kaimal and Finnigan (1994). The horizontal wind vector is defined as
\[ \vec{u}_{\text{hor}} = \sqrt{u^2 + v^2} \] (C.7)
and the total wind vector
\[ \vec{u}_{\text{tot}} = \sqrt{u^2 + v^2 + w^2} \] (C.8)
from this the angle between \( \vec{u} \) and \( \vec{u}_{\text{hor}} \) is
\[ \alpha = \arccos \left( \frac{\vec{u}}{\vec{u}_{\text{hor}}} \right) \] (C.9)
and the vertical tilt
\[ \gamma = \arccos \left( \frac{\vec{u}_{\text{tot}}}{\vec{u}_{\text{hor}}} \right) \] (C.10)
With these angles the first two rotations can be calculated. In effect the first rotation sets \( v \) to zero and the second rotation sets \( w \) to zero.

The covariances of the wind speed components and temperature and humidity (i.e. sensible and latent heat flux) were not rotated. Also the third rotation as proposed by Kaimal and Finnigan (1994), which forces the lateral momentum flux to zero was not performed. This rotation is necessary when the streamlines incline with respect to the local surface. For the present sites this was not considered to have a major influence on the results.

After the rotation corrections, a first estimate of the friction velocity \( u_* \) (m \( s^{-1} \)) is calculated as:
\[ u_* = \sqrt{\overline{u'v'^2} + \overline{u'w'^2}} \] (C.11)
as well as the Monin-Obukhov stability length \( L \) (m):
\[ L = -\frac{T_{\text{son}} u_*^3}{\kappa g \overline{w'u'T_{\text{son}}}} \] (C.12)
After the frequency response corrections the friction velocity and the stability length are recalculated.

C.3 Frequency response corrections

Due to sensor response, sensor separation, path length averaging, and signal processing, a part of the transporting eddies may be influenced or not detected by the system. To correct for this the flux loss is calculated using transfer functions describing the spectral loss of the system and model (co-)spectra (Moore, 1986). To prevent
C.3. Frequency response corrections

Aliasing, the effective transfer functions are calculated for frequencies \( n \) not exceeding the Nyquist frequency (i.e. half of the sampling frequency \( n_s \)). The convolution integral of the model spectra and the effective transfer function is calculated over the frequency range 0.00001 Hz to the Nyquist frequency.

Detrending of the data was done using an auto regressive moving average filter with a time constant \( \tau_d \) of 200 s. The response gain for this high pass filtering is:

\[
T_d(n) = \frac{(2\pi n \tau_d)^2}{1 + (2\pi n \tau_d)^2 / \alpha_d}
\]

where the time coefficient \( \alpha_d \) is taken as one.

The co-spectral transfer function for the transformation of analogue to digital signals is given by:

\[
T_a(n) = 1 + \left( \frac{n}{n_s - n} \right)^3
\]

where \( n_s \) is the sampling frequency.

The frequency response gains for the wind speed components, the sonic temperature (all from the Solent sonic anemometer) and for the humidity (from the Krypton hygrometer) are considered to be negligible.

The transfer functions for spatial averaging over a path with length \( x_p \) (\( x_p = 0.149 \) m for the sonic anemometer and \( x_p \approx 0.0125 \) m for the Krypton hygrometer) are for the humidity, the sonic temperature and the wind components \( u \) and \( v \):

\[
T_p(n) = \frac{1}{2\pi n u x_p} \left[ 3 + \exp \left( -2\pi n u x_p \right) - \frac{4 \left( 1 - \exp \left( -2\pi n u x_p \right) \right)}{2\pi n u x_p} \right]
\]

and for the vertical wind component \( w \):

\[
T_p(n) = \frac{4}{2\pi n u x_p} \left[ 1 + \frac{1}{2} \left\{ \exp \left( -2\pi n u x_p \right) - \frac{3 \left( 1 - \exp \left( -2\pi n u x_p \right) \right)}{2\pi n u x_p} \right\} \right]
\]

