On the influence of land-surface processes on the near-surface atmospheric state

Author(s):
Beroud, Jean-Marc

Publication Date:
1999

Permanent Link:
https://doi.org/10.3929/ethz-a-003810752

Rights / License:
In Copyright - Non-Commercial Use Permitted
On the Influence of Land-surface Processes on the Near-surface Atmospheric State

A dissertation submitted to the
SWISS FEDERAL INSTITUTE OF TECHNOLOGY
Zürich

for the degree of
DOCTOR OF NATURAL SCIENCES

presented by
JEAN-MARC BEROUĐ
Dipl. Phys.
born on January 24, 1971
citizen of Ecoteaux (VD)

accepted on the recommendation of
Prof. Dr. H.C. Davies, examiner
Prof. Dr. C. Schär, co-examiner
J. Quiby, co-examiner

Zürich, March 1999
# Contents

Abstract vii

Résumé ix

Zusammenfassung xi

1 **Introduction** 1

1.1 SVAT Schemes ........................................ 2

1.2 Recent Interests .................................... 3

1.3 Aims of This Study .................................. 4

1.4 Outline ................................................. 4

2 **Model’s Overview** 7

2.1 The Swiss Model ...................................... 7

2.2 The Soil Model ........................................ 9

2.2.1 The Heat Budget .................................. 10

2.2.2 The Water Budget ................................ 12

2.2.3 Sensible and Latent Heat Fluxes .................. 14

2.2.4 The Interception Reservoir ....................... 17

2.3 Screen-level Interpolation . ........................ 19

3 **Comparisons With Observations** 21

3.1 The Swiss Automatic Observing Network .......... 23

3.2 The Hydrological Station Büel ...................... 23

3.3 The Validation Method ............................... 24

3.4 Comparisons With ANETZ ............................ 26

3.4.1 Shortwave Radiation ............................ 26

3.4.2 Screen-level Variables ........................... 27

3.4.3 Potential Evaporation ............................ 30

3.4.4 Soil Temperatures ................................ 33

3.5 Comparisons With Büel .............................. 34

3.5.1 Global Radiation, Precipitation and Evapotranspiration . 34

3.5.2 Soil Moisture .................................... 38

3.5.3 Temperatures ...................................... 40
4 Sensitivity Experiments 41
   4.1 Case Studies 41
   4.2 Test Area 42
      4.2.1 Soil types 43
   4.3 Method 44
   4.4 Result of Experiments 45
      4.4.1 Vegetation 48
      4.4.2 Soil Moisture 53
      4.4.3 Soil Temperature 55
      4.4.4 Other Sensitivity Experiments 56
   4.5 Local Versus Large-scale Effects 60
   4.6 Discussion 61

5 Model’s Evapotranspiration 63
   5.1 Normalized Evapotranspiration 63
   5.2 Further Analysis of Case Studies 67
   5.3 The 2m Temperature Maximum 73

6 Variational Studies 75
   6.1 The Variational Method 76
   6.2 The Minimization Procedure 77
      6.2.1 The Quasi-Newton Method 78
   6.3 Method and Results 79
      6.3.1 Limitations 79
      6.3.2 2m Temperature Maximum 80
      6.3.3 Daily Evapotranspiration Sum 84
   6.4 Discussion 85

7 Conclusion 87
   7.1 Perspectives 88

A The Force-restore Method 91
   A.1 Derivation 91
   A.2 Outline of the Extended FR-method 92

B External Parameters 95
   B.1 Soil Types and Textures 95
   B.2 Albedo 95
   B.3 Vegetation 96
   B.4 Hydrological Terms 98

C The Variational Analysis Equation 99
   C.1 Derivation 99
   C.2 Illustration 101

References 103
Acknowledgments  111
Curriculum Vitae  113
Abstract

In numerical weather prediction models, the relevant physical processes at the earth's surface are described by soil-vegetation-atmosphere transfer schemes. Temperature and specific humidity, sensible and latent heat fluxes, as well as momentum fluxes, determine the lower boundary conditions for the atmosphere and influence significantly screen-level variables. Research geared to improving the 2m temperature and relative humidity forecasts provided by the operational model of the Swiss Meteorological Institute constitutes the main theme of this thesis.

Systematic errors found in the 2m summertime temperature are investigated through the analysis of soil model parameterizations, comparisons with observations and by the undertaking of sensitivity studies. Particular emphasis is placed upon the influence of orography on near-surface parameters. Finally, a variational technique based on such information is used to correct and improve the 2m temperature behavior.

The so-called extended "force-restore" method used to compute ground heat fluxes assumes constant heat capacity and conductivity values. In addition to this major limitation, the vegetation parameterization appears somewhat artificial, the rooting depth being dependent on time.

Experiments show that the evapotranspiration in rainy conditions and also the 2m temperature amplitude are overestimated by the model. Comparisons with observations from the Swiss Automatic Observing Network and the hydrological station Büel reveal that the 2m temperature maximum appears 1–2 hours earlier than in the observations.

In clear sky conditions, fraction of vegetation cover, rooting depth and soil moisture content control screen-level variables by modifying the Bowen-ratio, i.e. the partitioning between sensible and latent heat fluxes. Sensitivity studies show that evapotranspiration is particularly affected by the aforementioned parameters. In cloudy and rainy conditions, the surface energy fluxes are weak and planetary boundary layer conditions control screen-level variables.

Bare soil evaporation and surface soil moisture content constitute the relevant model parameters at altitudes higher than 1500 m, where vegetation is sparse. In contrast, at lower altitudes, the analysis of the evapotranspiration parameterization shows that deep soil moisture and vegetation transpiration parameters gain importance, although bare soil evaporation still remains a significant process.

