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MEPDIM

Version 1.0
Model Description

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1. INTRODUCTION

The new generation of dispersion models based on parameterization of the structure of the atmospheric boundary layer need improved meteorological input. The NILU Meteorological Processor for Dispersion Modelling (MEPDIM) is an attempt to fill this need.

Current regulating models normally use simplified meteorological input. The use Pasquill-Gifford stability classes which are valid over land with small roughness and which only crudely characterize the state of the atmospheric boundary layer. These model also use power-law representations of the wind profile that are only function of the stability class.

The physical basis for the meteorological preprocessor is provided by parameterization of the structure of the atmospheric boundary layer (ABL) including its interaction with the ground. Detailed information on this subject is given in van Ulden and Holtslag (1985), Holtslag and de Bruin (1988) and Gryning et al. (1987). The main parameters included in this metprocessor are the ABL depth (h), the surface heat flux (H_o) and the surface momentum flux τ_o.

This metprocessor is based on two optimal methods; the profile method and the energy budget method. The profile method needs as input vertical profiles of wind and temperature measurements from a tower, while the energy budget method need either cloud cover or direct measurements of net radiation.

2. ESTIMATION OF U_*, θ_* AND L

This metprocessor contains two methods to obtain the three main parameters in the atmospheric surface layer; the friction velocity (U_*), temperature scale (θ_*) and the Obukhov Length (L). The profile method is based on measurements of profiles of wind speed and temperature in a meteorological tower. The energy budget, based on the surface energy budget, is used when no information of the temperature difference is available. The theory for the two optional methods is described below.

2.1 The profile method

Based on the Monin-Obukhov similarity theory for the atmospheric surface layer (Monin and Yaglom, 1979), the friction velocity (U_*), temperature scale (θ_*) and the Obukhov Length (L) can be derived from profiles of wind speed U(z) and potential temperature θ(z).

The friction velocity can be related to a wind speed U_Z at level z as follows:

$$u_* = \frac{kU_z}{\ln(\frac{z}{z_o}) - \psi_M(\frac{z}{L}) + \psi_M(\frac{z_o}{L})}, \qquad (1)$$

where z_O and ψ_M are the surface roughness length and the stability function for momentum, respectively.

The turbulent temperature scale between two levels Z_2 and Z_1 is given by:

$$\theta_* = k\Delta\theta \left[\ln(\frac{z_2}{z_1}) - \psi_H(\frac{z_2}{L}) + \psi_H(\frac{z_1}{L}) \right], \qquad (2)$$

where $\Delta\theta$ is the temperature difference between two heights Z_2 and Z_1, k is the von Karman constant, ψ_H is the stability correction function of the temperature profile and L is the Obukhov length defined by:

$$L = \frac{u_*^2}{k\frac{g}{T}\theta_*} \qquad (3)$$

The most commonly used stability correction functions of wind and temperature profiles are (Dyer, 1974; Yaglom, 1977; Businger et al., 1971; Paulson, 1970).

L>0 (unstable):

$$\psi_H = 2\ln(\frac{1+x^2}{2}), \qquad (4)$$

$$\psi_M = 2\ln\left(\frac{1+x}{2}\right) + \ln\left(\frac{1+x^2}{2}\right) - 2\tan^{-1}(x) + \pi/2, \quad (5)$$

where

$$x = (1 - 16\, z / L)^{1/4},$$

and for L>0 (stable):

$$\psi_H = -5z / L, \quad (6)$$

$$\psi_H = \psi_M \quad (7)$$

Equation (6) are adequate for $z/L \leq 0.5$ (e.g. Dyer, 1974).

For larger values of z/L, several empirical forms are proposed in literature. Carson and Richards (1978) reviewed the topic and concluded that (6) remains applicable and that the findings of Hicks (1976) are most suitable to describe ψ_M. The latter has been confirmed by Holtslag (1984) for Cabauw wind profiles up to $z/L \approx 10$.

Holtslag and Bruin (1988) have proposed the following expression of ψ_M:

$$-\psi_M = a\frac{z}{L} + b\left(\frac{z}{L} - \frac{c}{d}\right)\exp\left(-d\frac{z}{L}\right) + \frac{bc}{d}, \quad (8)$$

where a = 0.7, b = 0.75, c = 5 and d = 0.35.