The transfer function for sensor separation of the Krypton hygrometer and the sonic anemometer is given by:

\[
T_s = \exp \left( -9.9 \left( \frac{n}{n_s} \right)^{1.5} \right)
\]

where \( s \) (\( s = 0.2 \) m) is the distance between the two sensors.
The composite transfer functions for the equipment used for this study are calculated as:

\[ T_{uu} = T_{pu}T_aT_d \]  
\[ T_{vv} = T_{pv}T_aT_d \]  
\[ T_{ww} = T_{pw}T_aT_d \]  
\[ T_{TT} = T_{pT}T_aT_d \]  
\[ T_{qq} = T_{pq}T_aT_d \]  
\[ T_{uw} = \sqrt{T_{uu}T_{ww}T_aT_d} \]  
\[ T_{vw} = \sqrt{T_{vv}T_{ww}T_aT_d} \]  
\[ T_{wT} = \sqrt{T_{TT}T_{ww}T_aT_d} \]  
\[ T_{qw} = \sqrt{T_{qq}T_{ww}T_aT_d} \]

For the theoretical spectra and co-spectra the formulations of Kaimal et al. (1972) are used. In their formulations the spectra are functions of the normalized frequency \( f = \frac{n(x-d)}{u} \) and stability \( \zeta = \frac{z-d}{L} \).

For stable conditions (\( \zeta < 0 \)):

\[ S_{uu} = \frac{1}{n} \frac{f}{0.2(0.838 + 1.172\zeta) + f^{5/3}3.124[0.2(0.838 + 1.172\zeta)]^{-2/3}} \]  
\[ S_{ww} = \frac{1}{n} \frac{f}{(0.838 + 1.172\zeta) + f^{5/3}3.124(0.838 + 1.172\zeta)^{-2/3}} \]  
\[ S_{TT} = \frac{1}{n} \frac{f}{[0.0961 + 0.644\zeta^{3/5}] + f^{5/3}3.124[0.0961 + 0.644\zeta^{3/5}]^{-2/3}} \]  
\[ S_{wT} = \frac{1}{n} \frac{f}{0.284 [1.0 + 6.3\zeta^{3/4}] + f^{2.2}12.34 \{0.284 [1.0 + 6.3\zeta^{3/4}] \}^{-1/2}} \]  
\[ S_{uw} = \frac{1}{n} \frac{f}{0.124 [1.0 + 7.9\zeta^{3/4}] + f^{2.2}12.34 \{0.124 [1.0 + 7.9\zeta^{3/4}] \}^{-1/2}} \]  

For unstable conditions (\( \zeta < 0 \)) the spectra are dependent of the boundary layer height. For the horizontal and vertical wind speed the formulations of Hojstrup (1981) are used. Here Moore (1986) is followed and the spectrum of temperature and the co-spectra of vertical wind speed with temperature and horizontal wind speed as
C.3. Frequency response corrections

given by Kaimal et al. (1972) are also used for unstable conditions:

\[
S_{uu} = \frac{1}{n} \left\{ \frac{210.0 f}{(1.0 + 33.0 f)^{5/3}} + \frac{f(-\zeta)^{2/3}}{(\frac{z}{z_i})^{5/3} + 2.2 (f)^{5/3}} \right\} \quad (C.32)
\]

\[
S_{ww} = \frac{1}{n} \left\{ \frac{1}{9.546 + 1.235(-\zeta)^{2/3} (\frac{z}{z_i})^{-2/3}} \right\} \quad (C.33)
\]

\[
S_{TT} = \frac{1}{n} \left\{ \frac{14.94 f}{(1.0 + 24.0 f)^{5/3}} \right\} \quad f < 0.15 \quad (C.34)
\]

\[
S_{TT} = \frac{1}{n} \left\{ \frac{6.827 f}{(1.0 + 12.5 f)^{5/3}} \right\} \quad f \geq 0.15 \quad (C.35)
\]

\[
S_{wT} = \frac{1}{n} \left\{ \frac{12.92 f}{(1.0 + 26.7 f)^{1.377}} \right\} \quad f < 0.54 \quad (C.36)
\]

\[
S_{wT} = \frac{1}{n} \left\{ \frac{4.378 f}{(1.0 + 3.8 f)^{2.4}} \right\} \quad f \geq 0.54 \quad (C.37)
\]

