ministerie van verkeer en waterstaat

koninklijk nederlands meteorologisch instituut

report of the workshop on boundary layer models in short range weather forecasting

including the abstract of all the presentations during the workshop.

. (De Bilt, The Netherlands, 10-12 March 1986)

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Contents

1. Introduction

During the years much research has been carried out to improve our understanding of a number of phenomena in the atmospheric boundary layer (ABL), e.g. the processes that govern the changes in the stable and convective ABL, the processes that govern the development of the clouds and fog in the ABL.

So far not much of the results of this work has been incorporated in forecast models. There exist a few one-dimensional ABL-models, and recently some Air Mass Transformation models have been developed, that combine a trajectory approach with a one-dimensional ABL-model. Also some fog models exist. The ABL parameterization in gridpoint numerical weather forecasting models concentrates mainly on the correct parameterization of the vertical fluxes of momentum, heat and water vapour in the ABL. In fact, the aim is to forecast the change in the dynamics of the atmosphere aloft. The development of the ABL itself is only poorly represented in these models. We had the opinion, that it was about time to organize a workshop on the topic: "Boundary Layers Models in Short Range Weather Forecasting". So a limited number of specialists in the field were brought together at the workshop, and finally there were 35 participants from 15 institutes. A list is given in Appendix A.

In this workshop we concentrated on the following questions:

- How far is the ABL-research at the moment, and what are the possibilities for forecasting the ABL-behaviour itself in an operational environment?
- If there are possibilities, what is the best way to proceed? Which models are promising? Which parameterization schemes are in principle able to give reliable forecast results?

In doing this, we focus ourselves on models and parameterization schemes that are able to forecast the following ABL-processes: The daily variation in temperature and ABL-height, the formation, development and dissipation of fractional clouds, layered clouds and fog in the ABL.

In the next section of this report we will describe the objectives of the workshop in more detail, while in section 3 the conclusions of the workshop will be given. The workshop program is given in Appendix B and the abstracts of the different contributions to this workshop are given in Appendix C.

2. Workshop objectives

Nowadays, there exist roughly two ways to describe the ABL in short-range weather forecasting models:

I The ABL-parameterization schemes in Limited Area Models (LAM's) and mesoscale models.

II One-dimensional operational ABL-models, which may be used at one geographical position for a short time period (often not more than 6 hours ahead), or which are combined with a trajectory approach, the so-called Air Mass Transformation models (AMT-models).

Table I gives the different characteristics for these models. In the workshop we want to find out which type of models can be applied for the forecasting of the different ABL-processes. In doing this, it may turn out, that for one ABL-process type I must be used, while for another ABL-process type II.

So far, a lot of operational experience is already obtained with LAM/Mesoscale models, therefore we concentrate in this workshop on the different types of 1-dimensional models and their possibilities. This is the subject of session I of the workshop: Different models and ABL-parameterization schemes will be treated, among them bulk parameterization schemes as developed by for instance Tennekes and Driedonks and applied by Randall in general circulation models and Reiff et al in an AMT-model. Also the K-model approach will be treated as it is applied in the Swedish approach of an AMT-model as developed by Karlsson. Further the partly non-local approach as developed and applied by Blackadar will be discussed and the non-local transilient approach as developed by Stull. Besides these techniques Burk will treat, the nesting of a high resolution ABL-formulation in a regional model and Golding will review the ABL-formulation in the British mesoscale model. Musson Genon will discuss the characteristics of a local dynamical interpretation model, which is currently under development. Later, in Session IIb, in discussing the parameterization of clouds and fog in the ABL, we will treat also several closure schemes.

During Session I we will concentrate on the following issues: Can the different models/schemes treat the different ABL-processes as the daily variation, stratiform and fractional cloudiness, and fog satisfactory, so that forecasting of these processes will be possible? This means for instance, are the schemes fast enough? In case that they are incorporated in a LAM or mesoscale model, can they run on a large main frame computer

Table 1

Different characteristics for the ABL-treatment in AMT-models and LAM-models for short-range weather forecasting

	AMT	LAM/Mesoscale models
Purpose	Local forecasting of ABL-processes	Regional forecasting of dynamical (and if possible ABL-) processes
Horizontal resolution	Variable (may in trajectory approach go down to a few kilometers/few hundred meters near the point of interest)	fixed 15-50 km
Vertical resolution	20-40 points; location of the points can be variable and may be changed according to the process which needs to be described, i.e. stratocumulus or fog, convective or stable ABL	fixed 5-10 layers up to 3 km
Runs/day	Whenever new information is available	2-4 times/day
Computer resources needed	Main frame or Personal Computer	Main frame
Limitations posed by the terrain	Works only in gently sloping terrain	Limited performance in mountainous terrain, depends on gridpoint resolution
Limitations posed by the synoptic situation	Does not work when large scale dynamics interacts with small scale ABL-turbulence i.e. land-sea breeze, nocturnal jet.	Does not forecast nocturnal jet. Forecast of other processes depend on gridpoint resolution

with 5-10 levels, 24-hours ahead, in approximately 15-30 minutes, even when they contain the evaporation, condensation and radiation schemes which are necessary to forecast these processes? Or, will an AMT-model approach be necessary with 20-40 levels in the vertical, giving a 24-hour forecast on a personal computer within a few minutes?

The next session, session II, is split into two parts: In session IIa the parameterization schemes for the surface fluxes will be reviewed. Questions to be addressed are: Are the present schemes accurate enough? How well must the vegetation layer be described to give reliable ABL-forecasts in LAM, mesoscale models and/or AMT-model approaches?

Session IIb is devoted to a few semi-operational/research models and techniques which describe fog and ABL-cloudiness. Questions to be addressed are: Is the present knowledge about these processes far enough developed already to try to forecast the evolution in time of these phenomena? If so, what are the most important physical processes which have to be parameterized to forecast these phenomena, and which processes are less important or can be parameterized by a climalogical value? What is the vertical resolution needed to forecast these processes in a reliable way? What input data are necessary and with what precision?

Session III, the last session, will concentrate more on the input data (session IIIa) and on the experience and possible applications of the different models (session IIIb). Questions to be addressed: Are the requested input data at present available or if not, is it possible to obtain them in the near future? Special attention will be given in this session to the developments in the retrieval of satellite soundings. And finally, are AMT-model or other 1-dimensional ABL-model approaches necessary to forecast the ABL-processes or will the development of computerpower make such models superfluous and will the ABL-processes in the near future, be forecasted in LAM models and in mesoscale models? To find an answer on these questions we listened to 21 presentations and had many discussions. The result of all this is presented below.

3. Conclusions

In this section we summarize the discussions and try to formulate some conclusions from them. The conclusions may therefore largely be seen as the consensus which was reached during the discussions.

During the workshop it became more and more clear, that the most important problems in the ABL, especially in relation to forecasting problems, are in the treatment of clouds in the ABL. In the clear ABL there are also some interesting problems, especially in the stable ABL over land; however, from the point of view of a forecaster, the present ABL-formulations give reasonable results on the daily variation of the temperature in the ABL and the ABL-height during clear sky conditions. However, problems arise when case the sky is cloudy. Partly these problems are due to an unsuccesful forecast of the cloudiness in the atmosphere aloft (this issue is not addressed during this workshop), partly the problems arise due to the development, maintenance or dissipation of fractional cloudiness, layered clouds and/or fog in the ABL.

3.1 The status of understanding the different cloud processes in the ABL

As a logical consequence of the above statements we first summarize the "status" of understanding the different clouds processes in the ABL. We realize, that much research, field work and theoretical research, is still needed to reveal the many aspects of the phenomena we want to forecast. However, in this workshop we concentrate more on what is reached already than to give a list of what is still missing.

3.1.1. The status of understanding and forecasting of fog

Fog can be split into two types of fog, the not fully developed radiation or ground fog, characterized by the fact that the sky (or clouds) above the fog still can be seen from the ground, and the mature fog in which this is not the case. For the mature fog, the important physical processes are known (see abstracts II-6, II-8 an III-4). The importance of gravitational droplet settling for the dissipation of fog needs still a lot of attention. Karlsson (III-4) showed, that it is possible to forecast mature fog rather succesfully in an AMT-approach, however, more objective verification is needed. Blaauboer showed that KNMI's AMT-model also had some success in forecasting fog (III-3). Wessels (II-8) showed, that it is possible to use a bulk-formulation to forecast fog, however, so far only one case was treated. The same holds for Musson Genon (II-6) who reproduced the evolution of a fog-case in a higher order closure 1-dimensional ABL-model, to be incorporated in a large scale model. Burk (II-9) showed the results of a comparative study between different semi-operational models, that tried to forecast fog above sea: all models have some failures, but the best performance was by the ones with a higher order closure scheme.

Fog is not treated well in LAM's and mesoscale models. In general it was concluded that the understanding of mature fog is such, that there is enough room to develop short range weather forecast models/techniques for it, while this is not the case for ground fog.

3.1.2 The status of understanding and forecasting of stratus/stratocumulus

A lot of work still has to be done on the understanding of the processes that drive stratus and stratocumulus, especially in the case when windshear at the top of the layered clouds is an important feature in the generation of turbulence. However, Duynkerke showed (II-7) that with an 1.5 order closure model the time evolution of some stratocumulus cases can be reproduced. Karlsson (III-4) gave a good example of a forecast of the evolution of a stratus-case in his model. In forecasting 143 cases, the model had 57% success, 23% moderate success and failed in 19% of the cases. These qualifications were given subjectively, but by independent forecasters. In LAM's and mesoscale models, layered clouds are so far not forecasted in a physical way.

It seems, that for the stratus/stratocumulus cases in which the radiation is the main driving force of the (upside down) turbulence at the top of the cloud layer (about 60% of the cases), the evolution of the stratus/stratocumulus can be reproduced, and that there is enough perspective to go on in trying to forecast those.

3.1.3 The status of understanding and forecasting of fractional cloudiness

The members of the workshop realised that lately a lot of work has been done in the improvement of the shallow and deep convection schemes in numerical grid point forecast models. We did not discuss these improvements at this workshop. Stull (II-5), however, showed some interesting results from a field experiment on the onset of cumulus convection and resulting fractional cloudiness. A statistical distribution of the height of the ABL combined with a statistical distribution of the lifting condensation level (LCL) made it possible to reproduce the onset and time evolution of fractional cloudiness much better than in cases in which this was neglected. It seems promising to use these statistical distributions rather than fixed values in schemes, which describe the evolution of shallow convection.

Much research work, however, has to be done in the field of transition of stratocumulus to cumulus.

3.2. Needs for reliable ABL-cloud and fog forecasts

The next question which we addressed in our discussions was: What are the needs for reliable ABL-clouds and fog forecasts? In these needs, we neglect the necessary timesteps in a model, as these are dependent on numerical schemes and difficult to guess.

3.2.1. Mature fog

The input data needed for a reliable short-range forecast of mature fog are the following: Detailed information about the underlying surface, temperature and humidity profiles in the ABL, the geostrophic wind, the existance of upper-level clouds, and, in case mature fog exists already, the thickness of the fog layer.

The vertical resolution to forecast mature fog is probably something like 10 layers in the lowest 100 meters, unless a bulk-formulation can be used succesfully. The necessary horizontal resolution will be determined by the changes in the terrain, above sea this may be 10-20 km, above land this may go down to 1-2 km.

The processes which have to be treated carefully are surface processes, radiation, the thermodynamic consequences of vertical mixing and further the gravitational droplet settling and dew formation.

3.2.2. Fractional cloudiness

The input data are the same as those in 3.2.1., but added to that knowledge of the temperature- and humidity profile above the ABL is necessary plus information about the large scale vertical velocity. The information about the underlying surface may be somewhat more crude than the information which is necessary to forecast mature fog.

The necessary vertical resolution near cloud base must be several layers of most 50-100 meter thick. However, if the height of the ABL and the width of the entrainment layer, and the height and width of the LCL are forecasted by an additional equation, no extra vertical resolution is needed.

The required horizontal resolution depends on the changes in terrain, roughness and surface fluxes (land/sea interface, lakes, sandy area's between wet cultivated area's etc). Often, the downstream distance from such transitions plays a role in the forecast problem.

3.2.3. Stratus and Stratocumulus

The necessary input data are the same as those in 3.2.2. If it is present, the thickness of the cloud layer is also required. How much information is necessary about the windshear at the top of the ABL (or cloud layer, if it is present) is not known yet. Knowledge of the underlying surface is only in a crude way necessary.

The required vertical resolution is about 25 meter over several hundreds of meters near the cloud layer. If bulk parameterizations will be possible, this requirement does not exist. The horizontal resolution required for forecasting layered clouds is normally not a limiting factor in the model: often the horizontal extension of layered clouds is hundred up to several hundreds of kilometers.

3.3. The possibilities to forecast ABL-clouds and fog

So far we discussed the needs for forecasting ABL-clouds, the next question is: Is it possible to fulfill these needs? And, related to that question, what types of ABL-models/schemes are in principle able to forecast them, given the computer resources and required data input?

3.3.1. Mature fog

As the vertical and horizontal resolution necessary to forecast mature fog is out of the question for LAM's and mesoscale models, a physical way of forecasting fog in these models will in the foreseeable future hardly be possible. However, this may not be the case when a reliable bulk parameterization scheme for the forecasting of mature fog is developed (See Wessels, II-8). Also the nesting of a detailed ABL-approach at certain parts of the regional forecasting domain that are sensitive to fog formation may circumvent this problem (See Burk, I-3). Further, above open sea the situation may be better as the required horizontal resolution is often not that fine. Finally, above land where the characteristics of the underlying surface often change over a distance of several kilometers, a statistical approach, in which the forecasted LAM parameters and detailed information about the underlying surface is incorporated, may lead to more or less reliable results.