Equation (8) is similar to the one proposed by van Ulden and Holtslag (1985) for $z/L \leq 10$. For larger values of z/L, eq. (8) results in linear profiles for wind and temperature.

The equations (2), (6) and (8) describes the temperature profile, while equations (1), (4) and (5) describes the wind profiles in the atmospheric surface layer.

2.2 The energy budget method

The surface energy balance over land can be written as

$$H + \lambda E + G = Q^*, \quad (9)$$

where H and λE are the fluxes of sensible and latent heat, respectively, G is the soil heat flux and Q^* is the net radiation. $H_o+\lambda E$ is the energy that is supplied to or extracted from the air, while Q^*-G is the source or sink for this energy.

The **sensible heat flux (H)** is given as:

$$H = -\delta \cdot C_p\, U_* \, \theta_*. \tag{10}$$

The **latent heat flux (λE)**, is described using the modified Priestley-Taylor (1972) model (van Ulden, Holtslag, 1983). This model is based on the experience that both the thermodynamic and the aerodynamic evaporation are strongly correlated with the equilibrium evaporation (wet surface, saturated air):

$$\lambda\, E_e = \frac{S}{1+S}(Q^* - G), \tag{11}$$

where S is the slope of the saturation enthalpy curve. This is the evaporation that would occur when the surface is wet and the air saturated.

Taking into account that λE_e and the humidity deficit (Δq) have a similar diurnal cycle, it is useful to split Δq into a part Δq_e that is correlated with λE_e and a term Δq_d that is not correlated. Then the evaporation can be correlated as (van Ulden and Holtslag, 1985):

$$\lambda\, E = \alpha \left[\frac{S}{S+1}(Q - G) - \beta \rho \lambda\, \Delta q_d\, U_* \right], \tag{12}$$

where δ and β are empirical coefficients.

The **net radiation (Q*)** consists of the net shortwave radiation K* that originates from the sun and the net longwave radiation:

$$Q^* = K^* + L^+ - L^-, \tag{13}$$

where L^+ and L^- is the incoming and outgoing longwave radiation, respectively. The net shortwave radiation can be parameterized as:

$$K^* = (a_1 \sin\phi + a_2)(1 - b_1\, N^{b2})(1 - r), \tag{14}$$

where $a_1 \sin\phi + a_2$ is the incoming solar radiation with clear skies and $1 - b_1\, N^{b2}$ is the interception of solar radiation by clouds. The reduction factor $1 - r$ is due to the reflection of incoming solar radiation by the surface, where r is the surface albedo.

The net longwave is given as:

$$L^+ - L^- = L_i^* + 4\,\sigma T_r^3\,(T_r - T_o), \tag{15}$$

where σ is the Stefan-Bolzman constant and

$$L_i^* = -\,\sigma T_r^4\,(1 - c_1\, T_r^2) + c_2\, N,$$

is called the isothermal net longwave radiation (i.e. $T_r = T_o$).

The last term in (15) is a correction term due to the temperature difference that normally occur. Holtslag and van Ulden (1983) found:

$$4\sigma T_r^{\,3}\,(T_r - T_o) = -\,C_H \cdot Q^*, \tag{16}$$

where C_H is an empirical heating coefficient approximated by:

$$C_H = 0.38\left[\frac{(1-\alpha)\,S+1}{S+1}\right]. \tag{17}$$

During the night time T_r-T_o is strongly affected by the wind speed. In this case the surface layer similarity can be used to eliminate T_r-T_o:

$$T_r - T_0 = \frac{\theta_*}{k}\left[\ln\left(\frac{z_r}{z_H}\right) + 5\frac{z_r}{L}\right] - \Gamma_d z_r, \tag{18}$$

where $\Gamma_d = 0.01$ K m^{-1} is the dry adiabatic lapse rate and where the surface reference height for heat z_H is used instead of z_0 because near the surface the resistance for heat transfer differs from that for momentum transfer (Garatt and Hicks, 1973). For short grass, typically, $(1/k)\ln(z_r/z_H) = 30$ (van Ulden and Holtslag, 1983).