\[
S_{uw} = \frac{1}{n} \left\{ \frac{20.78 f}{(1.0 + 31.0 f)^{1.577}} \right\} \quad f < 0.25 \quad (C.38)
\]

\[
S_{uw} = \frac{1}{n} \left\{ \frac{12.66 f}{(1.0 + 9.6 f)^{2.4}} \right\} \quad f \geq 0.25 \quad (C.39)
\]

where \( z_i \) (m) denotes the boundary layer height, which is not known most of the time and taken here as constant at 1000 m. Again according to Moore (1986) the same functions may be used for \( S_{uu} = S_{vv}, S_{TT} = S_{qq}, S_{uw} = S_{vw} \) and \( S_{wT} = S_{wq} \).

The flux loss correction factor \( c_{F_{xy}} \) is then the ratio of the integrated theoretical (co-)spectral distribution functions \( S_{xy} \) to the convolution integral of the (co-)spectral transfer functions \( T_{xy} \) and \( S_{xy} \)

\[
c_{F_{xy}} = \frac{\int_0^\infty S_{xy} dn}{\int_0^\infty T_{xy} S_{xy} dn} \quad (C.40)
\]

With the now corrected variances and covariances the friction velocity and stability length are recalculated.
C.4 Sensible heat flux and air temperature

The sonic temperature is close to virtual temperature (Kaimal and Kristensen, 1991) and is transformed to mean air temperature by a simplified relation:

\[ T = \frac{T_{\text{son}}}{1 + 0.51 \frac{\overline{\varphi}_{\text{abs}}}{\rho_a}} \approx T_{\text{son}} \left( 1 - 0.51 \frac{\overline{\varphi}_{\text{abs}}}{\rho_a} \right) \tag{C.41} \]

where \( \overline{\varphi}_{\text{abs}} \) (kg m\(^{-3}\)) denotes the absolute humidity and \( \rho_a \) (kg m\(^{-3}\)) the density of dry air at the sonic temperature. If available the humidity measured by the automatic weather station is used, if not the measurement of the eddy correlation system is used. Cross wind and humidity corrections are done on the covariance of the vertical wind speed and the temperature (Schotanus et al., 1983):

\[ \overline{w' T'} = \left( \frac{w' T'_{\text{son}}}{\rho_a} T_{\text{son}} \right) \frac{0.51 \overline{\varphi}_{\text{abs}}}{\rho_a} + 2 \overline{uT_{\text{son}} w' w'} \tag{C.42} \]

If the latent heat flux is not available, the humidity correction is derived from the latent heat flux calculated as the residue of the energy balance.

The sensible heat flux \( H \) (W m\(^{-2}\)) is:

\[ H = c_p \rho_a \overline{w' T'} \tag{C.43} \]

where \( c_p \) is specific heat of moist air (J kg\(^{-1}\) K\(^{-1}\)) and \( \rho_a \) the density of dry air (kg m\(^{-3}\)).

It should be noted that, except for the short periods during which raw data were collected, due to the limited number of covariances stored in the field the sonic temperature variance was not corrected for cross wind contamination nor for humidity fluctuations.

C.5 Latent heat flux

To correct for the fact that the vertical wind speed is unequal to zero if a sensible heat flux exists, Webb et al. (1980) proposed a correction for all scalars where the density is measured instead of the mixing ratio, which is the case when using a Krypton H\(_2\)O hygrometer

\[ \overline{w' \varphi_{\text{abs}}'} = 1 + \frac{M_d}{M_w} \frac{\overline{\varphi}_{\text{abs}}}{\rho_a} \left( \frac{w' \varphi_{\text{abs}}'}{\overline{\varphi}_{\text{abs}}' \overline{w' T'}} + \frac{\overline{\varphi}_{\text{abs}}' \overline{w' T'}}{T_{air}} \right) \tag{C.44} \]

The latent heat flux (W m\(^{-2}\)) is calculated as:

\[ \lambda_E E = \overline{w' \varphi_{\text{abs}}'} \lambda_E \tag{C.45} \]

where \( \lambda_E \) is the latent heat of water vapour (J kg\(^{-1}\)).
Appendix D