A 2D-VAR variational assimilation method is introduced in an attempt to improve the 2m temperature behavior by correcting the soil moisture initial condi-
tions. A penalty function, which measures the degree of misfit between observed and modeled 2m temperature maximum, is chosen for minimization. The addition of a background term related to the soil moisture first guess information proved necessary to obtain a distinct function minimum, as many soil moisture configurations can produce similar evapotranspiration rates. The correction of the temperature minima remains difficult due to the limitation of the current soil model.
Résumé

Dans les modèles de prévision numérique du temps, les principaux processus physiques de surface sont décrits par l'ensemble des transferts sol-végétation-atmosphère. La température et l'humidité spécifique, les flux de chaleur sensible et latente, ainsi que les flux de quantité de mouvement, déterminent les conditions aux limites inférieures de l'atmosphère et influencent de manière significative les paramètres de surface. La recherche portant sur l'amélioration des prévisions de température et d'humidité relative à 2 m fournies par le modèle opérationnel de l'institut suisse de météorologie, constitue le principal thème de cette thèse.

Les erreurs systématiques qui entachent la température estivale à 2 m sont examinées à travers l'analyse des paramétrisations du modèle de sol et les comparaisons avec les observations, ainsi que par la mise en œuvre d'études de sensibilité. L'influence de l'orographie sur les paramètres de surface est particulièrement mise en évidence. Finalement, une technique variationnelle basée sur de telles informations, est employée pour corriger et améliorer le comportement de la température à 2 m.

La méthode dite "extended force-restore" utilisée pour calculer les flux de chaleur dans le sol, suppose que la valeur des capacité et conductivité thermiques est constante. Outre cette limitation importante, la paramétrisation de la végétation apparaît quelque peu artificielle, la longueur des racines étant dépendante du temps.

Les expériences montrent que l'évapotranspiration par conditions pluvieuses, ainsi que l'amplitude de la température à 2 m, sont surestimés par le modèle. Les comparaisons avec les observations provenant du réseau suisse de mesure automatique et de la station hydrologique à Büel, indiquent que le maximum de la température à 2 m apparaît 1-2 heures plus tôt que dans les observations.

Par ciel dégagé, la fraction de couverture végétale, la profondeur des racines, ainsi que le contenu en eau du sol, contrôlent les paramètres de surface en modifiant le "Bowen-ratio", c.-à-d. le rapport des flux de chaleur latente et sensible. Les études de sensibilité montrent que l'évapotranspiration est particulièrement affectée par les paramètres sus-mentionnés. Par conditions nuageuses et pluvieuses, les flux d'énergie au sol sont faibles, et les conditions de la de couche limite planétaire contrôlent les paramètres de surface.

L'évaporation sur sol nu et le contenu hydrique du sol proche de la surface sont les paramètres déterminants du modèle aux altitudes supérieures à 1500 m, là où la végétation est clairsemée. En contraste, l'analyse de la paramétrisation de l'évapotranspiration montre qu'aux basses altitudes, la transpiration de la végétation et le contenu hydrique des couches profondes du sol gagnent en importance, quoique
l'évaporation sur sol nu reste encore un processus significatif.

Une méthode 2D-VAR d'assimilation variationnelle est introduite pour tenter d'améliorer le comportement de la température à 2 m, en corrigeant l'état initial du contenu hydrique du sol. Une fonction de pénalité, qui mesure l'écart entre la température à 2 m maximale observée et modélisée, doit être minimisée. L'ajout d'un terme supplémentaire, lié à l'état initial du contenu hydrique du sol, s'est révélé nécessaire, afin d'obtenir de cette fonction un minimum distinct. En effet, de nombreuses configurations hydriques du sol peuvent produire des taux d'évapotranspiration similaires. La correction de la température minimale demeure difficile, en raison des limitations du modèle de sol actuel.
Zusammenfassung

In numerischen Wettervorhersagemodellen werden die relevanten physikalischen Prozesse an der Erdoberfläche durch sog. Boden-Vegetation-Atmosphäre Austauschschemen beschrieben. Temperatur und spezifische Feuchte, fühlbare und latente Wärmeflüsse, sowie Impulslüsse stellen die unteren Randbedingungen für die Atmosphäre dar und beeinflussen die bodennahen Variablen erheblich. Die Verbesserung der Prognosen für die 2m-Temperatur und die relative Feuchte, wie sie vom operationellen Modell der Schweizerischen Meteorologischen Anstalt (SMA) geliefert werden, stellen das Hauptthema dieser Dissertation dar.


Die für die Berechnung der Bodenwärmeäusse verwendete sog. „extended force-restore“ Methode nimmt an, dass Wärmekapazität und -leitfähigkeit des Bodens konstant ist, was eine erhebliche Einschränkung darstellt. Hinzu kommt, dass die Vegetationsparameterisierung ein wenig künstlich erscheint, da die Wurzeltiefe von der Zeit abhängig ist.

Experimente machen deutlich, dass in regnerischen Bedingungen sowohl die Evapotranspiration, wie auch die 2m-Temperatur vom Modell überschätzt werden. Vergleiche mit Beobachtungen aus dem automatischen Beobachtungssnetz der SMA und der hydrologischen Station Büel zeigen, dass das 2m-Temperaturmaximum 1-2 Stunden früher als in den Beobachtungen erscheint.