The **soil heat flux (G)** is parameterized as follows (van Ulden and Holtslag, 1985):

$$G = -A_G\,(T_r - T_o), \tag{19}$$

where A_G is an empirical coefficient for the soil heat transfer, typically 5 $Wm^{-2}\,K^{-1}$. For daytime this leads to:

$$G = C_G \cdot Q^*, \tag{20}$$

where

$$C_G = \left(\frac{A_G}{4\sigma T_r^{\,3}}\right) C_H. \tag{21}$$

For night time hours the temperature difference is eliminated similar to the net radiation.

The equation for **Q_* during daytime hours** is obtained from (9), (10), (12), (13), (15), (16) and (20). This gives the following equation:

$$\theta_* = -\frac{[(1-\alpha)\,S+1]\,(1-C_G)\,Q_i^*}{(S+1)\,(1+C_H)\,\rho\,C_p\,U_*} + \alpha\,\theta_d, \tag{22}$$

where

$Q_i^* = K^* + L_i^*$ is the isothermal net radiation,

and $\theta_d = \beta\lambda\Delta q_d/C_p$ is an empirical temperature scale.

During **night time hours,** θ_* is obtained from (9), (10), (12), (13), (15), (18) and (19), which gives the following equation:

$$\theta_* = T_r\left\{\left[(d_1 v_*^2 + d_2 v_*^3)^2 + d_3 v_*^2 + d_4 v_*^3\right]^{1/2} - d_1 v_*^2 - d_2 v_*^3\right\}, \qquad (23)$$

where

$$v_* = u_* / (5gz_r)^{1/2},$$
$$d_1 = \frac{1}{2k}\ln\frac{z_r}{z_h},$$
$$d_2 = \frac{1}{2}(1+S)\rho C_p (5gz_r)^{1/2} / (4\sigma T_r^3 + A_G),$$
$$d_3 = -Q_i^* / (4\sigma T_r^4 + A_G T_r) + \Gamma_d z_r / T_r,$$
$$d_4 = (1+S)\rho C_p (5gz_r)^{1/2}\theta_d / (4\sigma T_r^4 + A_G T_r),$$

where Γ_d is the dry adiabatic lapse rate 0.01 K m^{-1} and $\alpha = 1$ has been used. In these equations van Ulden and Holtslag suggested as typical values: $z_r = 50$ m, $d_1 = 15$, $A_G =$ 5 W m^{-2} K^{-1} and $\theta_d = 0.033$ K. With these values the coefficients d_2, d_3 and d_4 still depend on the reference temperature T_r while d_3 also depends on N and K*.

For practical applications the T_r dependence is neglected and the constants are approximated by their values for $T_r = 283$ K. Holtslag and de Bruin (1988) then obtain (using $z_r = 50$ m, (l/k) $\ln(z_r/z_H) = 30$, $\theta_d = 0.033$) the values $(5gz_r)^{1/2} = 50$ m s^{-1}, $d_1 = 15$, $d_2 = 6600$, $d_4 = 1.55$ and

$$d_3 = (-K^* + 96 - 60N)/2870,$$

where the dry-adiabatic correction term has been absorbed in Q_i^*. The present results agree with the mean value $\theta_* = 0.08$ found by Venkatram (1980) for predominantly clear sky conditions. The advantage of the present approach is that solutions for θ_* and u_* are also obtained for low wind speed. Thus, in principle, the present method also gives a practical solution for very stable conditions. That such practical solutions are useful has been shown by Holtslag (1984).

Another advantage of the present method is that no special provisions have to be made for transition hours between day and night.

3. TECHNICAL DESCRIPTION OF THE ALGORITHM

The metprocessor MEPDIM has a modular construction, i.e. each main option is carried out in separate subroutines. This section outlines the structure of the processor and describes the methods to estimate hourly meteorological data such as mixing heights, surface fluxes, dispersion parameters and solar radiation. The methods and subroutines in MEPDIM are presented in Table 1.

Table 1: Methods and subroutines in MEPDIM.