The one-dimensional AMT-approach is very well suited as a physical way of forecasting fog, because a physical way of forecasting fog at a specific place is only possible with a model which contains the correct advection processes. Moreover, the AMT-approach has the possibility of variable horizontal and vertical resolution. The humidity for instance, is a sensitive parameter for forecasting fog and advection plays a dominant rôle in the evolution of the specific humidity at a certain place. Results of AMT-model approaches for forecasting fog showed encouraging results (Blaauboer III-3, Karlsson III-4).

3.3.2 Fractional cloudiness

ABL cumulus clouds can, in principle, be forecasted in all type of models. When the downstream distance of a transition in the underlying surface plays a role, the AMT-approach is best suited to treat this type of processes. The downstream distance at which ABL-cumulus clouds appear may depend on the temperature and humidity profile, the change in surface fluxes and the advection velocity. The downstream distance can vary from a few hundred meters till 50-100 km. The schemes to forecast fractional ABL-clouds can still be improved and a lot of research still has to be done.

3.3.3. Stratus/stratocumulus

To develop a scheme, that gives reliable forecasts of the time evolution of layered ABL-clouds is a difficult task to fulfill. Unless a reliable bulk formulation is developed, a high vertical resolution is needed. Besides that, a sofisticated parameterization scheme is needed of the radiation processes. Also the thickness of the layered clouds, if they are present, is necessary as input quantity. This thickness will not be measured easily. Forecasting layered clouds in the ABL will hardly be possible with LAM's or mesoscale models, as advection of stratus and stratocumulus plays often an important role in the forecasting of these clouds. The AMT-model is probably the best candidate to forecast these phenomena. Nesting of ABL-layers in a limited domain of a LAM model, or a passive 1-dimensional ABL approach in a large scale model, will often not solve the problem.

3.4. Conclusions concerning the input date and the analysis

3.4.1. Fog

The thickness of the fog layer is an important input value. It may be worthwile, to incorporate a guessed value of this parameter into the synoptic weather reports. In the near future information about the locations where fog exists at a specific time may be drawn from satellite pictures, even at night.

3.4.2. Stratus/Stratocumulus

For stratus/stratocumulus information about the thickness of the layered clouds is important. There may be several ways to obtain this information:

(i) Near regularly used airports, from aeroplanes flying through the layered cloud, while they ascend or decend. Procedures have to be developed, so that this information reaches the weather forecasting offices.

(ii) From satellite information the cloud top temperature may be obtained and so, with an uncertainly of 100-200 meter, the height of the cloud top. From the synoptic reports the height of the cloud base may be obtained. From a combination of these methods, the thickness can be derived. This method, however, will probably not be precise enough.

(iii) Wessels proposed a radiation method. Given the possible global radiation at a certain moment (from latitude, time of the day and time of the year) and a measurement of the global radiation at the surface, a relation may be derived between the percentage of reduction in the global radiation and the thickness of the layered clouds. Satellite pictures must then reveal the existance of higher clouds above the layer clouds, which may spoil the relation. This method looks worthwile enough to investigate. Climatological values of the turbidity at a specific location are also required. Above sea, however, it will not be easy to obtain information from such a method.

3.4.3. General

Information about the vertical structure of the ABL and the layers above the ABL can be obtained from radiosoundings. However, additional information of sodars and lidars may help a lot. It may be worthwile to set up networks of these equipment and disseminate the measurements as part of the synoptic weather reports.

Above sea, however, as the weather ships are going to be withdrawn, information about the ABL will pose a problem. Juvanon du Vachat (III-1) showed us, that the quality of TOVS in the foreseeable future will not improve so much, that reliable ABL-temperatures and humidities can be obtained from them. Some information near the surface of for instance stability and humidity may in the future become available from surface-buoys which some countries are going to employ. Information of the temperature- and humidity-profiles above the surface will at sea, however, be difficult to obtain.

Another problem is the analysis of ABL-parameters. Cats (III-2) made clear that, with information about the terrain and the synoptic weather reports, it is possible to make analysis of stability. With such an analysis and an analysis of the surface windfield, the windfield at a certain height in the ABL can be constructed. He also showed a method to make 3-hourly gridpoint guesses for the ABL-heights from radiosonde information plus the use of the AMT-model. Extra input from sodar and lidars, however, would be very welcome to refine such a method.

3.5. Conclusions concerning the surface fluxes

Blackadar (II-1) gave a general introduction into this subject, while De Bruin (II-2), Holtslag (II-3) and Karlsson (II-4) treated the practical side and operational experiences with the different approaches.
It turns out, that complicated models are needed as a kind of "truth" to check simple parameterization schemes. Often simple models can be used, for the particular problems addressed in this meeting.

Simple parameterization schemes for the surface fluxes from the following surfaces are available: sea, bare soil and grass. But, still a lot of attention is needed for schemes, which treat forests, cultivated area's, and snow/ice coverages. Holtslag (II-3) compared different parameterizations with observations, and concluded that especially the estimation of solar radiation from simple meteorological quantities is quite uncertain and needs improvement. Karlsson (II-4) found, that especially the albedo treatment of old/new snow and ice covered by thin snow layers, needs further attention. De Bruin (II-2) made clear, that for cultivated area's the Penman-Monteith approach and the Priestley-Taylor approach probably will be satisfactory. However local/regional information on the resistance coëfficients and the surface moisture availability is still needed here. It is hoped that this will become available from satellite information in the next 5-10 years.

3.6. The different approaches in the treatment of turbulence

The requirements of forecasting fog, fractional cloudiness and layered ABL-clouds leads to the need of flexible numerical schemes for forecasting the turbulent processes in the ABL. To save computer time it is worthwhile to develop bulk parameterization schemes. However, bulk schemes are only applicable when the circumstances make a clear distinction possible between the different catogories of ABL's: mixed ABL, stable ABL, fog, stratus, stratocumulus, fractional cloudiness and so on. Often nature developes profiles in between these well-defined categories and than a bulk formulation leads into difficulties. Example of these "in between situations" are: the ABL near sunrise and sunset, the ABL with weak externally forced circumstances like in winter and above sea.

As it is worthwile to give some characteristics of the different turbulence formulations, we present table II.

3.7. The different "1-dimensional" approaches

When we compare "the nesting of a high resolution ABL-model at a limited domain in a LAM" with "a passive nesting of a 1-dimensional ABL-model at one (or several) places in a LAM", than the first approach has the advantage that at least in that limited domain, (i) the advection processes are better described, (ii) the treatment of the interaction between large scale dynamics and small scale turbulence is, in principle, possible.

Comparing the nested ABL-model approach over a domain with the AMT-approach gives the following characteristics:

The nested approach has as an advantage that it can describe over the nested domain, in principle, processes like land/sea breezes and mountain/valley circulations. Further within the nested approach one does not have to calculate all the time new trajectories.

The AMT-approach has as an advantage, that the advection over larger distances (outside the limited domain) is described, which is often important for stratocumulus forecasts and sometimes in the forecast of fractional cloudiness.

Table II

Different characteristics of the different approaches for the treatment of turbulence in the ABL

In column A characteristics of bulk models are given, in column B of K-models, in column C of transilient turbulence theory models and in column D of higher order closure models. In the table + (-) means that the listed characteristic is (not) of application, while o means that the characteristic is of application when additional requirements are fulfilled.

CHARACTERISTIC	A	B	C	D
Order of Parameterization	0.5	1	1	1.5-3.0
Local Closure	+	+	-	+
Calculation of Mean Profiles	+	+	+	+
Calculation of Flux Profiles	-	+	+	+
Calculation of Turbulent Kinetic Energy	-	-	-	+
Applicable for Ideal Mixed Layers	+	-	+	+
Applicable for Ideal Nocturnal BL	+	+	+	+
Applicable for Real (Non-Ideal) BL	-	o	+	+
Applicable throughout the whole atmosphere	-	+	+	+
Relatively fast on computer*	++	+	o	-
Numerically stable	+	o	+	o
Can include Clouds and Radiation	+	+	+	+
Operationally tested	+	+	-	+

* Depends if a vector machine is used or not. This guess is based on Personal Computer experience.

An other advantage of the AMT approach is the possibility to vary the horizontal resolution and vertical resolution: At the down stream side of changes in terrain the horizontal resolution can be increased if that is necessary. The same holds for the vertical resolution when changes from a convective situation to a stable situation occur, or in case when fog or stratocumulus starts to appear.

3.8. Final remarks

The workshop could not address all the problems that are important in the forecasting of ABL-processes. We concentrated on ABL-clouds, especially on fog and stratocumulus. Further, by reading the conclusions one must take in mind, that we often were talking about the best ways to proceed "in the foreseeable future". We hope, however, to have set some lines for future research and applications.

Acknowledgement

We thank all participants for their contribution to the discussions. Sandra Klutz is thanked for her help in organizing the workshop and typing this report.

APPENDIX A

List of participants

a. From other institutions than KNMI:

- A.K. Blackadar	Pennsylvania State University, U.S.A.
- H.A.R. de Bruin L. Kroon	Agricultural University of Wageningen, The Netherlands.
- J. Businger	National Center for Atmospheric Research, Boulder, U.S.A.
- L. v.d. Berg	Institute of Meteorology and Oceanography, Utrecht, The Netherlands.
- S.D. Burk	Naval Environmental Prediction Research Facility, Monterey, U.S.A.
- D. v.d. Duyn Schouten	Royal Air Force, The Netherlands.
- K. Ericsson G. Omstedt J. Melgarejo	Swedish Meteorological and Hydrological Institute, Norrkoping, Sweden.
- B.B. Hicks	NOAA Air Resources Atmospheric Turbulence and Diffusion Division, Oak Ridge, U.S.A.
- B. Golding	Meteorological Office, Bracknell, England.
- R. Juvanon du Vachat L. Musson Genon	Meteorologie Nationale, Paris, France.
- E. Karlsson,	National Defence Research Institute, Umea, Sweden.
- J. Nemeghaire	Royal Meteorological Institute of Belgium, Brussels, Belgium.
- R.B. Stull	University of Wisconsin, Madison, U.S.A.
- N. Woetmann Nielsen B. Hansen Sass	HIRLAM project, Copenhagen, Denmark.

b. From KNMI :

A.P.M. Baede
D. Blaauboer
G. Cats
A.G.M. Driedonks
P.G. Duynkerke
L.M. Hafkenscheid
A.A.M. Holtslag
F.T.M. Nieuwstadt
W. Remmers
J. Reiff
T.L. van Stijn
H. Tennekes
S.A. Tjemkes
A.P. van Ulden
H.R.A. Wessels
J. Wieringa

APPENDIX B

Workshop Program

Monday 10 March :	9.00 - 9.30	Coffee
	9.30 - 9.40	Opening (Dr. H.M. Fijnaut, Director-in-Chief of KNMI)
	9.40 - 10.00	Workshop Ojectives (Hans Reiff)
SESSION I:	Operational Models with Boundary Layer Formulation. Chairman: Ad Driedonks	
I-1	10.00 - 10.30	The Air Mass Transformation (AMT-) model at KNMI (Hans Reiff)
	10.30 - 10.45	Discussion
	10.45 - 11.15	Coffee Break
I-2	11.15 - 11.45	A Numerical Boundary Layer Model in Operational Use (Edvard Karlsson)
	11.45 - 12.00	Discussion
I-3	12.00 - 12.30	Nesting of a High Resolution Planetary Boundary Layer Formulation in a Regional Forecast Model (Steve Burk)
	12.30 - 12.45	Discussion
	12.45 - 13.45	Lunch
I-4	13.45 - 14.10	The Penn. State/NCAR model - Treatment of the Boundary Layer (Al Blackadar)
	14.10 - 14.20	Discussion
I-5	14.20 - 14.40	A local Dynamical Interpretation Model (Luc Musson Genon)
	14.40 - 14.50	Discussion
	14.50 - 15.00	Tea Break
I-6	15.00 - 15.25	The English Meteorological Office Mesoscale Model: Its Current Status (Brian Golding)
	15.25 - 15.35	Discussion
I-7	15.35 - 16.00	A Transilient Turbulence Model of the Boundary Layer (Roland Stull)
	16.00 - 16.10	Discussion
	16.10 - 17.00	Session discussion and Summary
	17.00 - 18.00	Reception at room 390

Tuesday 11 March:

SESSION IIa :	Surface fluxes and the Treatment of Vegetation. Chairman : Aad van Ulden		
II-1	9.00 - 9.30	Surface Fluxes and the Treatment of Vegetation : General Introduction (Al Blackadar)	
	9.30 - 9.40	Discussion	
	9.40 - 10.00	Coffee Break	
II-2	10.00 - 10.30	Surface Fluxes and the Treatment of Vegetation: Practical Approaches (Henk de Bruin)	
	10.30 - 10.40	Discussion	
II-3	10.40 - 11.00	Surface Fluxes in the AMT-model at KNMI (Bert Holtslag)	
	11.00 - 11.10	Discussion	
II-4	11.10 - 11.30	Treatment of the Surface Fluxes and Vegetation in the Swedish Numerical Boundary Layer Model (Edvard Karlsson)	
	11.30 - 12.00	Session Discussion and Summary	
	12.00 - 13.00	Lunch	
SESSION IIb :	Parameterization of Clouds and Fog Chairman : Ad Driedonks		
	13.00 - 13.15	Introduction (Ad Driedonks)	
II-5	13.15 - 13.45	A Phenomenological/Statistical Approach to Fractional Cloudiness Parameterization (Roland Stull)	
II-6	14.00 - 14.30	A Fog Study with a Local Dynamical Interpretation Model (Luc Musson Genon)	
	14.30 - 14.45	Discussion	
	14.45 - 15.00	Tea Break	
II-7	15.00 - 15.20	Modelling Stratocumulus Fields (Peter Duynkerke)	
	15.20 - 15.30	Discussion	
II-8	15.30 - 16.00	Characteristic Physical Properties of Fog: Requirements for Models (Herman Wessels)	
	16.00 - 16.15	Discussion	
II-9	16.15 - 16.45	Fog Forecasts in Operational Models: A Comparative Study (Steve Burk)	
	16.45 - 17.00	Discussion	
	17.00 - 18.00	Session discussion and Summary	
	19.30	Diner at Restaurant in Utrecht	