Bei wolkenlosem Himmel werden die bodennahen Variablen von Vegetationsanteil, Wurzeltiefe und Bodenfeuchte beeinflusst, indem sie den „Bowen-ratio“, die Aufteilung zwischen fühlbarem und latentem Wärmefluss, ändern. Sensitivitätsstudien zeigen, dass die Evapotranspiration durch die obengenannten Parameter besonders stark beeinflusst wird. In bewölkten und regnerischen Bedingungen andererseits sind die Energieflüsse an der Oberfläche schwach, und der Zustand der planetaren Grenzschicht steuert die bodennahen Variablen.

Unbewachsener Boden und die Bodenfeuchte an der Oberfläche sind die relevanten Modellparameter auf Höhen über 1500 m, wo die Vegetation spärlich ist. Im Vergleich dazu zeigt die Analyse der Evapotranspirationsparametrisierung, dass mit abnehmender Höhe die Feuchte im tiefen Boden und die Vegetationtranspiration
an Bedeutung zunehmen und dass die Verdunstung vom unbewachsenen Boden ein wichtiger Prozess bleibt.

Es wird versucht, das 2m-Temperaturverhalten durch eine variationelle 2D-VAR Assimilationsmethode verbessern, in der die Anfangsbedingung der Bodenfeuchte geändert wird. Hierzu wird eine Kostfunktion minimiert, die ein Maß für den Unterschied zwischen beobachteter und modellierter 2m-Temperaturmaximum darstellt. Es stellt sich dabei heraus, dass ein Hintergrundfeld, das die "first guess" Bodenfeuchte enthält, notwendig ist, um ein eindeutiges Minimum der Kostfunktion zu erhalten, da viele Bodenfeuchtetemperaturen ähnliche Evapotranspirationsraten produzieren können. Die Korrektur des Temperaturminimums ist schwierig und auf die Beschränkungen des aktuellen Bodenmodells zurückzuführen.
Chapter 1
Introduction

The role of land-surface processes is increasingly being recognized as one of the key elements of weather forecasting and of the climate system. The various interactions at the earth’s surface are depicted in Fig. 1.1. Put simply, the soil-vegetation-atmosphere interface absorbs and transforms the major part of the received solar energy into sensible and latent heat during daytime and loses energy radiatively during night-time. How precipitation is recycled back to the atmosphere is also largely dictated by this interface.

Cloudiness and surface albedo determine the amount of incoming shortwave radiation that is effectively absorbed by the surface. Land properties such as soil types/textures (e.g. rock, sand, loam or peat) and vegetation (e.g. pasture, crops or forests) are decisive for the partitioning of this energy between sensible and latent heat fluxes. Vegetation regulates its transpiration according to external stresses (e.g. soil moisture and solar radiation) and is the key component of the latent heat flux, or evapotranspiration. The amount of precipitable water which can infiltrate the soil, be re-evaporated in the atmosphere and collected by various waterways depends also on the soil/vegetation distribution. During the night, sensible and latent heat fluxes remain weak and the energy transfers are dominated by longwave radiation exchanges. The amount of longwave radiation emitted and received back by the earth’s surface depends respectively on the ground surface temperature and the cloud cover, as well as the effects associated to greenhouse effect gases.

In models, heat and moisture transfers into the soil are represented by simple physical laws, i.e. Fourier’s equation of diffusion and Darcy’s law respectively. On the other hand, processes such as vegetation growth, precipitation interception and especially evapotranspiration can only be parameterized, because the involved physical mechanisms are too complex to be described in a simple way. Approximations, or parameterizations, validated with observations are used instead. The mathematical parameterizations describing the soil-vegetation-atmosphere transfers have been termed SVAT schemes, or soil models, and include all aforementioned processes.

The physical processes at the surface influence crucially the behavior of screen-level variables (i.e. 2m temperature, 2m relative humidity and 10m winds), on which this study is focused. The accuracy of the 2m temperature in particular, is
Fig. 1.1: Interactions at the earth's surface. The main physical processes between soil, vegetation and atmosphere are sketched with different soil/vegetation types. The wind influences the intensity of sensible and latent heat fluxes and is affected by the surface roughness (e.g. vegetation or buildings).

intimately bound to the parameterizations of SVAT schemes.

### 1.1 SVAT Schemes

The relevant aspects in land-surface parameterization were identified by Richardson (1922) in his classical book on numerical weather prediction (NWP). Soil models were not an important issue at the very beginning of NWP, the early computers being limited in both computation power and memory storage. Only crude surface conditions for temperature and moisture were specified instead.

The first soil models were highly simplified. Soil hydrology was described with a so-called “bucket” scheme (Manabe, 1969), in which a near-surface layer of soil is modeled with a bucket that can be filled by precipitation and emptied by evaporation and run-off. Ground surface temperature prediction was performed with the force-restore method, a simple and economical version of the heat diffusion equation (Bhumralkar, 1975). The inclusion of a layer of vegetation to calculate transpiration from plants, interception loss and storage of the canopy marked difference from the earlier models (Deardorff, 1978). Since, numerous SVAT schemes (see e.g. Sellers et al., 1986; Noilhan and Planton, 1989; Ducoudré et al., 1993; Bosilovich and Sun, 1995; Viterbo and Beljaars, 1995) are available to the climate and NWP
1.2 Recent Interests

The accurate parameterization of land-surface processes is vital for the NWP and is currently an active research field (Bougeault, 1997). To improve the understanding of the numerous physical interactions, the Project for Intercomparison of Land-surface Parameterization Schemes (PILPS) was established (Henderson-Sellers et al., 1993, 1995). First results show that the differences between individual models can still be large (Chen and al., 1997).