Parameter	Routine name	Description
CLON, CLAT, Z0, Z1, Z2	STATION	Description of the site and heights for observations
T2, T1, DD, FF	INTURB1	Input routine for the profile method
RNN, KIN, DD, FF	INTURB2	Input routine for the energy budget method using cloud cover (RNN)
QNET, T, DD, FF	INTURB3	Input routine for the energy budget method using net radiation
u_*, θ_*, L	TURB1	The profile method
u_*, θ_*, L	TURB2	The energy budget method, cloud cover
u_*, θ_*, L	TURB3	The energy budget method, net radiation
hmix	MIXHT	Mixing height
τ, HKIN, HSEN, HLAT, W_*	FLUX	Calculation of surface fluxes
σ_r, σ_W	DIFFPAR	Calculation of dispersion parameters
DD (±5°)	RANDOM	Random generator for wind direction if measured in decadegrees
SINPHI	SINSUN	Sinus to the solar elevation
QSTI	RADIAT	Calculation of isothermal net radiation based on time of the year and cloud cover
Functions:		
θ_*	TST2	Calculation of θ_* in TURB2
θ_*	TST3	Calculation of θ_* in TURB3
ψ_m	PSIM	Calculation of the wind profile
ψ_H	PSIH	Calculation of the wind profile
L	OBUK	Calculation of the Obuklov length
Statistics:		
	WINDDIR	Wind direction statistics
	MIXSORT	Statistics on mixing heights
	PRINT	Output of statistical results

STATION

This input routine requests description of the monitor site, such as the location of the station, the height of wind and temperature measurements and the surface roughness (z_o). The Davenport classes of surface roughness as given by Wieringa (1980) is given in Table 2.

Table 2: Terrain classification by Davenport (1960) and Wieringa (1980) in terms of effective surface roughness length z_0.

Class	Brief terrain description	z_0.(m)
1	Open sea, fetch at least 5 km	0.0002
2	Mud flats, snow; no vegetation, obstacles	0.005
3	Open flat terrain; grass, few isolated obstacles	0.03
4	Low crops; occasional large obstacles	0.10
5	High crops; scattered obstacles	0.25
6	Parkland, bushes; numerous obstacles	0.5
7	Regular large obstacle coverage (suburb, forest)	(1.0)
8	City center with high- and low-rise buildings	?-?

INTURB1

This subroutine has as input hourly meteorological data including temperature difference to be used in the profile method. This routine is used as input for TURB1. The routine also tests for valid data.

INTURB2

This subroutine has as input hourly meteorological data including cloud cover and optional the incoming shortwave radiation. This routine is used as input for TURB2. The routine also tests for valid data.

INTURB3

This subroutine has as input hourly meteorological data including net radiation. This routine is used as input for TURB3. The routine also tests for valid data.

TURB1

This routine containes the theory of the profile methods as described in section 2.1. This routine requests the temperature difference as input from INTURB1 and calculates the parameters u_*, θ_* and L.

TURB2

This routine containes the theory of the surface energy budget method as described in section 2.2. This routine requests cloud cover from INTURB2 and calculates the isothermal net radiation in RADIAT. The output of the routine is the basic parameters u_*, θ_* and L.

TURB3

This routine contains the same theory as TURB2, except that the input to describe the surface energy flux is measurements of the net radiation. The output of the routine is the same as for TURB1 and TURB2.

MIXHT

This subroutine calculates the height of the atmospheric boundary layer. This layer is generally regarded as the part of the atmosphere where the influence of surface friction and heating or cooling from the ground is observed. The height of the atmospheric boundary layer during neutral conditions is given by Deardorff (1971):

$$h = C_1 \cdot u_* / f, \tag{24}$$

where C_1 is an constant and f is the Coriolis parameter. The constant C_1 has a wide range in the literature: Delage (1974), guided by numerical modelling results, suggests the lower extreme of $C_1 = 0.05$. Deardorff (1971) uses the upper extreme of 0.30. Plate (1971) derived $C_1 = 0.185$, which is the value adopted by Benkley and Schulman (1979), as well as by several other modellers. In this metprocessor, an intermediate value $C_1 = 0.25$ is used, as recommended by Olesen et al. (1987). The use of (23) should be limited to atmospheric conditions that are sufficiently neutral. The following restriction can be used: $|U^*/(fL)|<4$ which corresponds to $|h/L|<1$.

For stable conditions, the formulae given by Zilitinkevich (1972) is used:

$$h = 0.4\,(\frac{u_* \cdot L}{f})^{1/2}. \tag{25}$$

To obtain a continuous calculation of the mixing height for L > 0, the selection of the two equations is varying dependent on the friction velocity.