Storage of heat in biomass

The heat storage in the biomass, $J_{\text{veg}}$ W m$^{-2}$ can be divided in a rapidly changing heat storage in the leaves and branches and a more slowly storage in the stems:

$$J_{\text{veg}} = J_{\text{leaf+branch}} + J_{\text{stem}} + J_{\text{undergrowth}}$$  \hspace{1cm} (D.1)

As the biomass of the undergrowth is small as compared to that of the trees, the change in heat storage of the undergrowth is neglected. The change in heat storage in branches and leaves is given by:

$$J_{\text{leaf+branch}} = \int_0^{z_{\text{tree}}} \rho_{\text{leaf+branch}} c_{\text{leaf+branch}} \frac{dT_{\text{veg}}}{dt} dz$$  \hspace{1cm} (D.2)

where $c_{\text{leaf+branch}} = 2647$ J kg$^{-1}$ K$^{-1}$ based on $c$ of dry wood corrected for the water content of the leaves and the branches.

For the estimation of the heat storage in the trunks, the method of Moore (1986) was followed, i.e. that the heat flux across the surface of a tree trunk with a diameter larger than 14 cm can be treated as an infinite slab. Ignoring the surface conductance the heat flux across a unit trunk surface at height $z$ is given by:

$$F(z) = \sqrt{\rho_{\text{stem}} c_{\text{stem}} k_T} A_T \cos \left( \omega t + \phi_T + \frac{\pi}{4} \right)$$  \hspace{1cm} (D.3)

where $k_T$ (W m$^{-1}$ K$^{-1}$) is the thermal conductivity of the stems and $\phi_T$ is the phase angle of the diurnal cycle of $T$. As $T_{\text{veg}}$ was not measured, is was assumed that it was equal to $T_a$. For those cases where no measurements of $T_a$ under the canopy were available, the $T_a$ measured above the canopy was used. Under the assumption that the variation of $F$ with height is negligible the change of heat storage in the trunk has been calculated as:

$$J_{\text{stem}} = F T_{AI}$$  \hspace{1cm} (D.4)

where $T_{AI}$ (-) is the tree area index.

If no specific measurements were available, the mass of the tree stems has been calculated using the volumes and the specific mass of the stems. The mass of the branches and leaves was estimated as a fraction of the stem biomass. The specific
heat of the stems was estimated as the weighted average of the specific heat of dry wood and water and adding a correction for heat of wetting hygroscopic material (Moore, 1986). The relation of Skaar (1972) $c_0 = 4.857 + 1113 \text{ (J kg}^{-1}\text{K}^{-1})$ for the specific heat of cellulose was used for that of dry wood. The thermal conductivity $k_T$ of the stems was calculated applying (Siau, 1971):

$$k_T = [\rho_{stem}(2.0 + 5.5\theta_{stem}) + 238] \cdot 10^{-4} \quad (D.5)$$

where $\theta_{stem}$ (m$^3$ m$^{-3}$) is the moisture content of the stem.

To estimate the stem volumes $V_{stem}$ (m$^3$) the allometric functions of Dik (1990) and Dik (1996) were used:

$$V_{stem} = a_1 D_{BH}^{a_2} z_{tree}^{a_3} \quad (D.6)$$

where $a_1$, $a_2$ and $a_3$ are tree species specific constants, $D_{BH}$ (cm) is the diameter at breast height, $z_{tree}$ (m) the tree height. The values of $a_1$, $a_2$ and $a_3$ for the main tree species of this study are given in Table D.1.

The density of the stems is taken from Laming et al. (1978). Estimates of the percentage biomass of branches and leaves are from Cannell (1982).

For the pine trees at the Loobos site specific measurements were available and allometric relations for the dry weight of the biomass $M_X$ (kg) were derived, using the same type of model as for $V_{stem}$:

$$M_X = a_1 D_{BH}^{a_2} z_{tree}^{a_3} \quad (D.7)$$

In Table D.2 the values of the constants to estimate the dry weight of the biomass are shown.

For beech the allometric relations found by Bartelink (1998) for the central part of the Netherlands have been used (see Table D.3).
Table D.2: Parameters for the allometric relations between the dry biomass $M_X$ (kg) of leaves, branches and stems for Scots pine.