1.2 Recent Interests

The quality of near-surface parameters goes hand in hand with the development and refinement of soil-vegetation-atmosphere transfer schemes and planetary boundary layer (PBL) parameterizations. Systematic errors found in screen-level variables originate most of the time from a deficiency in SVAT schemes or PBL parameterizations. For instance, surface and 2m temperatures were often significantly overestimated by the HIRLAM model at night. This behavior was related to the poor surface parameterization, especially of the ground heat fluxes and soil diffusivity (Rontu, 1995; Saas and McDonald, 1995; Saas and Järvenoja, 1996). The 2m temperature forecasts of the ECMWF global model were improved noticeably by introducing soil freezing and by increasing turbulent diffusion of heat in the stable PBL (Beljaars et al., 1996a).

Soil moisture inconsistencies are also signaled by screen-level variable drifts and might even affect mesoscale circulation (Ookouchi et al., 1984). For example, the misprediction by the ECMWF global model of anomalous rainfall over United States during July 1993 revealed a model sensitivity to the land-surface parameterization and soil wetness anomalies (Beljaars et al., 1996b). Also, the soil water content has been shown to be of great importance for the soil-precipitation feedback at a regional scale (see e.g. Schär et al., 1988).

The feasibility to retrieve soil moisture fields by assimilating screen-level information with a variational method was demonstrated for the first time by Mahfouf and Noilhan (1991). Similar studies have followed since (see e.g. Calvet et al., 1998; Callies et al., 1998; Rhodin et al., 1998). In global climate models (GCMs), more realistic soil moisture fields might improve the partitioning of sensible and latent heat fluxes, which is particularly problematic during the summertime period (Schulz et al., 1998).

Several studies with heterogeneous soil/vegetation conditions and different flux aggregation techniques have already been conducted (see e.g. Zhong and Doran, 1995; Mölders et al., 1996; Mölders and Raabe, 1996; Hu and Islam, 1998). It has been shown that run-off and evapotranspiration were greater with inclusion of sub-grid variability than without (Ghan et al., 1997). This is of primary concern for the accuracy of the hydrological cycle in GCMs, which have a coarse horizontal spatial resolution (Abramopoulos et al., 1998). Note that sub-grid inhomogeneities are usually not taken into account in NWP models.
1.3 Aims of This Study

The aim of this study is to determine the origins of some systematic errors found in summertime screen-level variables predicted by the Swiss Model (SM) and to seek to improve their forecast quality with help of a variational method. The behavior of screen-level variables with respect to the orography are especially emphasized in this study.

The appropriateness and performance of SVAT scheme formulations in the pre-Alps and Alps are not very well known, since observations are difficult and sparse. Soil, atmosphere and vegetation conditions are also quite different at these altitudes, compared with the well studied lower areas. Field experiments have shown that vegetation still controls the evapotranspiration in mountainous regions (Konzelman et al., 1997). It is also worth noting that the effect of orography on evaporation has been only studied for small hills (Huntingford et al., 1998), but remains largely unknown for mountains. Estimates of the reduction of evapotranspiration with height show a large spread (Lang, 1981).

The overall strategy for deriving better screen-level variables is the following: first, the formulation of the SVAT scheme used in the SM is closely examined for possible limitations (Chap. 2, App. A and B). Its performance is then evaluated with observations and detected errors are analyzed (Chap. 3). Experiments are then carried out to determine to which parameter screen-level variables are the most sensitive (Chap. 4). The evapotranspiration parameterization is also evaluated (Chap. 5). Finally, this information is used to set up and tune the variational method (Chap. 6 and App. C).

This study is restricted to the summer period. In Winter, the snow layer disconnects physically the ground from the atmosphere and reduces thereby the usefulness of a variational method based upon soil parameters only. Moreover, snow is an additional complex medium, which is poorly modeled in NWP models.

1.4 Outline

This thesis is organized as follows. Chapter two is focused on the SVAT scheme of the SM. The components of the thermal and hydrological cycles are briefly introduced, their purpose are explained and their strengths and weaknesses are discussed. Parallels are established with the current parameterizations used in NWP and global climate models. A short description of the SM and a note on the interpolation of screen-level variables are also provided.

The derivation of the force-restore method and the outline of its extension, which is used to compute ground temperature, are given in App. A. Definitions of widely used hydrological terms, tables of soil types/textures thermal and hydrological properties, and descriptions of the albedo and vegetation parameterizations are available in App. B.

The third chapter is devoted to the comparisons of the SM with observations from the Swiss automatic observing network and in situ measurements from the hydrological station Büel. These two data sources are presented, as well as the
validation method used for the standard verification of the SM. More specifically, comparisons of global radiation, precipitation, screen-level variables, evapotranspiration, soil temperature and soil moisture are analyzed for the months June to September 1995/96.

Chapter four deals with the sensitivity of screen-level variables with respect to soil model parameter changes. The accuracy and consistency of the investigation method is discussed. Impacts upon soil-atmosphere energy fluxes and screen-level variables are shown for distinct changes carried out in a small domain comprising Switzerland. More precisely, albedo, roughness length, vegetation, soil temperature and soil moisture changes are investigated for a rainy day (July 8, 1996) and a sunny day (July 22, 1996). The meteorological situation for these two case studies is also available.

Chapter five contains an analysis of the evapotranspiration parameterization as a function of the orography. Several aspects concerning soil moisture changes based on the experiments of the previous chapter are further discussed.