FLUX

The fluxes of heat, momentum and evaporation can be obtained from observed profiles of wind and temperature using the similarity relations for the atmospheric surface layer. These relations are based on Monin-Obukhov similarity theory, which assumes stationary and horizontally homogeneous conditions. The flux of momentum or the surface shear stress is related to the friction velocity u_* by

$$\tau = \rho u_*^2, \tag{26}$$

where ρ is the density of air. This parameter is used as a scaling parameter in the unstable atmospheric boundary layer (ABL).

The flux of sensible heat H is related to u_* and the temperature scale θ_* by

$$H_{sen} = -\rho C_p\, u_* \theta_*, \tag{27}$$

where C_p is the specific heat at constant pressure.

The convective velocity scale is defined as:

$$w_* = (gH_{sen}\ H_{mix} / T)^{1/3}, \tag{28}$$

where g is the acceleration of gravity. This is the turbulent velocity scale in the unstable ABL.

The kinematic surface heat flux is related to the sensible heat flux by:

$$H_{kin} = H_{sen} / \rho_o C_p. \tag{29}$$

For the different scaling regions in the atmosphere, scales for velocity, temperature and length can be obtained from the above parameters to be used in dispersion modelling.

DIFFPAR

This routine calculates the standard deviation of velocity fluctuations in the horizontal (σ_v) and vertical (σ_w). The horizontal dispersion parameter σ_v is related to u_*, h_{mix} and L by:

$$(\sigma_v / u_*)^2 = 0.35\,(-\frac{h_{mix}}{kL})^{2/3} + (2 - z / h_{mix}) \tag{30}$$

for unstable conditions and $h_{mix}/L < 1$. This equation is based on an empirical model for shear-produced variance by Brost et al. (1982) and for buoyancy-produced variance described by Caughey (1982). For stable conditions is used:

$$(\sigma_v / u_*)^2 = 2(1 - z / h_{mix}) \tag{31}$$

as proposed by Gryning et al. (1987).

An estimate of σ_w for $z/L < 1$ can be obtained from

$$(\sigma_w / u_*)^2 = 1.5\,[z / (-kL)]^{2/3} \exp(-2z / h_{mix}) + (1.7 - z / h_{mix})\ for\ L < 0. \tag{32}$$

The equation is based on the theory by Brost (1982) and Bærentsen and Berkowics (1984).

For stable conditions, σ_w can be estimated by

$$(\sigma_w / u_*)^2 = 1.7(1 - z / h_{mix})^{3/2}, \tag{33}$$

as proposed by Nieuwstadt (1984).

RANDOM

This subroutine calculates a random number between -5 and +5 to be added to the old generation of wind data registered in decadegrees.

SINPHI

The solar elevation (ϕ) for a given time and location may be calculated by simplifying well-know astronomical formulas. The solar longitude (SL) in radians can be evaluated from

$$SL = 4.871 + 0.0175d + 0.033\sin(0.0175d), \tag{34}$$

where d is the Julian day. The solar declination (δ) follows from

$$\delta = \arcsin(0.398\sin(SL)). \tag{35}$$

Using the above estimates for SL and δ we can compute the hour angle (h) through which the earth must turn to bring the meridian of the given location directly under the sun:

$$h = \lambda_w + 0.043\sin(2SL) - 0.033\sin(0.0175d) + 0.262t - \pi, \tag{36}$$

where λ_w is the western longitude (rad) of the location and t is universal time in hours. The solar elevation (ϕ) follows from the above relations by applying (Sellers, 1965):

$$\sin\phi = \sin\delta\sin\psi + \cos\delta\cos\psi\cos h, \tag{37}$$

where ψ is the latitude of the location (rad). With the above scheme for the calculation of ϕ the accuracy of ϕ is within 0.05 rad, which is acceptable within the present context.

RADIAT

This routine calculates the isothermal net radiation Q_i* by using the radiation scheme by van Ulden and Holtslag (1985). Input parameters for estimation of Q_i* is the surface albedo (r), the cloud cover, sinus to the solar elevation and the incoming short wave radiation.

The albedo r describes the effect of the surface on the net incoming solar radiation, which is important in the surface radiation budget. The set of albedoes used in this metprocessor is given in Table 3. The albedo in Table 3 is given as the per cent of incoming radiation which is reflected by the ground.