Wednesday 12 March:

SESSION IIIa :	Retrieval and Analysis of Meteorological Data for Operational Boundary Layer Model Applications Chairman : Hans Reiff		
III-1	9.00 - 9.45	On the use of High-Resolution Satellite Data for Mesoscale Analysis (Régis Juvanon du Vachat)	
	9.45 - 10.00	Discussion	
	10.00 - 10.30	Coffee Break	
III-2	10.30 - 11.00	Analysis of Boundary Layer Parameters (Gerard Cats)	
	11.00 - 11.10	Discussion	
	11.10 - 11.30	Session Discussion and Summary	
SESSION IIIb :	Applications Chairman : Hans Reiff		
	11.30 - 11.45	Introductory Remarks on Applications (Hans Reiff)	
III-3	11.45 - 12.15	Operational Experience with the AMT-model at KNMI (Dick Blaauboer)	
	12.15 - 12.30	Discussion	
	12.30 - 13.30	Lunch	
III-4	13.30 - 14.00	Operational Experience with the Swedish Numerical Boundary Layer Model (Edvard Karlsson)	
	14.00 - 14.15	Discussion	
III-5	14.15 - 14.45	Operational Experience with the U.S. Navy Models (Steve Burk)	
	14.45 - 15.00	Discussion	
	15.00 - 15.30	Tea Break	
SESSION IIIc :	15.30 - 16.30	Summary and Conclusions of the Workshop : A Round Table Discussion Chairman : Hans Reiff	
	16.30	Closing of the Seminar	

APPENDIX C

This appendix contains the abstracts of all the presentations during the workshop.

The number on each abstract refers to the number in the program of appendix B.

The Air Mass Transformation (AMT-) Model at KNMI

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1. Introduction

The purpose of the AMT-model is to give local forecasts for the period 6-24 hours ahead. We thereby concentrate on the boundary-layer. For practical reasons at the moment a forecast period of 12-24 hours ahead is used. Besides that, also 36-hours forecasts are made to get an impression how fast the skill of the forecast decreases with the length of the forecast period.

2. The AMT-model concept

The model consists of a trajectory part (Reiff et al., 1984) and an 1-dimensional ABL-part. Backward trajectories are calculated from our place of interest. For this, all the three components, of the winds forecasted by the European Center's model are used. This defines a source area. In the source area data from neighbouring radiosoundings are used to define initial temperature- and humidity profiles. The 1-dimensional ABL-model is used to calculate the changes in these profiles along the trajectories towards our place of interest. The model keeps track of the exchanges of heat and water vapour between the ABL and the underlying surface (land and sea, see also abstract II-3 of Holtslag) and the entrainment at the top of the boundary layer. Surface fluxes are calculated along the lowest trajectory. The trajectories ending at our place of interest higher up in the atmosphere, which may come from a different direction, are meant to forecast the part of the profile above the ABL correct.

3. The ABL-part of the model

For the boundary layer part of the model bulk formulations are used. During unstable conditions the mixed layer approach developped by Tennekes (1973) and Driedonks (1982) is used, for the stable boundary layer the approach developped by Nieuwstadt et al. (1981). During unstable conditions the models uses well-mixed potential temperature and humidity profiles in the ABL capped by an inversion, while during stable conditions the potential temperature is not-well mixed, but assumed to be linear. The boundary layer height during stable conditions is governed by a rate-equation and tends to go to an equilibrium value.

So far the model is dry and does not contain radiation divergence processes in the atmosphere. This means that the model is able to describe the development of the ABL up to the point where processes as fog and stratocumulus occur. In the present version, the model is not able to describe the evolution of ABL-cloud and fog processes.

The model does not have a fixed vertical spacing. It keeps track of all the significant points that define the initial temperature and humidity profiles. It is able to create or dissolve points in the vertical, and is therefore able to put a detailed vertical resolution at that height in the atmosphere there where it is needed.

4. Results

The model has been tested with data from 1981-1982. Analysed windfields and observed cloud amounts were used in this test. It turns out, that the model performs best during spring, summer and autumm. During these seasons the mean absolute error (mae) in the hindcast for the air temperature at 1.5 m (T) at noon is 1.5°C, while the mae for the absolute humidity in the ABL (Q) at noon is 10 to 15% off the observed q. The correlation coefficient between observed T and model T were for the three seasons between 0.8 and 0.95, for q they were between 0.8 and 0.9. As the ABL during the winter is more shallow, its characteristics are more difficult to forecast (v.d. Berg et al., 1985). By dividing our material into subsets we found that during convective circumstances the model hindcasts were at best when the sky was partly cloudy or clear ($N \leq 6/8$), while during the nights the model worked best in slightly stable circumstances ($N > 6/8$).

To test the model during changes from land to sea use was made of data from two mesoscale experiments that were held during 1981 in the German bight. Results showed a satisfactory performance. (Driedonks et al. 1985). Since summer 1984 the model has been tested in operational circumstances, since spring 1985 it is in operational use in KNMI's weather forecasting office. Results will be shown by Blaauboer, see abstract III-3.

5. Conclusions

From the herefore described experiments and some sensitivity tests the following conclusions may be drawn:

1. The model works well for the intended situations so in gentle sloping terrain in situations with advection.
2. In the Netherlands the advection process has to be included when forecasts for 12 hours or more are made. It contributed significantly to the results. The source area is than often more than 500 km away from our point of interest.
3. To obtain reliable results a knowledge of the ABL characteristics in the source area is necessary.
4. At least in one case (advection over the Southern Chinese Sea during a cold surge in Winter MONEX) it is shown, that different advection of air in the ABL and above it was necessary to obtain good results.
5. An advantage of the above described concept is that the model puts its vertical resolution there where it is needed. Especially when fog and stratocumulus are going to be incorporated in the model this is necessary.

   A disadvantage is, that the concept cannot be used when large scale dynamics and ABL-processes interact, for instance during land/sea breezes or mountain valley circulations. Another disadvantage is, that local forecasts are made by one model run only for one moment. When forecasts are necessary some hours later, new trajectories have to be calculated.
6. The advantage of the 1-dimensional bulk model used in the ABL parameterization is its simplicity. In practice, we found out that there are many situations in between "clear convective circumstances and "stable situations", for instance during the transition hours from day and night, above sea, or when the ABL is not in equilibrium with the underlying surface (transition sea - land, outside of lakes).

   An improved bulk formulation has to be developed or multilevel formulations have to be used in these situations.

6. Future developments

- At the moment the model is also available on an IBM-PC / AT, so that forecasters in remote areas (with access to forecasted windmaps and radiosoundings) can make there own forecasts, in an interactive way.

It turns out that the present, dry, version of the model uses only 30 seconds to make a 12 hour forecast (without input or output).

- The surface flux parameterizations, especially those during the night, will be improved. We use independent data sets for this, tested at moderate latitudes under as many circumstances as possible. To stress the interactive use, we will make it possible to the forecaster to overrule the default values for some parameters (see also the abstract of Holtslag, II-3).
- A "second-generation" AMT-model will be developped in near future, that will contain condensation, evaporation and radiative divergence processes in the atmosphere.

7. References

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A Numerical Boundary Layer Model in Operational Use

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An numerical boundary layer model is now in operational use at the two Air Force Bases F21 in Lulea and F10 in Ängelholm. At F21 the model has been subject to tests for 2 years before the operational start. At F10 the use of the model started in autum 1985.

Trajectory

The horizontal advection in the model is treated by moving a column of air along a mean trajectory in the atmospheric boundary layer. The length of the trajectory may vary between 20 to 500 km depending on situation and wind conditions. The forecast length can vary from some hours to 12 hours.

One-dimensional boundary layer model

During the movement along this trajectory a numerical one-dimensional boundary layer model of K-type is applied to the air column. The air column is moved by the wind velocity computed by the one-dimensional boundary layer model at the height of maximum turbulence. The dependent variables of the model are two horizontal wind velocity components, temperature in air, temperature in soil and total specific humidity. Liquid water content is defined as the difference between the total specific humidity and the saturation value.

Along the trajectory new horizontal grid points are defined when the surface characteristics are changed. These new boundary conditions are applied to the one-dimensional boundary layer model when the air column reaches the grid points. The different conditions that has to be defined at each horizontal grid point are:

Surface albedo
Vegetation height

Heat capacity of vegetation

Heat capacity and heat conductivity in soil

Water surface temperature (for grid point at sea)

Evaporation properties of dry vegetation

Amount of water intercepted on wet vegetation

Gradient wind velocity

Higher cloud albedo

Terrain height

Latitude and longitude

Distance from starting point

The vertical grid of the one-dimensional boundary layer model consits of 36 grid points of which 5 are below the surface. In addition to these 36 grid points there are 8 grid points at higher levels for the calculation of long wave radiation. The time step of the model is 60 seconds.

Effect of vegetation

The effect of vegetation on the winds is modelled by conventional methods, i.e. by defining a displacement level and a roughness height, both as a function of the vegetation height.

The effect of vegetation on the temperature is treated by solving the thermodynamic prognostic equation inside the canopy, with the assumption that the turbulent exchange coefficient for heat is assumed to be equal to the value just above the displacement level. At the surface, below a possible vegetation cover, the thermodynamic equation is solved for a thin layer of soil , air and vegetation, using a mean heat capacity for the soil and the canopy and using heat fluxes to and from this thin layer.

Over water the surface temperature is prescribed and the thermodynamic equation is solved down to the surface. When starting to test the model it was obvious that this boundary condition needed a very high resolution close to the surface, especially in instable conditions. To get correct fluxes and satisfactory prediction results we had to put the first grid point above the surface at 0.001 m.

Contrary to the heat flux in the thermodynamic equation, the humidity boundary condition is introduced at the displacement level as a flux balance starting from the formulation

$$E = Ha \times Ep,$$

where Ha is the so called Halstead parameter, E is the actual humidity flux and Ep is the humidity flux over a wet surface.

As E and Ep can be expressed by flux equations we can get a relation between the specific humidity, q(d), at the displacement level and the specific humidity, q(d + 1), at the first grid point above the displacement level

$$q(d) = Ha\ q_s(d) \quad + \quad (1-Ha)\ q(d+1).$$

If the temperature is below zero, q_s is the ice saturation value. As this equation is a "flux balance" the accuration in the solution of the humidity equation will be dependent on the mesh width at the displacement level.

By calculating Ha in the case of evaporation from dry vegetation, we first tried to start from Monteith's evaporation equation. However that method has not been successful because we got to large evaporation, eg computed dew points were to high and clearing of fog and stratus was to slow. I am not sure of the reason of the failure, but it could be coupled to the grid mesh at displacement level where the boundary condition is formulated. Other reasons may be the formulation of the aerodynamic resistances, or the used values for the surface resistance of the vegetation.

Instead of Monteiths equation we now are using a constant value for Ha = 0.05 for dry vegetation in all situations. This means that dry vegetation is assumed to evaporate 5% of the evaporation that should occure from a wet surface.

In the case of wet vegetation we use intercepted water on the vegetation following an equation according to Bringfeldt

$$Ha = C/S,$$

where C is amount of water intercepted on the vegetation and S is maximum amount of water intercepted on the vegetation. If we have vegetation and the temperature is falling giving dew, Ha is put equal to 1.0 which means that the surface is wet. If the temperature is changed to raising, Ha is changed to the evaporation formulation for dry or wet vegetation.

At water, snow or ice surfaces the humidity is assumed to be saturated (Ha = 1.0). In the case of snow covered forests the humidity is also assumed to be saturated at the dispacement level.

The latent heat flux from vegetation or other surfaces is introduced into the thermodynamic equation at the displacement level using the flux of moisture to or from the surface.

Since the thermodynamic equation is solved inside the vegetation cover it is natural to distribute the radiation fluxes linearily between the surface and the displacement level. The flux divergence is then introduced into the thermodynamic equation giving a distributed heating or cooling inside the vegetation. Since the sun radiation has very large influence on the boundary layer, the surface albedos are very important. The values we are using are based on different measurement investigations eg Pertu. However there are still uncertainties especially concerning the winter forest conditions.

Input data

The model is not run in a strict routine schedule. Instead the meteorologists decides when the model is run, depending on the weather situation. There is no automatic analyses system to chose input data for the model. Therefore before running the model there is first some subjective analyses work to be done in order to chose the input data and then to put these into the computer.

As the base for chosing the input data we use the available weather analyses and forecast, special trajectory forcasts from the LAM - model from Swedish Meterological and Hydrolological Institute (SMHI), surface observations and soundings in the region and possible special aircraft observations of cloud bases and cloud tops. Also the sea surface temperature charts from SMHI, distributed 2-3 times a week, are very important information.