In chapter six, the variational analysis technical framework is used to improve 2m temperature forecasts by providing more adequate soil moisture initial conditions to the model. The variational method is presented and new formulations of the cost function are tested with a simplified one-dimensional version of the SM. The quality of first guess and retrieved soil moisture fields are discussed. The derivation of the variational analysis equation and an illustration of its effective use are given in App. C.

Finally, an outlook is given in the last chapter.
Chapter 2

Model’s Overview

The general framework is set with the brief but concise technical overview of the SM, which is the main tool used in this study. A careful description of the SVAT scheme included in the SM follows and constitutes the base reference material throughout this thesis. This overview of the soil model also provides the opportunity to critically discuss the merits of the implemented parameterizations. This is necessary given that many mathematical formulations, in particular those concerning the modeling of vegetation and evapotranspiration, remain specific to each SVAT scheme. This presentation also emphasizes the intrinsic complexity of the close inter-relationships between soil, vegetation and atmosphere processes, and contributes to a better understanding of their implementation in this particular NWP model.

2.1 The Swiss Model

The SM originated as the Europa Model (EM), a NWP meso-β-scale model developed by the German Weather Service (DWD). Both are limited area models based on the hydrostatic set of primitive equations and operating on a rotated longitude/latitude grid. The SM runs operationally at the Swiss Meteorological Institute (SMI) since September 1994 and provides, twice a day, high resolution weather forecasts up to 48 hours. Initial and hourly boundary conditions are supplied by DWD from their continuous EM-based assimilation cycle and are then interpolated on the SM grid.

The current version of the SM covers central Europe (see Fig. 2.1a) with a horizontal resolution of 0.125° (~ 14 km) and owns 20 terrain-following hybrid vertical layers of upward increasing thickness (145 × 145 × 20 grid-points). Model equations are discretized in finite-differences using a semi-implicit scheme and solved with a time step of 240 s. The parameterized processes include a radiation transfer scheme (Ritter and Geleyn, 1992), Kessler-type cloud microphysics and mass-flux moist convection (Tiedke, 1989), a boundary-layer and turbulence formulation (Mellor and Yamada, 1974), a surface layer formulation (Louis, 1979), a soil model (Jacobsen and Heise, 1982) and a fourth-order horizontal diffusion.

Fig. 2.1: Maps from the Swiss Model. The whole SM domain (a) and the test area (with model grid mesh) used for the sensitivity studies (b) are represented. The topography is shaded every 100 m and contoured at 800 and 1500 m.
2.2 The Soil Model

The parameterization of land surface processes (LSP) in NWP is important for a number of reasons. Temperature and specific humidity, as well as sensible and latent heat fluxes at the soil-vegetation interface constitute lower boundary conditions for the atmosphere. Sensible and latent heat fluxes influence boundary layer exchanges and the intensity of convection. These processes affect the low level cloudiness, which in turn modifies the radiative balance at the interface. This is in effect the main surface-atmosphere feedback mechanism. Furthermore, the land-surface schemes are also largely responsible for the quality of model-produced near-surface weather parameters such as screen-level temperature and dew point (Viterbo, 1995). Finally, the correct partitioning between sensible and latent heat fluxes determines the soil wetness, which acts as one of forcing low frequency atmospheric variability (Delworth and Manabe, 1989).

Snow and freezing processes are handled in the soil model, but are not described in this overview for the reasons mentioned in Chap 1. This also applies to an alternate but optional parameterization of the vegetation transpiration from Dickinson (1984). The first temperature and humidity model levels in the PBL have approximately the following heights: 30, 130 and 300 m. (A full description of the SVAT scheme with snow and freezing parameterization included is available in Schrodin...
2.2.1 The Heat Budget

The soil heat transfer is assumed to obey the following Fourier law of diffusion:

\[
(pC) \frac{\partial T}{\partial t} = \lambda \frac{\partial^2 T}{\partial z^2}
\]  

(2.1)

where \( T \) is the soil temperature (K), \( z \) is the vertical coordinate (distance from the surface, positive downward) in m, \( (pC) \) is the volumetric soil heat capacity (J m\(^{-3}\) K\(^{-1}\)) and \( \lambda \) is the heat conductivity (W m\(^{-1}\) K\(^{-1}\)). The above equation assumes that heat fluxes are predominantly in the vertical direction, the effects of phase changes in the soil and the heat transfer associated to the vertical movement of water in the soil can be neglected (de Vries, 1975), and also the effects of hysteresis (Milly, 1982).

By discretizing the soil in multiple layers with decreasing thicknesses towards the ground surface, Eqn. 2.1 can then be solved with finite differences. The choice for the number of layers and for their spatial arrangement is completely free, which is the main advantage of this method. For convenience, the same soil discretization is usually used for both thermal and hydrological processes. The two-layer approximation of the heat diffusion equation reads:

\[
(pC) \Delta z_B \frac{\partial}{\partial t} \left( \frac{T_B + T_M}{2} \right) = G - G_M
\]

\[
(pC) \Delta z_M \frac{\partial}{\partial t} \left( \frac{T_M + T_U}{2} \right) = G_M - G_U
\]

(2.2)

where \( \Delta z_{B,M} \) are the layer thicknesses (m) and \( G_{M,U} \) are the soil heat fluxes at soil layer boundaries (W m\(^{-2}\)), as depicted in Fig. 2.2. The top boundary condition is the net heat flux \( G \), i.e. sum of the radiative, latent and sensible heat fluxes:

\[
G = (1 - \alpha) S^1 + R^1 - \epsilon \sigma T_B^4 + H + LE_{tot}
\]

(2.3)

where \( \alpha \) is the surface albedo (-), \( S^1 \) and \( R^1 \) are the incoming shortwave and longwave radiation respectively (W m\(^{-2}\)), \( \epsilon \) is the emissivity (-), \( \sigma \) is the Stefan-Boltzmann constant (W m\(^{-2}\) K\(^{-4}\)) and \( L \) is the latent heat of evaporation (J kg\(^{-1}\)). \( H \) and \( E_{tot} \) are the sensible heat flux (W m\(^{-2}\)) and the evapotranspiration (kg m\(^{-2}\) s\(^{-1}\)) respectively. The bottom boundary condition is specified like a heat reservoir with infinite capacity, i.e. \( \partial_t T_U = 0 \).