Table 3: Selection of surface albedoes (%) used in this metprocessor.

Index	Surface	Albedo (%)
1	Dark soil	10
2	Forest	15
3	Grass, lush	20
4	Sand	25
5	Ice (old), sand (dark)	30
6	Ice (new)	35
7	Snow (old)	40
8	Snow (normal)	60
9	Snow (fresh)	80

The isothermal net radiation is given by:

$$Q_i^* = L_i^* + K_i^*, \qquad (38)$$

where

$$L_i^* = -\sigma T^4(1 - c_1 T^2) + c_2 N \qquad (39)$$

i the isothermal longwave radiation.

The isothermal shortwave radiation is given by:

$$K_i^* = K_{in}^*(1 - r), \qquad (40)$$

where

$$K_{in} = (a_1 \sin\phi + a_2)\cdot(1 - b_1 N^{b_2}) \qquad (41)$$

is the incoming shortwave radiation. This parameter can be given as direct input in the metprocessor.

TST2

This function calculates θ_* as a function of the friction velocity (u_*) and isothermal net radiation ($\theta_*{}^i$) according to van Ulden and Holtslag (1985) as described in Section 2.

TST3

This function calculates θ_* as a function of the friction velocity (u_*) and net radiation (Q_{net}) according to van Ulden and Holtslag (1985) as described in Section 2.

PSIM

This function calculates the stability correction function in the surface layer wind profile (ψ_m) as defined in equations (6) and (8).

PSIH

This function calculates the stability correction function in the surface layer temperature profile (ψ_H) as defined in equations (4) and (5).

OBUK

This function calculates the Obukhov length (L) from similarity theory. The Obukhov length reflects the height at which the contributions to the turbulent kinetic energy from buoyancy forces and from the shear stress are comparable (Obukhov, 1946).

L is related to u_* and Q_* by

$$L = \frac{T \cdot u_*^2}{k \cdot g \cdot Q_*}, \quad (42)$$

where k is the von Karman constant.

The profiles described in Section 2 includes corrections for very stable conditions. However, during stable conditions and low wind speed, the Obukhov length drops below 5 which leads to mixing heights below 10 m.

For dispersion modelling, NILU has included the critical Obukhov length as described by Holtslag and van Ulden (1982). For z/L > 1 the temperature scale is defined as:

$$\theta_* = 0.09\,(1 - 0.5 \cdot N^2), \quad (43)$$

which is in agreement with $\theta_* = 0.08$K obtained by Venkatram (1980).

By using standard the temperature profile ($\psi = \div\alpha\ z/L$), we can obtain a quadratic equation in L, whose solution can be written as:

$$L = (L_u - L_o) + \{L_u(L_n - 2L_o)\}^{1/2}, \quad (44)$$

where L_o and L_n are length scales given by

$$L_o = \frac{\alpha z}{L_n(\frac{z}{z_o})} \quad (45)$$

and

$$L_n = \frac{KU^2T}{2g\,\theta_*(L_n(\frac{z}{z_o})^2)}, \quad (46)$$

where $\alpha = 5$, k = 0.41, g =9.81. Real solution exists, however, for $L_n \geq 2\,L_o$ only. The lower limit $L_o = 12.8$ m for z = 10 m and $z_o = 0.2$. van Ulden and Holtslag suggested an extension of L below L_o for very stable conditions, but this method is not included in this metprocessor.

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ABSTRACT
Norwegian Institute of Air Research (NILU) has developed a **Me**teorological **P**reprosessor for **Di**spersion **M**odelling (MEPDIM). Based on tower measurements and the energy budget of the atmosphere, basic parameters for modelling atmospheric dispersion are calculated by parameterization of the structure of the atmospheric boundary layer. The theory is based on the methods of van Ulden and Holtslag.

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ABSTRACT (in Norwegian)
Norsk institutt for luftforskning (NILU) har utviklet en meteorologisk preprosessor til bruk for spredningsberegninger. Ved bruk av meteorologiske målinger beregnes spredningsparametre ved parameterisering av atmosfærens grensesjikt baser på teorien utviklet av van Ulden og Holtslag.

** Classification*	*A*	*Unclassified (can be ordered from NILU)*
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