In order to reduce the amount of parameters that has to be put into the computer a database with preparerd trajectories and corresponding latitudes and longitudes and surface characteristics along the trajectories are stored in the computer. If no prepared trajectory is suitable there is the possibility to design a special "own trajectory" composed from a data base of surface types with corresponding parameters for the surface characteristics. When chosing initial temperature and humidity profiles one can not just take a sounding and put it into the model. Instead one has to use all available observations and make a subjective analyses of the profiles taking into account cloud observations, radiation and turbulence conditions etc. The initial liquid amount of water in stratus or fog is defined by a "fictious" dew point (including all liquid water) and some thumb rules telling how to chose the values in stratus and fog. The method of defining cloud water by a "fictious" dew point, including the liquid water, gives automatically a natural coupling to the temperature and is a more natural way of defining liquid water for an operating meteorologist, than for example giving it in g/m^3.

Results

Experiences are now available from test and operational use at the Air Force Base F21 in Lulea during more than 2 years. At F10 in Ängelholm there are experiences from only 4 months and no verification is yet made for F10. Unfortunately, it has not yet been possible to compare systematicaly the model predictions with independent forecasts made by conventional methods. Instead we have made a subjective verification with scale values "good", "medium" and "bad". The score is defined in that way that the value "medium" gives useful but not perfect information to the forecast. The verification was based on the fog/stratus situation and the temperature at the end of and along the trajectory. A number of 143 cases at F21 have been analysed giving 57% "good" 23% "medium" and 19% "bad" cases.

Proposals for further development

There still remain problems in the model that has to be corrected. For example the discription of surface and the need for more surface types. The surface albedo values used in the model have to be refined especially for forests in winter time. Another improvement that should be made is to introduce some form of initialisation, eg adopting the model calculation to two consecutive observations before starting the prediction.

To improve the physics of the model the evaporation from falling precipitation and the fall out of cold stratus and fog precipitation and orografic precipitation should be introduced. The evaporation from the vegetation also need improvements. There is also a suspicion that the eddy exchange coefficient for heat is too small when starting a prediction in very stable stratification.

Finally, there are still some model faults that have to be eliminated, eg the model now allows the temperature in snow to increase above 0 degree Centigrade. Another fault is the calculation of heat flux through ice without snow, which gives to large values in daytime.

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Nesting of a High Resolution Planetary Boundary Layer Formulation in a Regional Forecast Model

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In these brief talks I will review some of the boundary layer modeling efforts which have occurred or are under development at the Naval Environmental Prediction Research Facility (NEPRF). In my first presentation, I will cover the development of a new boundary layer formulation for the Naval Operational Regional Atmospheric Prediction System (NORAPS).

NORAPS is a limited area forecast model, generally run with a horizontal grid spacing of ~80 km and a 109 x 82 x 12 model domain. The new boundary layer formulation which we are developing for NORAPS uses a "level 2.5" closure (using the terminology of the Mellor and Yamada (1974) classification hierarchy) in which the turbulent kinetic energy equation is solved prognostically; other second order moments are calculated diagnostically. A complete long and shortwave radiation code is used to compute fluxes in-cloud as well as clear air. The new model physics includes nonconvective cloudiness and precipitation, with liquid water content and the fraction of cloud contained in a grid volume computed (Sommeria and Deardorff, 1977). Over water, stability dependent transfer coefficients are used with bulk formulations for the surface fluxes (Louis, 1979). Over land, we will test prognostic surface heat and moisture budget expressions that simulate the presence of a plant canopy (Deardorff 1978), however, it may well be necessary to use a simpler parameterization ultimately. The current NORAPS model contains a Kuo-type cumulus parameterization which we will continue to use.

A unique feature of our new paramerization for the regional model is the nesting of a high-resolution boundary layer grid within the coarser mesh of the regional model. Thus, at each grid point of the regional model (spaced ~80 km apart) we compute relatively high-resolution boundary layer profiles of wind, temperature, and humidity. Figure 1 shows schematically the structure of the nested grid in the vertical (the figure stops at $\sigma = 0.4$, but in actuality the model extends to $\sigma = 0.1$). Therefore, at each point of the regional model in the horizontal, we will have a

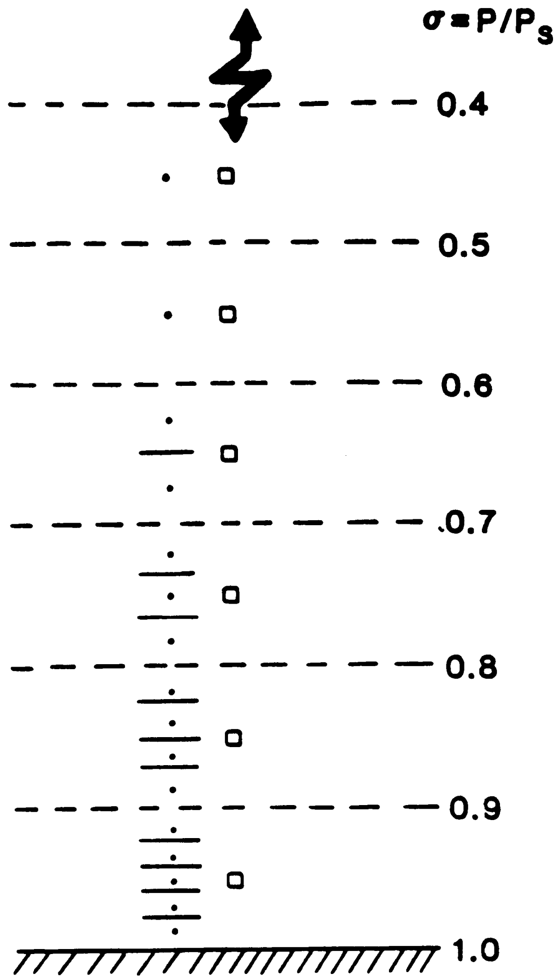

Fig.1 Distribution of fine mesh point nested within NORAPS coars grid. Dots are fine mesh points, squares are coarse grid points.

vertical structure as shown in Fig. 1.

The coarse grid and the fine grid are interactive every timestep. Large-scale tendency terms from the dynamic calculations done on the coarse mesh are interpolated to the fine mesh as forcing functions. The profiles computed on the fine mesh, in turn, are layer-averaged for communication back to the coarse grid. Thus, although not supplying a forecast "sounding" at a particular point, at each gridpoint we will have a forecast representative of boundary layer structure averaged over a domain 80 km on a side. This kind of boundary layer detail conceivably could be useful to a model which wished to make a detailed forecast on a smaller scale, e.g., an Air Mass transformation Model.

A question that arises from the construction of such a model is "why not use a bulk boundary layer formulation?". Part of the answer certainly lies in the personal knowledge and basis of the model builder! Beyond that, however, I believe it is necessary to look at the requirements for formulating the boundary layer physics in the two types of models. In the simplified approaches to second order closure (level 3, 2½, 2, etc.) the formulation describing entrainment at the top of the boundary layer is very little altered whether dealing with a clear or cloudy boundary layer. If the modeler uses conservative variables (e.g. liquid water, potential temperature and total moisture) he is automatically in a position to deal with a cloudy boundary layer. The closure constants used in the model are derived from laboratory data and do not need to be altered for each new physical situation. On the other hand, entrainment parameterization used in bulk models become much more complex when dealing with the cloudy boundary layer and their formulation continually strains the ingenuity of the model designer as new situations are encountered. The history of these cloud-top entrainment formulations is such that they have become increasingly complex over the years, more intensive users of computing power. It must be acknowledged that they have also become progressively more accurate, and thus it is still very difficult to decide which path is the best here.

In my second talk I will discuss results from a model intercomparison study in which the ability to forecast fog/stratus development and dissipation was tested. Five different research models were run as trajectory models. The modelers were supplied with initial conditions and the surface forcing, but the verifying conditions were withheld. There were six different test cases, which included advection and stratus-lowering fogs, as well as diurnally varying stratus conditions. The results from this study are presented in extensive detail in a Calspan Report [Mack et al, 1983). The data necessary for others to exercise their models is there, and may represent a good starting point for those adding moisture parameterizations to their trajectory models.

I also discuss results of a comparison between a third-order closure model (Bougeault, 1986) and a streamlined version of the "level-3" closure model that I work with. In this version of my model, the number of points in the vertical has been reduced to 25, the time step increased to five minutes, the radiation portion of the code called only every ½ hour, and extensive code optimization conducted. This has made possible conversion of the model to a desktop minicomputer (HP-9845).
With this model, I set up the initial conditions as described by Bougeault (1986), and found that many features of my model simulation of the behavior of a stratus deck over a seven day period were in agreement with the third-order closure model results. However, Bougeault showed a large liquid water content (near 1 g/kg in his results at 0600), while my values of q_1 peak near 0.2 g/kg at this time. This difference may result from the cloud droplet sedimentation parameterization contained in my model and lacking in Bougeaults. In any case, model validation using a detailed third-order closure model as benchmark provides a good means of further evaluating the performance of a simpler formulation.

In my third lecture I will describe the Navy Operational Local Atmospheric Prediction System (NOLAPS). This forecast system consists of the level-3 1-D turbulence model coupled to the output fields of the U.S. Navy's global forecast model. The global model fields are used to provide advective forcing to the detailed boundary layer forecast. This marine planetary boundary layer model is intitialized with a selected ship sounding . A 24-hour forecast of sounding structure is made and the modified refractive index structure is derived from this. This forecast refractive profile can be input into a separate microwave propagation code to predict atmospheric effects on radar beam transmission.

More recently this model, with its code virtually unaltered, has been tested with an alternative method of initialization. In this technique a "pseudo-sounding" constructed from the large-scale fields is used for the initialization. Clearly the initial profile in this case is very coarse, and the model must grow its own boundary layer. To do this properly, the entrainment process (in cloudy and clear cases) must be working accurately if the boundary layer is to develop realistically. Of course the model response to surface fluxes and large-scale forcing need also be treated well if the proper boundary layer development is to occur. This is one reason why tests of the model against Benchmark codes, such as Bougeaults third order closure model, (under conditions in which no boundary layer structure is evident in the sounding initially) are particularly important.

Recently the NOLAPS model has undergone an operational evaluation. The results however, are still being evaluated as of this writing. This test consisted of running the model in an "on-request" operational form for a 45 day period. U.S. aircraft carriers from around the world supplied initial soundings to the model at the central operational site (Fleet Numerical Oceanographic Center, Monterey, California). The forecast soundings supplied back to the carriers will be compared against verifying observations. After a lengthy, and detailed examination of results during this operational test period, a decision will be made as to whether the model is ready for routine operation usage, or whether further refinements are required.

Further description of this model forecast system, and a discussion of model validation using station ship soundings may be found in Burk (1985); Burk and Thompson (1982); and Thompson and Burk (1983).

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The Penn State/NCAR Model - Treatment of the Boundary Layer

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The boundary layer module is a medium resolution version of an one-dimensional high-resolution model developed by the author (1976, 1979a, 1979b) and adapted by Zhang and Anthes (1982). It uses a force-restore treatment of the soil surface for the temperature boundary condition. Turbulent transports are determined by the local gradients in near-neutral and stable conditions and by a non-gradient type of parameterization under strongly buoyant convective conditions.

The ground is represented in a simplified but efficient way by a slab model developed by the author (1976). The lower boundary is in thermal contact with an infinite reservoir at constant temperature. The upper surface intercepts the net radiation and partitions it in accordance with the surface heat balance equation. In the Penn State version the surface is not vegetated. By suitable selection of the interfacial transfer coefficient at the lower surface and the slab heat capacity per unit area, one can achieve the correct amplitude and phase of the surface temperature of the slab, as determined by an analytical solution of a homogeneous soil model.

In near neutral and stable conditions the calculation of turbulent fluxes for local gradients is justified. The exchange coefficients for heat, water vapor, and momentum are assumed to be equal. The magnitude of the coefficient is determined at each location in time and space by local conditions in a way that may be considered to be an extension of well know empirical conditions in the surface layer. The equation used is of the form

$$K = l^2 S \; f(Ri)$$

where l is a scale length, S is the vertical shear of the horizontal wind, and Ri is the local Richardson number. The function most generally used is $(1 - \underline{Ri} / \underline{Rc})$, where Rc is the critical Richardson number, but better agreement with the surface layer behavior would prevail if the square of this function were used. The behaviour of the prediction is insensitive to the form of the function that is used as well as to the length scale.

Under strong surface heating, the fluxes within the mixed layer are not determined by local gradients, but rather by the relationship between the heated surface layer and the buoyantly agitated layer above it. The temperature of the surface layer, which is nominally 10 meters deep, is free to float in response to the heat input through its lower interface and the losses from thermals that detach themselves and rise buoyantly, entraining heat, to the entire depth of the mixed layer. In effect, each layer exchanges heat with the surface layer rather than with adjacent layers. The amount of heat leaving the surface layer is determined by the surface layer temperature and that of the first layer above, and regulates itself in response to the surface temperature of the ground. The amount of heat determines the amount of mass exchanged in each time step, and the temperature distribution above determines the depth of the layer through which it is partitioned. The level of penetration of the rising thermals is generally somewhat higher than the level where buoyancy ceases, and the downward entrainment of air through the capping inversion plays an important role in heating the mixed layer in the daytime and in maintaining its depth in a subsiding environment. One of the consequences of this model is that the potential temperature of the mixed layer tends to increase gradually upward, and this prediction is exactly what is generally observed above the superadiabatic surface layer under free-convective conditions.