The fluxes \( G_{M,U} \) are not discretized in finite differences, but are specified with the extended force-restore (EFR) method developed by Jacobsen and Heise (1982). This is an extension of the force-restore (FR) method that is an approximation of Eqn. 2.1 for a single harmonic forcing (see App. A, pp. 91 for the derivation).
2.2 The Soil Model

The EFR-method can simulate the earth surface temperature for two preselected frequencies of harmonic forcing without amplitude or phase errors as compared to the equation of heat conduction and can be shown to be equivalent to Eqs. 2.2 if the fluxes $G_{M,U}$ and the layer thicknesses $\Delta z_{B,M}$ are defined in a consistent manner (see also App. A, pp. 92 for an outline). The depths of the soil layers are close to the e-folding depth of the temperature waves penetration into the ground. For two prescribed external harmonic forcing of periods $\tau_1 = 1$ day and $\tau_2 = 5\tau_1$, the expression of the heat fluxes $G_M$ and $G_U$ yields:

$$G_M = \sqrt{\frac{\lambda (\rho C)}{\tau_1}} \cdot (1.58(T_B - T_U) - 1.28(T_M - T_U))$$

$$G_U = \sqrt{\frac{\lambda (\rho C)}{\tau_1}} \cdot 0.68(T_M - T_U)$$

(2.4)

where $T_B$, $T_M$ and $T_U$ are soil temperatures (see Fig. 2.2). These settings give a reasonably good response of the surface temperature from several hours up to some days. Actually, for this particular choice of $\tau_1$ and $\tau_2$, the harmonic temperature variations of periods $0.25\tau_1 \leq \tau \leq 22\tau_1$ are simulated with at most a 10% amplitude error and 25° phase error (Jacobsen and Heise, 1982).

Several drawbacks arise with the EFR-method. First of all, the soil discretization is limited to two layers and has an extra dependence on the soil type. Furthermore, the temperature $T_B$ is a "skin" temperature completely decoupled from the vegetation, which reacts very quickly to the solar radiation (small thermal inertia). This is not viewed to be problematic by Viterbo and Beljaars (1995), who introduced this concept in the most recent ECMWF soil model.

The volumetric soil heat capacity and soil thermal conductivity are expressed as function of available water-holding capacity $\bar{\theta} = \frac{1}{2} \theta_{ava}$ (the availability $\theta_{ava}$ and porosity $\theta_{sat}$ are hydraulic parameters that are defined in App. B, pp. 98):

$$\rho C = (\rho C)_w + (\rho C)_w \bar{\theta}$$

$$\lambda = \lambda_w + \left(0.25 + \frac{0.3\Delta\lambda}{1 + 0.75\Delta\lambda}\right) \Delta\lambda.$$  

$$\min\left\{\frac{4\bar{\theta}}{\theta_{sat}}, 1 + \left(\frac{4\bar{\theta}}{\theta_{sat}} - 1\right) \cdot \frac{1 + 0.35\Delta\lambda}{1 + 1.95\Delta\lambda}\right\}$$

(2.5)

(2.6)

Note that Eqn. 2.5 is in effect incomplete (i.e. the porosity of the soil is not taken into account) and should read:

$$\rho C_{sat} = (1 - \theta_{sat}) (\rho C)_w + (\rho C)_w \bar{\theta}$$

(2.7)

This means that the heat capacity and the ground heat fluxes are currently overestimated (see Table. 2.1).
soil textures | $(\rho C)_{\text{min}}$ | $(\rho C)$ | $(\rho C)_{\text{std}}$ | $(\rho C)_{\text{max}}$ | $\lambda_{\text{min}}$ | $\lambda$ | $\lambda_{\text{max}}$
--- | --- | --- | --- | --- | --- | --- | ---
sand | 1.28 | 1.78 | 1.31 | 2.80 | 0.30 | 1.45 | 2.40
sandy loam | 1.35 | 2.10 | 1.50 | 3.21 | 0.28 | 1.55 | 2.40
loam | 1.42 | 2.36 | 1.71 | 3.32 | 0.25 | 1.09 | 1.58
clay loam | 1.50 | 2.66 | 1.95 | 3.49 | 0.21 | 1.14 | 1.55
clay | 1.63 | 3.13 | 2.31 | 3.75 | 0.18 | 1.22 | 1.50
peat | 0.58 | 2.73 | 2.23 | 4.19 | 0.06 | 0.35 | 0.50

Table 2.1: Soil thermal properties. $(\rho C)$ is the heat capacity ($10^6$ J K$^{-1}$ m$^{-3}$) and $\lambda$ is the thermal conductivity (W K$^{-1}$ m$^{-1}$). Minimum and maximum values are reached when the soil is dry or saturated with water respectively. $(\rho C)_{\text{std}}$ is the value obtained with the correct formulation of the heat capacity (see Eqn. 2.7).