<table>
<thead>
<tr>
<th></th>
<th>$a_1$</th>
<th>$a_2$</th>
<th>$a_3$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Leaves</td>
<td>0.0178</td>
<td>3.889</td>
<td>-2.575</td>
</tr>
<tr>
<td>Branches</td>
<td>0.0269</td>
<td>4.439</td>
<td>-3.053</td>
</tr>
<tr>
<td>Stems</td>
<td>0.0322</td>
<td>1.726</td>
<td>1.099</td>
</tr>
</tbody>
</table>

Table D.3: Parameters for the allometric relations between the dry biomass $M_X$ (kg) of leaves, branches and stems for beech.

<table>
<thead>
<tr>
<th></th>
<th>$a_1$</th>
<th>$a_2$</th>
<th>$a_3$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Leaves</td>
<td>0.0167</td>
<td>2.951</td>
<td>-1.101</td>
</tr>
<tr>
<td>Branches</td>
<td>0.0114</td>
<td>3.682</td>
<td>-1.031</td>
</tr>
<tr>
<td>Stem</td>
<td>0.0109</td>
<td>1.951</td>
<td>1.262</td>
</tr>
</tbody>
</table>

For American oak at the Edesebos site a specific estimation of the total biomass has been derived during an earlier study by Hendriks et al. (1990), and this estimate of the total biomass has also been used for this study. The volume of twigs and branches with a diameter less than 5 cm was estimated as 15% of the volume of the stem and branches greater than 5 cm. The wet biomass including small branches and twigs was estimated in 1988 as 19.7 kg m$^{-2}$ and in 1989 as 20.9 kg m$^{-2}$.
# Appendix E

## Soil water availability

**Table E.1:** Soil water content $\theta$ (m$^3$ m$^{-3}$) for different soil water pressures $\psi$ (Pa) at different soil depths at the different forest sites. The negative depth at the Edesebos site indicates that the sample is taken in litter layer. *) At the Bankenbos site the second sample at $z = 0.94$ m has a high loam content.

| Site       | Depth (m) | $|\psi|$ | $10^{1.5}$ | $10^{1.7}$ | $10^{2.1}$ | $10^{3.3}$ | $10^{4.2}$ |
|------------|-----------|--------|------------|------------|------------|------------|------------|
| Bankenbos  | 0.14      | 0.48   | 0.44       | 0.32       | 0.07       | 0.02       |
| Forest: larch | 0.36   | 0.35   | 0.30       | 0.18       | 0.03       | 0.01       |
| Soil: loamy           | 0.60   | 0.30   | 0.23       | 0.08       | 0.01       | 0.01       |
| sand       | 0.94      | 0.27   | 0.23       | 0.12       | 0.01       | 0.01       |
| Edesebos   | -0.03     | 0.50   | 0.42       | 0.25       | 0.03       | 0.01       |
| Forest: oak           | 0.15   | 0.32   | 0.25       | 0.14       | 0.02       | 0.01       |
| Soil: loamy           | 0.37   | 0.20   | 0.12       | 0.03       | 0.00       | 0.00       |
| sand       | 0.55      | 0.25   | 0.20       | 0.11       | 0.01       | 0.00       |
|            | 1.00      | 0.14   | 0.11       | 0.07       | 0.02       | 0.01       |
| Fleditebos | 0.07      | 0.46   | 0.43       | 0.37       | 0.25       | 0.18       |
| Forest: poplar          | 0.40   | 0.43   | 0.42       | 0.38       | 0.29       | 0.24       |
| Soil: clay            | 0.66    | 0.52   | 0.51       | 0.48       | 0.39       | 0.34       |
|            | 0.74    | 0.62   | 0.62       | 0.61       | 0.57       | 0.54       |
| Kampina    | 0.07      | 0.40   | 0.35       | 0.24       | 0.05       | 0.02       |
| Forest: mixed          | 0.34    | 0.39   | 0.37       | 0.19       | 0.12       | 0.07       |
| Soil: sand                | 0.54   | 0.33   | 0.31       | 0.16       | 0.01       | 0.01       |
| Loobos     | 0.14      | 0.35   | 0.21       | 0.04       | 0.01       | 0.01       |
| Forest: pine          | 0.54    | 0.35   | 0.20       | 0.02       | 0.01       | 0.01       |
| Soil: sand                |        |        |            |            |            |            |
Appendix F