In general the treatment of momentum and water vapor fluxes is identical to that of heat. Under free convection, the exchanged air is assumed to exchange its momentum and water vapor. However, since the amount of the exchanged mass is known, the conservation equations provide the fluxes at the top of the surface layer. As a result, the surface layer budget equations provide the rate of evapotranspiration and wind stress at the interface. In the case of moisture, an availability factor depending on the land use precipitation history is used.

Because of the nonlinear dependence of the Richardson number on wind shear, the calculated values of the exchange coefficient are strongly dependent on the resolution used. It has been found by tests that lower resolution models can be made to behave quite well by increasing the critical Richardson number.

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A local Dynamical Intrepretation Model

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The aim of a local dynamical interpretation model is to improve the results of synoptic models by adaptation to local conditions. In this way, although statistical methods have been employed with some succes, they do not allow an understanding of the physical phenomenom involved in the planetary boundary layer. It is now possible to use dynamical methods as result of recent advances both in turbulent theory and in synoptic prediction. We use an one-dimensional (1-D) boundary layer model including liquid water cycle coupled to a large scale (LS) synoptic model. Here we present the 1-D model with its parameterizations, the coupling procedure and the application of this 1-D model to a fog event simulation.

Basic equations of the 1-D model

They are derived from the Reynolds system, with the Boussinesq approximation. The concept of exchange coefficients is used for the vertical divergence of turbulent fluxes.

Dynamic equation

In this case we can write

$$\frac{\partial U}{\partial t} + \bar{u}\left(\frac{\partial U}{\partial x}\right)_{LS} + \bar{v}\left(\frac{\partial U}{\partial y}\right)_{LS} + (\bar{w})_{LS}\frac{\partial U}{\partial z} = -if\,(U - U_G) + \frac{\partial}{\partial z} K \frac{\partial U}{\partial z}$$

where $U = \bar{u} + i\bar{v}$, $U_G = u_g + iv_g$, f is the coriolis parameter and K the wind exchange coefficient.

Thermodynamic equations

They are derived from the conservation of moist static energy. Thus we can use conservative variables through condensation processes.

$$\bar{\theta}_l = \bar{\theta} - \frac{L}{Cp}\frac{\bar{\theta}}{\bar{T}}\bar{q}_l, \quad \bar{q}_W = \bar{q} + \bar{q}_l \quad \text{(vapour + liquid)}$$

$$\frac{\partial \overline{q}_W}{\partial t} + \overline{u}\left(\frac{\partial \overline{q}_W}{\partial x}\right)_{LS} + \overline{v}\left(\frac{\partial \overline{q}_W}{\partial y}\right)_{LS} + (\overline{w})_{LS}\frac{\partial \overline{q}_W}{\partial z} = \frac{\partial}{\partial z} K_q \frac{\partial \overline{q}_W}{\partial z}$$

$$\frac{\partial \overline{\theta}_1}{\partial t} + \overline{u}\left(\frac{\partial \overline{\theta}_1}{\partial x}\right)_{LS} + \overline{v}\left(\frac{\partial \overline{\theta}_1}{\partial y}\right)_{LS} + (\overline{w})_{LS}\frac{\partial \overline{\theta}_1}{\partial z} = Rd + \frac{\partial}{\partial z}\left(K_\theta\left(\frac{\partial \overline{\theta}_1}{\partial z} - \gamma_g\right)\right) - \frac{L}{Cp}\frac{\theta}{T}\frac{\partial G}{\partial z}$$

where K_q, K_θ are exchange coefficients, $\partial G/\partial z$ introduced by Brown (1976) represents the gravitational droplet settling, and γ_g allows the description of counter gradient heat fluxes. Rd represents the radiative effects.

The different parameterizations

Turbulence

We use the eddy kinetic energy to compute exchange coefficients. The crucial point in this closure is the choice of dissipative and diffusive mixing lengths. We use those of Therry and Lacarrère (1983), because they result from comparison with a third-order closure model of turbulence.

Radiation

For the thermal radiation, the radiative transfer equation is used with the "intergrated" emissivity approximation, in a similar form to the one proposed by Sasamori (1968). In cloudy conditions, we use an absorptivity due to liquid water: Acloud = 1. -(1.-Aclear) exp(- K_{ext} Uh201) with K_{ext} = 120 $m^2\ kg^{-1}$ and Uh201 is the liquid water path.
For the solar radiation, the Fouquart and Bonnel (1980) scheme is used. It describes Raleigh diffusion, ozone, water vapour, carbon dioxide, and cloud drop absorption including multiple scattering.

Condensation

A subgrid scale condensation scheme is used to simulate liquid water appearance. We suppose that the liquid water content distribution obeys a statistical law in the grid volume with a skewed distribution function following Bougeault (1981).

Earth-atmosphere interaction

The energy exchanges are parameterized in terms of fluxes which are the lower boundary conditions of the 1-D model. To calculate these fluxes, we predict the ground surface temperature and moisture following Deardorff

(1978). The fluxes are computed with these parameters at z_o (roughness height) and at the lowest level of the model with the surface layer universal laws, (even in cloudy air with $\overline{\theta}_l$ and $\overline{q}_W$)

Coupling procedure

Above 3000 m, we assume that the mean variables (U, V, θ,q) obey to the large scale tendency. In the boundary layer, the equation system presented is used by computing horizontal gradients and vertical velocity from the LS model. Between 2500 and 3000 m we use a mixing of the two formulations.

Simulation of a Fog event (L. Musson-Genon, 1986)

This one-dimensional boundary layer model (without advection terms for wind and moisture and LS coupling) is used to simulate a fog event at Cabauw on August 3, 1977. The model is able to describe the mechanisms occuring in fog evolution from its appearance to its disappearance. The data set used is the most complete ever published but it is yet difficult to validate the different parameterizations. Nevertheless we point out again the importance of turbulent transport and radiation, the usefulness of subgrid scale parameterization and we show the ability of the model to quantitatively reproduce the measured evolution.

Concluding remarks

This local dynamical interpretation method is used to simulate the Trappes sounding from 00h to 36h with the french operational model EMERAUDE. This method gives results in clear air condition but not in cloudy air with rain. Therefore, future works consist to include ice treatment and precipitation scheme and to test this method in different typical synoptic situations. We also want to couple the 1-D model with a fine mesh operational model PERIDOT.
For fog application, we hope to use the data from a permanent instrumentation site in north of France to validate the different parameterizations used.

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I-6

The Meteorological Office Mesoscale Model: Its Current Status

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Summary

A numerical forecast model with very fine resolution, is being developed as a short period forecast tool to give detailed guidance on local weather up to a day ahead. The processes represented in the model have been specially developed to take account of the scales represented. Surface synoptic reports are incorporated into the initial data to give mesoscale detail on boundary layer and cloud variables. Trials of the system have shown considerable skill in surface temperature and wind forecasts. The precipitation forecasts are superior to previous numerical models and show a realistic orographic enhancement. Cloud and fog forecasts are still of rather poor quality although recent improvements in the cloud results are very encouraging.

Model characteristics

Dynamics : - Non-hydrostatic, Compressible, "Semi"-implicit integration

Grid : - 15 km horizontal resolution (is normal use)
- levels at 10, 110, 310, 610, 1010, 1810, 2110, 2810, 3610, 4510, 5510, 6610, 1610, 9110, 10510, 12010 m above the ground.
- At this resolution the non-hydrostatic, compressible effects should be negligible.

Radiation: - Surface balance only (cloud top balance being tested)
- Short wave - Cloud transmission depends on total liquid water path & solar elevation.

- Long wave - Surface emissivity = 1
  cloud emissivity depends on total liquid water path
  Lowest cloud base temperature used
  Water vapour emissivity depends on total water vapour path near temperature of bottom 3 layers used.

Precipitation : - Fractional cloudiness from Turbulence scheme.
Snow/ice-cloud falls at 1 ms^{-1} and melts at 0°C.
Rain produced locally & cloud water density is accreted on rain falling from above (much more efficient)
Snow reaches ground if freezing level below 310 m.

Turbulence : - 1½ order closure (after Yamada & Mellor level 2.5) i.e. prognostic TKE, diagnostic mixing length.
Conservative variables θ_1, q_t include phase changes.
Liquid water diagnosed after Sommeria & Deardorff but using top hat probability function.

Surface fluxes: - Ground heat flux determined using one soil temperature and solving diffusion equation from surface temperature. Atmospheric fluxes = C_D * gradient, C_D calculated from Monin-Obukhov similarity as function of Richardson number and roughness.
Moisture availability controlled by "resistance" which depends on weekly calculated real soil moisture deficit, time of day, and wetness of surface deduced from model precipitation/dew/evaporation.

Intialisation: - Surface synoptic reports used to define boundary layer corrections to first guessfield of wind, temperature, humidity.
Cloud base, cloud top and cloud density using rainfall observations.
We hope soon to use satellite data for cloud top and radar data for rainfall rate.
The first guess is a combination of a 6-hour fine mesh forecast (75 km res.) and a 3-hours mesoscale forecast.

Verification

Surface temperature and wind are as good as the human product. Humidity is biassed too high.

Cloud amount is reasonably forecast under most situations but is biassed low. Cloud base is biassed low and is poorly forecast.

Visibility is poorly forecast.

Precipitation is well forecast in a climatological sense but is not very good at the rain/no rain boundary.

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Most recent published reference - Golding & Machin (1984):
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Pub. by ESA.

A Transilient Turbulence Model of the Boundary Layer

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THE DISCRETE, RESPONSIVE FORM

• **An array [c] of "transilient coefficients" can be formed that describe the fraction of air mixed from any grid point (j) to any other grid point (i) during timestep Δt.** Let c_{ij} represent an element of this matrix. If $S_j(t)$ represents the amount of quantity S in grid box j at time t, then the forecast for S at any grid point i at the end of one timestep Δt is represented by

$$S_i(t+\Delta t) = \Sigma_j \, c_{ij}(t,\Delta t) \, S_j(t).$$

S can represent any variable that is conserved during vertical movement, such as potential temperature, specific humidity, wind components, or tracer concentration. For saturated vertical motion, alternate conserved variables can be used instead.

• **The transilient forecast is absolutely numerically stable for a forward timestep.** This stability comes about because of constraints on the values of of the elements of the [c] matrix. Namely, to insure conservation of air mass and of S, and to require that turbulence always increases randomness and entropy, each element of the matrix must be between 0 and 1, and the sum of each row and each column of the matrix must equal unity. The implications of this numerical stability are profound: large timesteps and/or grid spacings may be used if necessary to reduce computational expenses.

• **Transilient turbulence closure is a non-local, first-order parameterization.** Since i and j need not necessarily be neighbors, this turbulence scheme is <u>non-local</u>. The word "transilient" implies "jump over" or "leap across", as based on the Latin root "transilire".

• **The array [c] is a function of the physics causing mixing, and of Δt and Δz.** Since the physics of geophysical flows is constantly changing during the day, [c] is thus a function of time. The Δt and Δz dependence come in to insure that mixing and dispersion happen at a rate governed by the physics, regardless of the specification of, or changes of, the discretization.

• **A "responsive" parameterization for [c] is possible by making it a function of the Richardson Number matrix, [R].** Each element of this matrix, R_{ij}, represents the bulk Richardson number between grid points i and j. This particular form of closure is called responsive because [c] is re-evaluated each timestep in response to the current value of [R]. This scheme is also <u>first order</u> because [R] is a function of mean stability and shear.