Finally, note that there is the requirement that $(\rho C)$ and $\lambda$ be constant with time. This has clearly negative impacts, especially when the soil is too dry or saturates after rain events. Standard, minimum and maximum values of $(\rho C)_o$ and $\lambda$, as compared to the fixed current values, are available in Table 2.1. Note that the differences can be large between dry and saturated soils.

2.2.2 The Water Budget

The vertical movement of water in the unsaturated zone of the soil matrix obeys Richardson’s equation:

$$\rho_w \frac{\partial \theta}{\partial t} = -\frac{\partial F}{\partial z} + \rho_w S_\theta$$

(2.8)

where $\rho_w$ is the water density (kg m$^{-3}$), $\theta$ is the volumetric soil water content (m$^3$ m$^{-3}$), $F$ is the vertical water flux in the soil (positive downward, kg m$^{-2}$ s$^{-1}$) and $S_\theta$ is the volumetric source term (m$^3$ m$^{-3}$ s$^{-1}$) corresponding to infiltration, root uptake and ground run-off.

Using Darcy’s law extended to the flow of water in the unsaturated case, $F$ can be specified as:

$$F = -\rho_w \left(D \frac{\partial \theta}{\partial z} - K\right)$$

(2.9)

See Richards (1931) and Philip (1957) for the conditions under which Eqs. 2.8 and 2.9 are valid. Note that Richardson’s and Darcy’s relations are the standard equations used in all soil models. $D$ and $K$ are the hydraulic diffusivity (m$^2$ s$^{-1}$) and hydraulic conductivity (m s$^{-1}$) respectively. They are specified for each layer $k$ as a function of the weighted soil water content $\bar{\theta}_k$:

$$D_k = D_{\text{sat}} \cdot e^{D_{k1}(\theta_{\text{sat}} - \bar{\theta}_k)/\left(\theta_{\text{sat}} - \theta_{\text{dvp}}\right)}$$

$$K_k = K_{\text{sat}} \cdot e^{K_{k1}(\theta_{\text{sat}} - \bar{\theta}_k)/\left(\theta_{\text{sat}} - \theta_{\text{dvp}}\right)}$$

(2.10)
2.2 The Soil Model

where \( \overline{\theta_k} = a \min \{ \theta_k, \theta_{k+1} \} + (1 - a) \max \{ \theta_k, \theta_{k+1} \} \). \( D_{\text{sat}} \) and \( K_{\text{sat}} \) are the hydraulic diffusivity and hydraulic conductivity at saturation respectively. With a coefficient \( a = 0.8 \), the weight is set upon minimum soil moisture values that favor slow moisture exchange between layers (low \( D \) and \( K \) values). This avoids also numerical instabilities related to the highly dynamical nature of hydraulic parameters (coverage of several orders of magnitude). The use of the parametric relations of Clapp and Hornberger (1978) for the formulation of the hydraulic diffusivity and conductivity is a more common practice:

\[
K = K_{\text{sat}} \left( \frac{\theta}{\theta_{\text{sat}}} \right)^{2b+3} \\
D = \frac{bK_{\text{sat}}\psi_{\text{sat}}}{\theta_{\text{sat}}} \left( \frac{\theta}{\theta_{\text{sat}}} \right)^{b+2}
\]

(2.11)

where \( b \) is the coefficient of Clapp and Hornberger defined for each soil texture (Cosby et al., 1984) and \( \psi_{\text{sat}} \) (m of water) is the so-called “matric” potential at saturation, representing the work required to extract water from the soil against capillarity and gravity forces. These parameters can also be used to describe the thermal conductivity:

\[
\lambda = \frac{a}{\psi_{\text{sat}} \log_{10} \left( \frac{\theta}{\theta_{\text{sat}}} \right)}^{b/\log_{10} 10}
\]

(2.12)

where the coefficient \( a = 3.8 \). This formulation, following McCumber and Pielke (1981), is used in recent soil models and forms a consistent set with the previous equations. However, this is not implemented in the SM. Finally, the water budget equations are:

\[
\rho_w \Delta z_k \frac{\partial \theta_k}{\partial t} = I + (1 - \sigma_W)(E_h + E_{v_k}) - F_2 - R_1 \\
\rho_w \Delta z_k \frac{\partial \theta_k}{\partial t} = (1 - \sigma_W)E_{v_k} + F_k - F_{k+1} - R_k \quad k = 2, ..., N
\]

(2.13)

where \( \Delta z_k \) is the soil layer thickness (m) and \( \sigma_W \) is the interception reservoir cover fraction (m² m⁻²), which is defined later (see Eqn. 2.30). \( I \) is the infiltration rate, \( E_h \) and \( E_{v_k} \) are the evaporation rate from bare soil and vegetation respectively, \( R_k \) is the run-off and \( F_k \) is the moist flux between layers \( k \) and \( k - 1 \). All above defined fluxes have units of kg m⁻² s⁻¹.

When an active layer becomes too dry and its soil moisture value drops below the air dryness point, the fluxes are corrected in order to set \( \theta_k \) to the minimum moisture value \( \theta_{\text{adp}} \) at the next time step. The top boundary condition is precipitation minus evaporation and the bottom boundary condition is specified like a water reservoir with infinite capacity, i.e. \( \partial_t \theta_3 = 0 \), and can be interpreted as a no drainage condition (the so-called “bedrock” condition).
Run-off

Ground run-off is triggered when $\Delta F_k = F_k - F_{k+1} > 0$ and $\theta_k > \theta_{fc}$:

$$R_k = \frac{\theta_k - \theta_{fc}}{\theta_{sat} - \theta_{fc}} \Delta F_k$$  \hfill (2.14)

This parameterization is an attempt to simulate a minimal drainage, which plays an important role in hydrology, but this is not the same as if free drainage was really specified as the bottom boundary condition. The introduction of gravitational drainage greatly improved the simulation of the annual cycle of soil moisture (Mahfouf and Noilhan, 1996). PILPS results show also that run-off have a non-negligible impact on the annual soil moisture cycle (Koster and Milly, 1997).