Small scale spatial variability in storage capacity of intercepted water

The year to year changes in $L_{AI}$ and $c_{veg}$ for a specific site are relatively small if compared to the variation in $C$. The use of the fixed throughfall buckets allows to compare for each individual bucket the throughfall fraction $T_f/P$ for the same tree species with different $V_{AI}$. At such a small spatial scale $P$ and other meteorological conditions can be assumed identical for each time step, differences in $T_f/P$ between the buckets can be attributed directly to differences in $C$ and $V_{AI}$. Fig. F.1 shows the relation between $V_{AI}$ and $T_f/P$ for each individual bucket at the pine forest of the Loobos site. The results are based on the sum of $T_f$ measured the week preceding the date when the $V_{AI}$ measurements were made. Weekly $P$ varied from 1.8 to 57 mm in this period.

For all weeks in the period December 1996 to December 1997 the regression results

![Figure F.1: Weekly throughfall $T_f$ measured at each bucket as a fraction of gross precipitation $P$ on a weekly basis depicted as a function of the vegetation area index ($V_{AI}$). The data are averaged values (with standard errors) over 12 weeks in the period of December 1996 to December 1997 for the pine stand at Loobos.](image-url)
(R² = 0.61) are:

\[
\frac{T_f}{P} = 0.90 - 0.099V_{AI}
\]  

(F.1)

For this site with its relatively low \(L_{AI}\), an almost similar relation can be obtained by using the gap fraction instead of \(V_{AI}\). Plotting the interception loss \(E_i\), i.e. \(P\) minus \(T_f\), shows slightly more scatter (R² = 0.46):

\[
E_i = 2.44 + 2.13V_{AI}
\]  

(F.2)

In all cases drip points show up as outliers. At these drip points sometimes the throughfall exceeded the gross precipitation amount resulting in a negative interception loss as is observed by many other studies (e.g. Lloyd and Marques, 1988). The slope of the regression line and the variance explained (R² = 0.19) reduces for high values of \(P\) e.g. for \(P > 40\) mm:

\[
\frac{T_f}{P} = 0.88 - 0.051V_{AI}
\]  

(F.3)

This change in the relationship may be explained as at high rainfall intensities the influence of the amount of water intercepted by the leaves and eventually evaporated will be relatively small. At the same time the amount of water falling freely to the soil surface and the water dripping from the leaves will be relatively large. These two effects combined reduce the differences in water retention between the covered and non-covered areas and at the same time increase the relative effects of drip points.

At the poplar forest of the Fleditebos site the canopy varies from an almost closed cover in summer to bare in winter. This change in leaf area gives a negative correlation in summer; similar to the pine forest, but for the winter period the correlation between \(T_f/P\) and \(V_{AI}\) is positive (see Fig. F.2).

This positive correlation during the non-foliated period of the trees is caused by the \(V_{AI}\) that is measured, representing the amount of branches above the bucket during this period. The bare branches do not block the throughfall as in the summer period, but enhance throughfall by functioning as drip points. These data of the poplar forest show that \(E_i\) in winter even for deciduous trees are an important component of the water balance (10 - 20% of \(P\)).

The change in the direction of slope over the year may be explained by the smaller branches and twigs acting as preferential flow paths towards drip points during the non-foliated period. However, the wide scatter (R² = 0.09 – 0.16) for the individual dates especially during the leaf bearing period seems to point to the conclusion that at least at this deciduous site there is no clear simple relation between the leaf area and \(C\) at this spatial and temporal scale. In contrast, at the pine forest of the Loobos site, changes in \(V_{AI}\) or \(c_{veg}\) have a linear effect on \(T_f/P\) and thus on \(C\). This relationship
Figure F.2: Weekly throughfall $T_f$ measured at each bucket as a fraction of gross precipitation $P$ on a weekly basis as a function of the vegetation area index $V_{AI}$. The data are averaged values for 3 periods of 1, 2 and 1 week in 1997 for the poplar forest at the Fleditebos site.

makes it possible to extrapolate the results of the pine forest to different sites with approximately the same tree characteristics (tree height and crown diameter), but with different $V_{AI}$ or $c_{veg}$. 