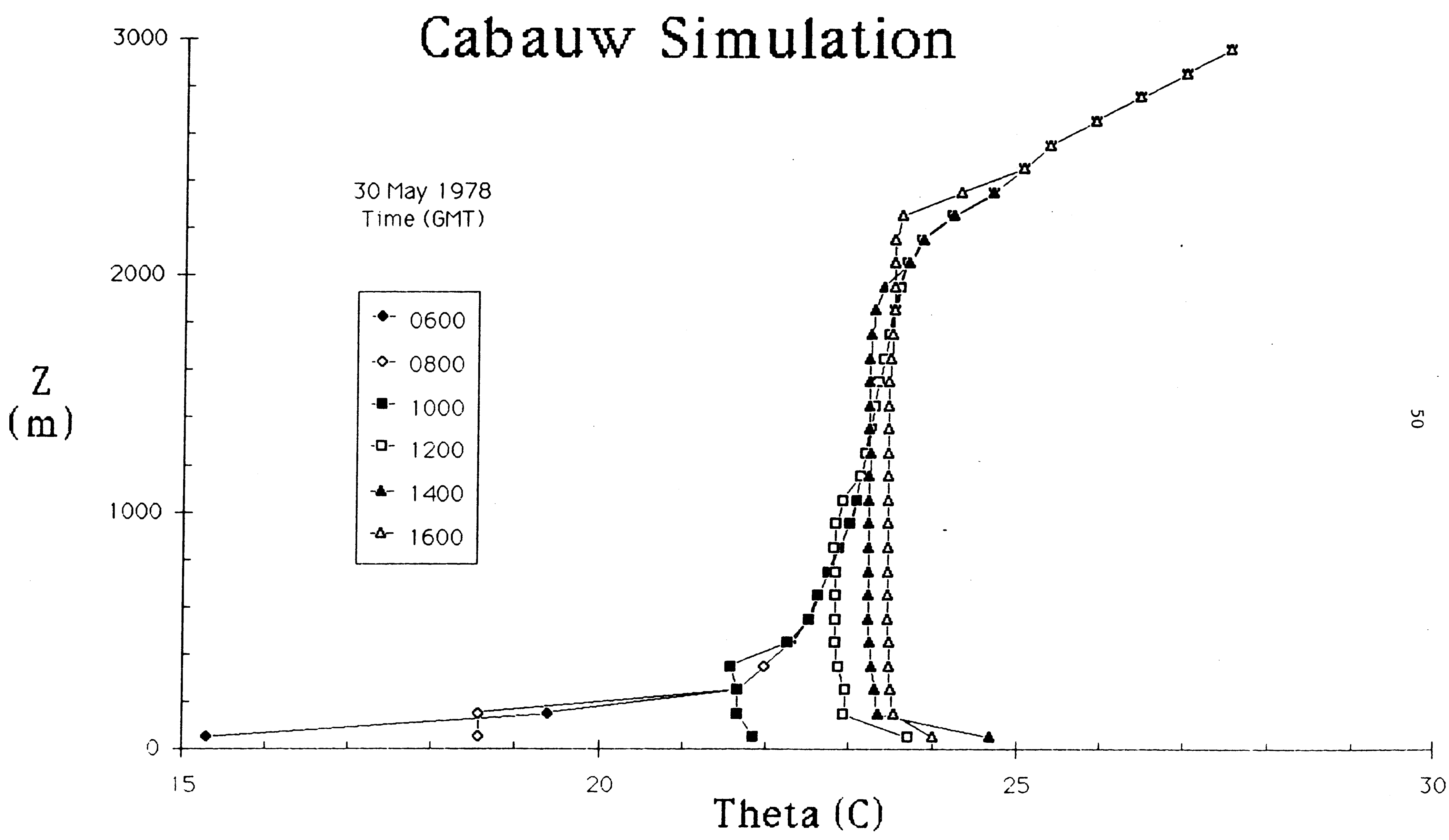
• **One parameterization:** There is no unique parameterization for [c] as a function of [R]; it is a matter of judgment on the part of the investigator to choose an appropriate closure approximation. The scheme presently in use by the author is based on $c_{ij} \propto w_{ij} (1 - R_{ij}/R_T)$, where R_T is a critical Richardson number specifying the termination value of turbulence, and w_{ij} is a weighting factor given by $w_{ij} = \Delta t\, U_0 L_0 / [(i-j)\, \Delta z]^2$. The proportionality factor in the equation for c_{ij} is there because additional normalization must be performed to insure that the sum of each row and column of c_{ij} equals unity. Turbulence is assumed not to occur until R_{ij} is fallen below a critical onset value of the Richardson number, Rc. Presently in use are $R_c = 0.21$, and $R_T = 1.0$. The only free parameters are thus: R_c, R_T, and the product $U_0 L_0$.

FORECAST PROCEDURE

• **Turbulence is viewed as an attempt by the atmosphere eliminate instabilities generated dynamically and thermodynamically.** Namely, if static or dynamic instabilities form by body or external forcings, then turbulent mixing occurs in such a way as to partially undo the original instability. This is nothing more than an adaptation of ,eChatelier's principal. Such a concept is an intimate part of the timestep implementation described next.

• **Each timestep is split into two parts: dynamics and mixing.** In the dynamics part, body forcings such as Coriolis force, pressure gradient force, diabatic heating, precipitation fallout or evaporation, chemical reactions, advection, and similar processes are applied, using whatever timestep schemes are appropriate. During this first part, boundary conditions are also applied, to change the values of variables at only the one grid point adjacent to the boundary. Then, the turbulent mixing part is applied as follows: 1) Calculate the [R] matrix; 2) Solve for the [c] matrix; 3) Use that one [c] matrix to mix each conservative variable in the model.

• **Operations per timestep are proportional to n^2, where n is the number of grid points in the column.** No matrix inversion is involved in this transilient parameterization. The matrix multiplication of [c][S] can also be vectorized on some computers for greater efficiency. The [R] and [C] matrices are symmetric, hence storage requirements involve only about half of each matrix.


Cabauw Simulation
30 May 1978
Time (GMT)
0600
0800
1000
1200
1400
1600
Z
(m)
3000
2000
1000
0
15
20
25
30
Theta (C)

Surface Fluxes and the Treatment of Vegetation: General Introduction.

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The logarithmic distribution of variables near the boundary makes it impractical to model them all way to the edge. Instead, the fluxes must be parameterized in terms of the values of the variables at some finite height a. The appropriate surface flux equations are analogous to Ohms law. The aerodynamic resistance can be evaluated from the stress equation using wind profile similarity, and the application of the resistance to other fluxes is called the Reynolds analogy. Because of the special nature of momentum transfer at a rough boundary the Reynolds analogy is unlikely to be completely successful.

In general, models require the values of the variables at the boundary. In some cases these can be predicted from external considerations, as for example, the sea surface temperature. Usually, however, the flux is known or calculated from a budget equation and the surface values are derived from the aerodynamic resistance laws.

The importance of latent heat in the surface heat balance equation requires the water budget of soil to be determined. The water flux in soil depends on both the water density gradient and the temperature gradient. Heating the surface tends to drive moisture downward but vegetation works in the opposite way. Therefore, the treatment of vegetation is necessary. The combined fluxes from the vegetation and ground can be treated as two parallel resistances in series with the aerodynamic resistance.

The vegetative resistance can be treated as a stomatal resistance in series with an interfacial resistance connected with the transfer from the leaf surfaces to the air around them. The stomatal resistance depends on the rate of insolation and the soil water content in the root domain. The latter dependence requires a treatment of the plant water budget and its control on stomatal resistance. A model developed by the author is described.

The presence of a vegetative canopy also requires a network for the treatment of the heat fluxes from the ground and canopy.

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Surface Fluxes and the Treatment of Vegetation: Practical approaches

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The fluxes of heat, water vapour and momentum are determined by interactive processes that take place within the atmosphere, the vegetation layer and the soil. To describe these processes properly detailed models are needed, containing several non-meteorological parameters, such as albedo, roughness length, zero-plane displacement, plant architecture, depth of the root zone, conductivity for heat and liquid water of the soil, depth of the ground water table etc.

For short range weather forecasting these models are two complicated and simplifications have to be made. The "single-leaf" approach, known as the Penman-Monteith equation, is an example of such a simplified model. It requires as input the available energy (net radiation minus soil heat flux), the temperature, humidity and windspeed at reference level (usually 2 m) and the surface resistance. The latter depends on soil moisture, etc. The Penman-Monteith equation is used extensively in agriculture, hydrology and recently in meteorological models. Nevertheless, it still is too complicated for practical applications.

A further simplification was made by Priestley and Taylor (1972). They found that the fluxes for well watered surfaces are primarily determined by net radiation and temperature. Several authors applied the Priestley-Taylor method also to non-well watered surfaces by taking the parameter α dependent on e.g. soil moisture content. De Bruin and Holtslag (1982) showed that this modified Priestley-Taylor approach yields similar results as the more complicated Penman-Monteith formula. This can be explained by the fact that a) the surface fluxes are interrelated with the input parameters for the Penman-Monteith equation. (De Bruin, 1983; McNaughton and Spriggs, 1986) and b) the input parameters itselve are interrelated (De Bruin and Holtslag, 1982).

The modified Priestley-Taylor approach is applied in the KNMI-AMT model (Reiff et al, 1984).

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Surface Fluxes in the Air Mass Transformation model at KNMI

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In the AMT Model at KNMI, temperature and humidity profiles are calculated along predicted trajectories (Reiff et al., 1984). For the development of the Atmospheric Boundary Layer profiles, the surface fluxes of heat and momentum are needed in terms of predictable quantities. Above the sea the fluxes are calculated by bulk relations between the atmosphere and the surface. We use drag coefficients as a function of stability as proposed by Burridge and Gadd (1977). The sea surface temperature is prescribed along the trajectory, based on observations in the preceeding period. During the forecast period the sea surface temperature is held constant and the sea surface is assumed to be saturated. Above land however, the surface temperature may have a strong diurnal variation. For that reason we parameterize the surface fluxes with the aid of the surface radiation and energy budget. During daytime we use a scheme by Holtslag and Van Ulden (1983). The input parameters of this scheme are solar elevation, total cloud cover, air temperature and wind speed. The total cloud cover has to be provided along the trajectory by the meteorologist. During nighttime we use a simplification of the approach by Van Ulden and Holtslag (1985). Here the vapor flux is taken zero and the sensible heat flux is taken proportional to the friction velocity. In the presentation we discuss the performance of the schemes in comparison with observations at Cabauw, the Netherlands. Possible improvements of the scheme are discussed. These may include the estimation of solar radiation and the partitioning of sensible and latent heat over the available energy above land.

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II-4

Treatment of the Surface Fluxes and Vegetation in the Swedish Numerical Boundary Layer Model

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Abstract is included in model description of I-2.

A Phenomenological/Statistical Approach to Fractional Cloudiness Parameterization

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TYPES OF FAIR-WEATHER SCATTERED CUMULUS

• **Scattered fair-weather cumulus clouds should be classified into forced, active, and passive clouds.** Forced clouds are the (semi-)passive tops of mixed layer thermals. They neither remove (vent) air from the mixed layer, nor do tney produce a cloud-induced subsidence between clouds. Active clouds have achieved positive buoyancy and can grow independently of the evolution of the mixed layer thermal that triggered it. They vent pollutants and induce subsidence. Passive clouds are disconnected from the mixed layer, but continue to shade the ground.

• **To model forced cloud onset time and cover, one must consider the variations in local lifting condensation level (LCL zone), and the variations in local top of the mixed layer (entrainment zone).** The inclusion of these zones makes the cloud forecast more accurate, and less sensitive to forecast errors than does the neglect of these zone (ie, than considering only mean LCL or mean mixed layer depth).

• **The fraction of forced clouds is proportional to the convolution of the probability distributions of the LCL and the thermal height, within their respective zones.** [See Wilde, et al (1985), *JCAM*, **24**, 640-657.] This is represented schematically by the overlap in LCL and entrainment zones in the figure below.

LIFTING CONDENSATION LEVEL

• **Each thermal can have its own lifting condensation level (LCL), which may be different from the average LCL.**

• **Forced cloudcover fraction is proportional to the number of thermals that reach their individual LCL**, (not the number of thermals that reach an average LCL).

• **The range of individual LCLs within a subgrid scale region is called the LCL zone.** An estimate of this zone thickness and location is possible using surface layer air (see following remarks about thermal cores). Typical values of LCL zone thickness observed during BLX83 over a mixture of farmland, pasture, and irigated fields ranged from about 200 to 500m in thickness during the day in May and June. The LCL zone is roughly centered on the mean LCL height. The probability of finding a LCL different from the mean decreases with distance of the individual LCL from the mean.

• Neglect of the LCL variation can lead to the following problems in diagnosing forced clouds:
1) Onset time of clouds is too late (by up to a few hours in some cases).
2) Cloud cover becomes too large (overcast)
3) Diagnosed clouds linger longer in the evening than observed clouds.

ENTRAINMENT ZONE

• Lateral entrainment into many thermals is small enough to leave an undiluted core to the thermal. This core can carry air with surface layer humidity up to the top of the mixed layer. This core air will be associated with the first cumulus clouds, because the core air is the moistest air near the top of the mixed layer.

• Air in the entrainment zone can be grouped into three categories: 1) unmixed air from the surface layer, 2) unmixed air from the free atmosphere, and 3) air that consists of the mixture of the previous two groups. The surface layer air is there because of the undiluted core of thermals previously described. The free atmosphere air results from the entraiment of air from above the mixed layer. Deardorff's entrainment zone results don't distinguish between the mixed and surface layer air; therefore, they don't adequately resolve the fact that the mixed layer is drier and less likely to form clouds than the surface layer air.

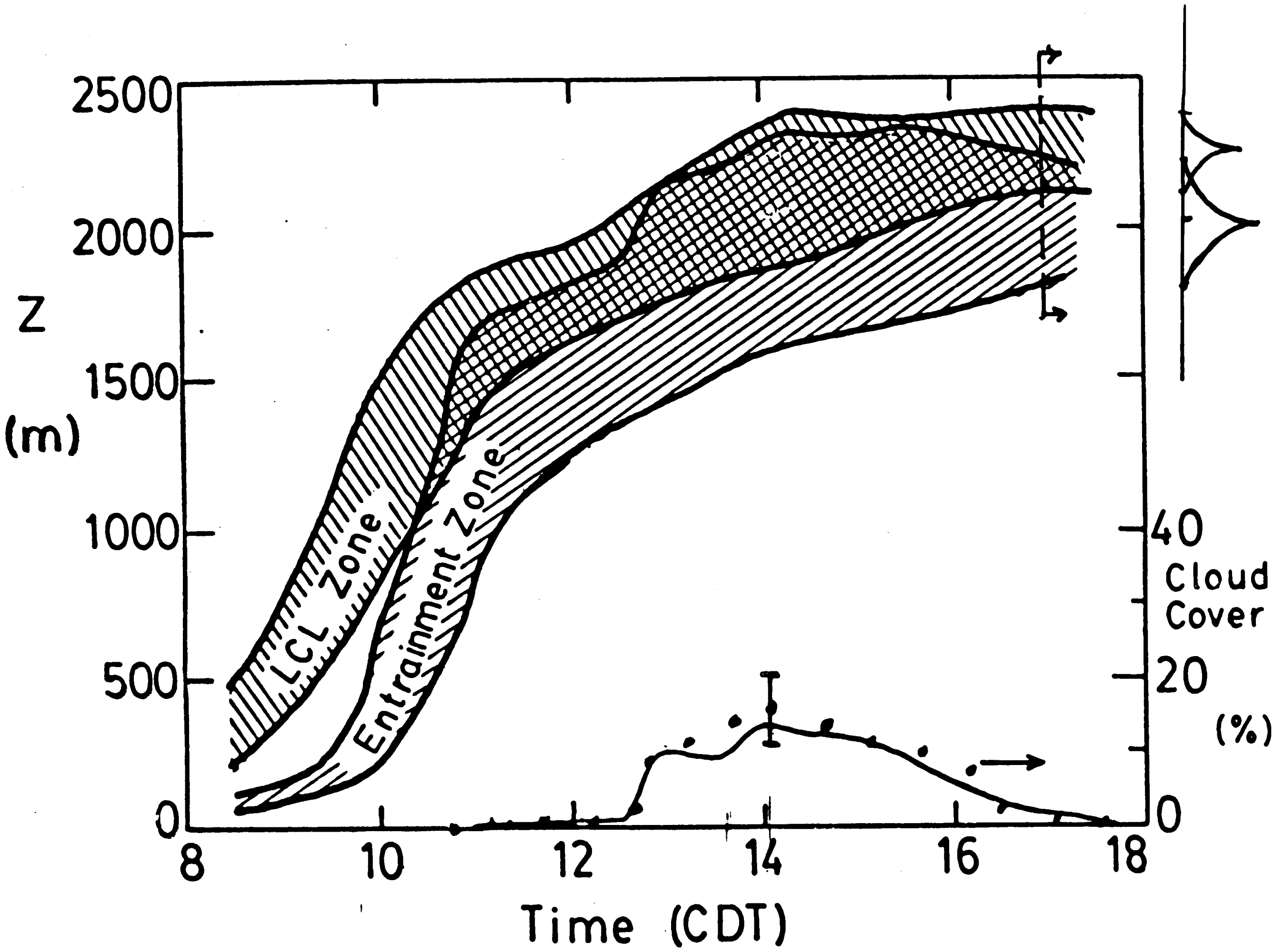

Fig.1 : The zones of local lifting condensation level (LCL) and variations in the top of the mixed layer (Entrainment Zone), together with the fraction of cloud cover as a function of time.

A Fog Case Study with a Local Dynamical Interpretation Model

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Abstract is included in model description of I-5.

Modelling of Stratocumulus fields

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Stratocumulus (Sc) clouds may cover extensive areas of the world, often exceeding 10^6 km^2, for a considerable period of time. Apart from their frequent occurrence, the importance of Sc-decks lies primarily in the large changes in the radiation balance which accompany it (important for climate and local weather forecasting) and the large changes in the interaction between the ocean and the atmosphere (important for climate, air pollution modelling etc.).

In a horizontally homogeneous Sc-deck the evolution of the cloud depends upon the combined effect of different physical processes such as, wind shear at the surface, wind shear at ABL top, a buoyancy flux at the surface, long wave radiative cooling at cloud top, shortwave radiative heating inside the cloud layer, phase changes and subsidence. Improved understanding of the cloud-topped ABL has been gained from detailed observational studies. Brost et.al. (1982 a,b) analysed data collected off the coast of California and Nicholls (1984) described a case of stratocumulus over the North Sea. The observations by Brost et al. (1982 a,b) and Nicholls (1984) showed that different combinations of physical processes may lead to a totally different turbulent structure of the ABL.

A model is developed that incorporates the most important physical processes (Duynkerke and Driedonks, 1986), to study their influence on the turbulent structure of the cloud-topped ABL. Moreover the model results are compared with observational data of Nicholls (1984) and Brost et al. (1982 a,b). From a comparison of the model results with the observational data of Nicholls (1984) and Brost et al. (1982 a,b) it is clear that with the model the observed turbulent structure of the Sc-deck can be simulated quite well.

The model is an ensemble-averaged model with multiple layers in the vertical. Turbulence closure is formulated by using an equation for the turbulent kinetic energy and either a diagnostic formulation of the integral length scale or a parameterized version of the dissipation equation. More details on the model are given in Duynkerke and Driedonks (1986). The model is also used to study the diurnal variation of a Sc-deck over sea. The sea surface temperature is a few tenth's of a degree higher than the air above it. In the early morning longwave radiative cooling at cloud top is the most important process, which destabilizes the ABL and produces mixing down to the surface. As the day progresses shortwave radiative heating becomes more important. The combined effect of long wave radiation, short wave radiation and entrainment of warm air from above the inversion is that the cloud layer is heated, whereas the temperature in the sub-cloud layer remains the same. As a result a stable layer is formed near cloud base and the cloud layer and sub-cloud layer are decoupled. This decoupling can be clearly seen from the moisture flux in Figure 1. The dry entrained air is only mixed over the cloud layer whereas the moisture input from the sea surface is only mixed over the sub-cloud layer. In the late afternoon and night the shortwave radiation disappears and the longwave cooling at cloud top produces mixing throughout the whole ABL. From the moisture flux in Figure 1 one can see that in the sub-cloud layer the total water content decreases whereas in the cloud layer it increases, due to the redistribution of the water vapour out of the sub cloud layer (brought in during day time) over the whole ABL.

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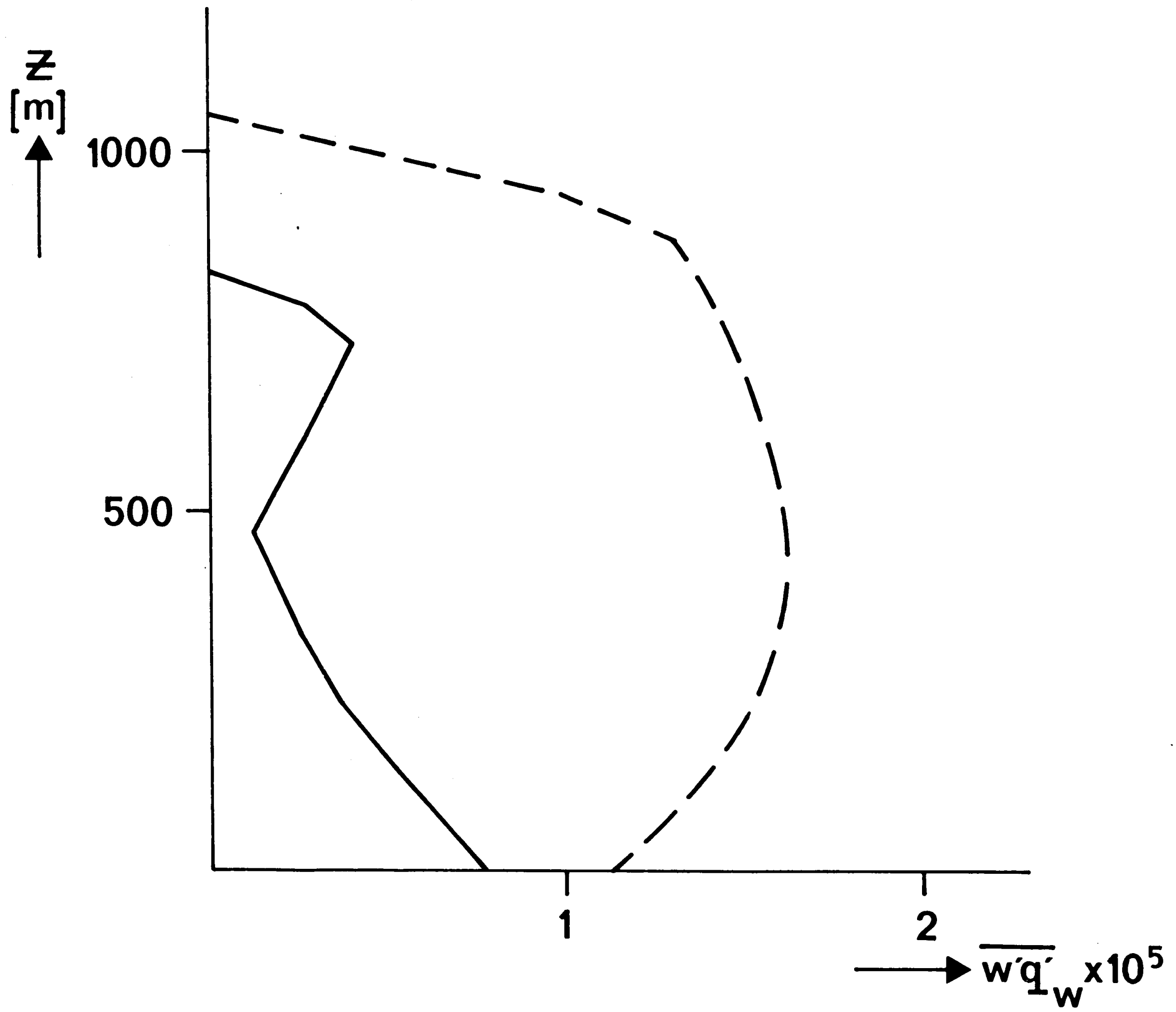

Fig 1: The moisture flux as a function of height around noon (-) and around mid-night (--)

Characteristic Physical Properties of Fog: Requirements for Models

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Important processes during the formation, the maintenance and the dissipation of radiation fog over land have been observed along a meteorological tower. Some suggestions for modelling are illustrated with a simple one-dimensional bulk model with five layers.
Since 1972 radiation fogs are observed at the Cabauw tower with 10 transmissometers up to a height of 180 m. In addition profiles of temperature and wind, radiation parameters, soil heat flux, etc. are measured. In later years also detailed humidity profiles are available.
Radiational cooling of the ground starts around sunset in clear nights. The nocturnal boundary layer grows with a certain rate depending on the geostrophic wind speed. Stable temperature profiles show all kinds of micro-structure: this is mainly caused by differential advection bringing together decoupled air layers from various directions over an imhomogeneous terrain. In the model we concentrate on the average characteristics and assume the accumulated cooling to be distributed over a triangle with prescribed height. However, if the bulk Richardson number exceeds a certain value, a trapezium under a lifted inversion is used.
As the ground fog grows, both in density and height, the droplets together form a closed surface taking over the IR emission from the ground. This closed surface implies, that also the vertical visibility is obstructed.
If the sky gets obscured, important changes are observed. The profiles become well-mixed and the cooling surface rises to just under the fog top. Because the mixed layer is wet-adiabatic, the fog density increases with height. This transition of fog is therefore encouraged by a positive feed-back.

The final stage -after sunrise- is characterised by the start of convention and the lifting of the fog. The rise and entrainment of the mixed layer resemble very much the dry case.
The wind profile also changes with the transition from ground fog to mature fog. The mixed layer is more or less decoupled from the free atmosphere. The fog might even propagate by a thermal wind caused by the

horizontal distribution of cooling.

From the observational evidence we distinguish two different stages of radiation fog: ground fog and mature fog. The last stage may eventually lift as stratus. Ground fog forms in most clear nights somewhere in the country more or less like noise. Its appearance depends very much on local topographic factors. Forecasting ground fog is probably best attempted with statistical methods based on observations and/or model simulations. The -more rare- transition to the second stage is of special interest, because this stage may frequently survive after sunrise and become a nuisance for human activities. With operational interests in mind we therefore try to forecast the occurence of sufficient cooling for the fog transition to occur in favoured locations. An additional cause of mature fog is its advection from the sea.

The transition of fog is seldom observed to occur gradually. At least in Cabauw ground fog is usually overrun by mature fog that formed elsewhere. The mature fog stage of radiation fog is therefore rightfully called "advection fog" although advection might not be the cause of its formation. Advection, of course, complicates local observations. Mature fog is not very sensitive to surface properties: it behaves like a rather stable radiation machine that can propagate over hundreds of kilometers. This characteristic is another justification for our special interest in forecasting the mature fog stage. The dispersal of mature fog or stratus by the sun is equally contagious. Once a hole has been burned, the surface heats so that newly arriving cloud will dissolve too. The fog height has a strong influence on the possibility of dispersal by sunlight: a thick fog layer needs more, but receives less solar radiation!

If we want to forecast whether the cooling will be sufficient for the fog to obscure the sky, we need careful estimates of the surface fluxes. The first problem to be answered, however, is why fog forms at all? One would rather expect dew formation to be the dominant process. In a K-model over bare ground the dew point decreases at the same rate as the temperature. At least two factors assist in fog formation. Firstly the air and its aerosol content cool by radiation. Secondly, altough the top of the vegetation cools after sunset, evaporation from the warm soil and the plants will go on for a while.

If only ground fog is present at sunrise the amount of dew (attracted from the air and also from the soil) will typically be at least 30 times larger than the liquid water present in fog over the same area. So dew formation will retard the formation of ground fog! In mature fog the settling of fog droplets is important and also their capture in the vegetation. This capture is caused by the stronger surface wind in a mixed layer. The lifetime of droplets in mature fog may be less than 20 minutes. The production rate in that stage exceeds the dew deposition, which might then even stop and change to evaporation. A rough estimate of the condensation rate near the fog top is 30 g./m^2 per hr. From a radiative cooling of 60 W/m^2 about 30% is needed for the fog production, the rest is used to maintain the turbulence, to compensate entrainment and subsidence or eventually to intensify the fog. After sunrise the wet vegetation may retard the dispersal with an extra one or two hours. This is a reason why a fog model must be able to forecast vegetation wetting.

Modelling the microphysics is necessary to account for the settling and capture of the droplets and also for the conversion of liquid water content to shortwave and longwave optical properties. By approximating the droplet spectrum as a gamma-distribution all these aspects can be described by only two parameters: the droplet concentration and the width of the distribution. It might further be necessary to account for the aerosol concentration in the radiation computations.

The vegetation is essential for the fluxes of radiation, momentum, sensible heat, vapour and liquid water. We attempted a compatible treatment of all these aspects with a model vegetation described by only two parameters: height and leaf area density. The results for the surface roughness and the (stability-dependent) thermal resistance were tested with measurements of the Cabauw grass cover. The vegetation model was also used to estimate the droplet capture. The procedure followed can easily be applied to other types of vegetation.

A radiation fog model has to consider at least four layers below the free atmosphere: boundary layer, surface layer, vegetation and soil. The boundary layer profiles are used to distribute the accumulated cooling and drying. At each time step the condensation and the resulting radiative properties are computed. Surface layer resistance follows from established flux-profile relations. Longwave cooling is applied at the top of the vegetation. The temperature and dewpoint at the top of the vegetation follow from the energy balance at the upper and lower vegetation surfaces. If the vegetation is not wet, its evaporation is restricted by a stomatal resistance. In principle the model will also work for the computation of daytime fluxes. The soil surface is assumed wet and its temperature is related to the soil heat flux by a method adapted from Shaffer. Upper boundary conditions are temperature and humidity at the top of the surface layer which follow from the previous time step.

This simple model -with only a few independent parameters- includes the most important aspects of fog development, and especially the distinction between ground- and mature fog. The input data are: geostrophic wind and cloud amount, both as a function of time, and the initial temperature and humidity deficit at sunset. The horizontal variation of the input and the soil and vegetation parameters can be considered in sensitivity studies with the model.

The preliminary results for a Cabauw fog case seem realistic, but they are difficult to verify because of the ever-present advection. Tuning the model parameters on measurements for one location would therefore be a bad policy. It would e.g. be better to simulate fog cimatology.
The organisation of the model and the input parameters chosen, anticipate direct use for forecasting, e.g. in an Air Mass Transformation (AMT) model.

Fog Forecasts in Operational Models: A Comparative Study

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Abstract is included in model description of I-3.

On the use of High-Resolution Satellite Data for Mesoscale Analysis

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A mesoscale analysis has been developed in the French Weather Service to provide the initial conditions for a short-range numerical weather prediction model over France (mesh size = 35 km, Peridot project cf. R. Juvanon du Vachat, 1983). Since we lack conventional data to define the vertical structure of the atmosphere at this horizontal resolution, we make use of radiances data from the NOAA's satellites (HIRS-2 radiometer). These data have a resolution which is approximately the same as our mesh size and they are directly inserted in the analysis scheme (optimal interpolation : 3-dimensional, multivariate) without any retrieval procedures.

In addition to the HIRS-2 radiance data, we also use cloudiness information deduced from the AVHRR instrument (Phulpin et al., 1983) in order to discriminate between clear sky pixels or cloud-covered pixels. Up to now, we only insert clear-sky radiances in the analysis or a humidity bogus for completely covered pixels. We are now currently developing a method in order to use partially covered pixels, on the basis of the AVHRR information.

In the presentation of the analysis with satellite data, much attention will be given to the characteristics of the satellite information (Smith et al., 1979) and to the direct assimilation of radiances (Durand et al, 1983, 1985). The problem of the sensitivity of this analysis and its subsequent forecast to the satellite information will be discussed with the results of some experiments.

Finally, some consideration will be given to other approaches, which are known as the Physical Retrieval TOVS Package (Smith et al., 1983 and 1985) and the Improved Intialization Inversion Method (Chedin et al., 1985).

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Analysis of boundary layer parameters

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Models for operational weather forecasting have to be provided with an analysis, that is a set of initial values for the prognostic variables. The analysis should be based on observations. Boundary layer models, however, usually have prognostic variables that are not regularly observed directly. The analysis then can resort to a set of pseudo-observations, that are generated with a model from observational data.

Such generation of pseudo-observations is common in existing analysis procedures. For example, observations at earlier times are extrapolated with a prognostic model to the so-called first-guess field at the analysis time, and wind data are converted with a diagnostic relation based on geostrophy to surface pressure gradient data in many surface pressure analyis schemes.

A boundary layer model can be provided with an initial wind profile in the lowest few tens of meters by the use of a model developed by Holtslag (1984). This model uses flux-profile relationships, where the surface momentum and heat fluxes over land are estimated from the synoptically observed 10-meter speed U_{10} and cloud cover. The applicability of this model has been extended to areas over sea, by estimating the surface fluxes from U_{10} and sea- air temperature difference. Thus pseudo-observations of the low level wind profile (up to a height of, say, 100 m) and of the fluxes in terms of the friction velocity U_* and Monin-Obukhov length L are available at almost all SYNOP and SHIP reporting positions. Contours of a field of L, straightforwardly interpolated from those pseudo-observations, will be presented.

Often a boundary layer model carries the boundary layer height as a prognostic variable. The pseudo-observations of this quantity, reguired for an analysis, can be obtained with the air mass transformation (AMT) model (Reiff et al., 1984). In this analysis made of the AMT model, the model is run along the 24-hour trajectories that start at the available radiosounding stations. Usually, the generated set of pseudo-observations has a fair distribution over land and sea areas; yet, every now and then additional pseudo-observations have to be added. The boundary layer height may vary strongly at the land sea transition. Therefore the horizontal interpolation scheme has been devised to concentrate the gradients along the coast.

At the workshop the analyses of the boundary layer height through a summer day will be presented (Cats et al. 1985). In the shown cycle, the additional data had been manually generated. Procedures to generate them automatically are under investigation. Fully objective analyses of boundary layer height promise to be feasible along these lines.

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III-3

Operational experience with the AMT model at KNMI

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The Air Mass transformation Model is in semi-operational use at KNMI since spring 1984. About three experienced forecasters used the AMT-model in an operational surrounding, during which the input and output procedures were defined and small changes in the model were made. Since spring 1985 the AMT-model is in operational use, which means that about 8 different forecasters run the interactive version of the AMT-model.
The results at the +12 hour forecasts of the AMT-model, starting at 00 GMT and ending in De Bilt, the Netherlands (approximately 60 km from the sea) at 12 GMT are given in Table 1.

When one compares the MAE and MAE* for T_0 for the different seasons it is seen, that the temperature forecast of the model is still slightly worse than the forecaster. However, one has to take in mind, that the forecaster does not forecast for a specific time 12 GMT, as the model does, but forecasts the maximum temperature which is easier. Further, the results of <u>all</u> AMT-model forecasts are taken into account in these numbers. It may well be, that a well-defined subset of these cases scores better than the forecaster does. Further research is needed on this point.

The strength of the model at the moment is its humidity forecast. The MAE and CC of q_0 perform well in winter and spring. In summer the scores decrease, but that is probably due to a program error which was introduced during a program change at the end of spring 1985.

How good the ABL-height is forecasted, is diffucult to judge as the ABL-model forecasts are an area-averaged h, while the observed h is a point value taken from the radiosonde. It is well known, that this value varies considerably from place to place (over hundreds of meters). The 700 and 500 mb scores are comparable with those obtained from the ECMWF-model (not shown here).In general the winter forecasts in the ABL are worse than the spring forecasts as may be expected.

The results of the 24-hour forecasts starting at 00 GMT and ending at 00 GMT in De Bilt are given in table II. In comparing MAE and MAE* for T_0, one sees that the AMT-model results are still slightly less than the forecaster. Only in summer the difference increases, probably due to a program error as said before.

Table 1

Verification results of the +12 hour AMT-model forecasts

		T_0	q_0	h	T_7	q_7	T_5	q_5
winter	N	14	14	14	14	14	10	10
'85	M	-1.3	3.1	536	14.4	0.66	23.7	0.28
	ME	0.02	0.13	-10	1.04	0.33	1.67	0.03
	MAE	2.8	0.42	256	1.8	0.52	1.8	0.09
	MAE*	2.2						
	CC	0.85	0.96	0.74	0.96	0.08	0.99	0.93
spring	N	30	30	29	30	30	29	29
'85	M	12.7	6.7	1002	24.0	2.10	34.8	0.62
	ME	-0.02	-0.20	-77	0.62	-0.06	-0.74	0.03
	MAE	1.8	0.66	438	1.8	0.66	1.7	0.24
	MAE*	1.7						
	CC	0.93	0.91	0.38	0.87	0.77	0.86	0.68
summer	N	56	56	56	56	56	50	50
'85	M	16.7	8.5	1227	27.5	2.45	38.5	0.86
	ME	-0.79	0.79	-174	0.20	0.23	0.57	0.12
	MAE	1.5	1.20	578	1.3	0.79	1.5	0.41
	MAE*	1.2						
	CC	0.86	0.69	0.28	0.96	0.64	0.93	0.63

N = Number of cases

M = Mean of observed values

ME = Mean error : $\Sigma (X_{mod} - X_{obs}) / N$

MAE = Mean absolute error : $\Sigma \| X_{mod} - X_{obs} \| / N$

MAE* = Mean absolute error of the forecaster for the maximum temperature

CC = Correlation coefficient

T_0, T_7 and T_5 are the potential temperatures at 1.5 m, 700 mb, 500 mb (in °C) respectively

q_0, q_7 and q_5 are the specific humidities at 1.5 m, 700 mb, 500 mb (in g/kg) respectively

h is the height of the boundary layer (m)

Table II

Verification results of the +24 hour AMT-model forecasts

		T_0	q_0	h	T_7	q_7	T_5	q_5
autumm	N	20	20		19	19	18	18
'84	M	10.1	7.2		26.2	2.28	36.9	0.83
	ME	0.87	0.76		0.41	0.21	-0.83	0.25
	MAE	1.8	0.83		2.3	1.2	2.8	0.50
	MAE*	1.6						
	CC	0.61	0.84		0.83	0.48	0.88	0.48
winter	N	49	49		45	45	24	24
'85	M	-3.5	3.1		15.4	0.93	27.4	0.49
	ME	0.98	0.63		-0.08	0.40	-0.24	0.33
	MAE	2.8	0.77		2.4	0.68	3.0	0.59
	MAE*	2.4						
	CC	0.90	0.92		0.91	0.71	0.74	0.45
spring	N	50	50		49	49	38	38
'85	M	5.7	5.5		21.2	1.87	31.7	0.60
	ME	0.41	0.53		-0.13	0.02	0.20	0.13
	MAE	2.0	0.83		2.5	0.82	2.3	0.30
	MAE*	1.6						
	CC	0.89	0.82		0.88	0.64	0.85	0.55
summer	N	58	58		58	58	48	48
'85	M	11.5	8.1		27.7	2.89	39.6	1.03
	ME	1.55	1.93		0.52	-0.05	-0.92	0.24
	MAE	2.0	2.0		2.2	1.06	3.6	0.66
	MAE*	1.3						
	CC	0.80	0.72		0.87	0.53	0.63	0.32

MAE* is the mean absolute error made by the forecaster in forecasting the minimum temperature. The other symbols are defined below table I.

Also at night the strength of the model, as it is now (the dry version) is its specific humidity forecast. The errors at 700 and 500 mb are still small, and comparable with errors in the forecasted ECMWF fields (not shown here). The absolute error (ME) in q_0 increases in time, as may be expected in this AMT-model version, in which condensation in the ABL and transport of humidity from the ABL into the atmosphere aloft is neglected sofar.

In table II no score is given for the ABL-height, because this height is difficult to obtain from radiosounds during night time.

It is difficult to compare the results of these forecasts with the results of the AMT-model on analysed windfields and observed cloud amounts as published before (Reiff et al, 1984), because here different years are taken into account. However, as we devide our material in cases "advected from sea" and "completely advected over land" we obtain the same results (not shown here) : the AMT-model performs best in cases with advection from sea.

Besides verification of the ABL-parameters T, q and h we also have verified the occurrance of fog and boundary layer clouds (see for definitions Reiff et al, 1984). In table III the amount of forecasted cloud amount (divided into three classes) is compared with the observed cloud amount. In brackets the +24 hour persistency score is given.

From this table it is seen, that 50% of the ABL-clouds are forecasted correct into one of these three classes, 24-hour persistence scores 46%. Therefore this, dry version of the model is not better than persistence. Further, it can be seen that the model forecasts too many times ABL-clouds. This is not surprising, as the present, dry, version of the model makes the ABL too wet, as has been seen already in Table I and II.

Table III

	observed cloud amount		
	0/8	1/8-4/8	5/8-8/8
forecast at 0/8	3 (6)	2 (3)	1 (4)
forecast at 1/8-4/8	9 (8)	5 (12)	7 (18)
forecast at 5/8-8/8	3 (1)	27 (20)	42 (28)

N = 100 (100)

Table IV

	observed	
	no fog	fog
no fog forecasted	123 (124)	14 (19)
forecasted fog	24 (23)	13 (8)
N = 174 (174)		

Table IV compares the +24 hour fog forecast with observations. (See for definitions Reiff et al. 1984). In brackets the results of +24 hour persistence are given. From table IV it is seen, that the model forecasts 48% of the observed fog-cases correct, persistence does that 30%.

From the above results it is clear, that still a lot of work has to be done. The strenght of the present, dry, version of the model is the specific humidity forecast of the ABL, which, as has been shown in Reiff et al. (1982) only can be forecasted correct if the advection is forecasted correct.

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Operational Experience with the Swedish Numerical Boundary Layer Model

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Abstract is included in model description of I-2.

Operational Experience with the US Navy Models

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Abstract is included in model description of I-3.