2.2.3 Sensible and Latent Heat Fluxes

The parameterization scheme for sensible and latent fluxes is based on the Monin-Obukhov similarity theory and is well supported by experimental evidence (Businger et al., 1971). A review of flux-profile relationships is also available in Dyer (1974). The sensible heat flux can be written as:

$$H = \rho_h C_p c_h |u_h|(T_h - T_B)$$  \hfill (2.15)

where $\rho_h$ is the air density (kg m$^{-3}$), $C_p$ is the heat capacity of air at constant pressure (m$^2$ s$^{-2}$ K$^{-1}$), $c_h$ is the heat transfer coefficient (-), $|u_h|$ is the wind speed at height $h$ (m s$^{-1}$) and $T_B$ is the surface temperature (K). To assure consistency, the original temperature $T_h$ in Eqn. 2.15 is replaced by the virtual potential temperature in the SM.

The potential evaporation flux $E_{pot}$ is the amount of water evaporated per unit of area and per unit of time from an idealized extensive free water surface under existing atmospheric conditions (Shuttleworth, 1993):

$$E_{pot} = \rho_h c_H |u_h|(q_h - q_{sat}(T_B, p_s))$$  \hfill (2.16)

where $q$ is the specific humidity (gr gr$^{-1}$) and $p_s$ is the surface pressure (Pa). The transfer coefficient is identical for heat and moisture fluxes and is calculated following Louis (1979):

$$c_H = \frac{\kappa^2}{\ln\left(\frac{h}{z_o}\right) \ln\left(\frac{h}{z_H}\right)} F\left(R_{ib}, \frac{h}{z_o}, \frac{h}{z_H}\right)$$  \hfill (2.17)

where $\kappa$ is the von Karman constant (-), $z_o$ and $z_H$ are the roughness lengths (m) for momentum and heat respectively, $R_{ib}$ is the bulk Richardson number (-) and $F$ are the stability functions specified in ECMWF (1991). The atmosphere is
considered as unstable, neutral or stable for the respective conditions: $Ri_b < 0$, $Ri_b = 0$ and $Ri_b > 0$.

Instead of using this simple formulation in terms of the Richardson’s number, one may use transfer coefficients expressed as a function of the Obukhov length (Beljaars and Holtslag, 1991). The advantage is that the empirical stability functions can be specified as they have been measured and the surface roughness lengths for momentum, heat and moisture can be chosen independently. Nevertheless, an iterative process to convert the Richardson number into an Obukhov length is required at each time step.

A thorough theoretical review of PBL processes can be found in Stull (1988).

**Bare Soil Evaporation**

Bare soil evaporation results from a combination of two physical processes, i.e. molecular diffusion from the water trapped in the pores of the soil matrix up to the land surface level and laminar exchanges in the air between the roughness length $z_H$ and screen-level height.

Modeling the evaporation as a bulk transfer of water vapor between $z_H$ and the lowest atmospheric model level is done following $\alpha$ or $\beta$-methods. $\alpha$-methods regulate the evaporation by modulating the saturated specific humidity at the surface:

$$E_h = (1 - \sigma)\rho_h c_H |u_h| (q_h - \alpha q_{sat}(T_B, p_h))$$

(2.18)

where $\sigma$ is the vegetation cover fraction (m² m⁻²). In comparison, $\beta$-methods control the specific humidity gradient:

$$E_h = (1 - \sigma)\beta_b^2 F_{pot}$$

(2.19)

where $(1 - \sigma)$ is the bare ground fraction. This latter formulation is used to parameterize bare soil evaporation and vegetation transpiration in the SM. Note that $E_h$ is linear with respect to potential evaporation in $\beta$-methods. The coefficient $\beta_b$ is defined as:

$$\beta_b = \begin{cases} 0 & \theta_1 \leq \theta_{adp} \\ \frac{(\theta_1 - \theta_{adp})}{(\theta_f - \theta_{adp})} & \theta_{adp} < \theta_1 < \theta_f \\ 1 & \theta_1 \geq \theta_f \end{cases}$$

(2.20)

Mahfouf and Noilhan (1991) and Mihailović et al. (1995) made comparative studies of several formulations of evaporation over bare ground. These methods are classified in bulk parameterization approaches ($\alpha$ and $\beta$-types, following Kondo and Saigusa (1990)) and threshold methods. Nevertheless, the choice of the most suitable formulation is still unclear, despite all available studies on that subject.
Transpiration

A similar $\beta$-method is used to compute vegetation transpiration. An additional coefficient $r_{vk}$ depending on the rooting depth provides a selective control of the root uptake in each layer $k$:

$$E_{v_k} = \sigma r_{v_k} \beta_{v_k}^2 E_{pot} \quad k = 1, \ldots, N$$

with

$$\beta_{v_k} = \begin{cases} 
0 & \theta_k \leq \theta_{wp} \\
(\theta_k - \theta_{wp})/(\theta_{tp} - \theta_{wp}) & \theta_{wp} < \theta_k < \theta_{tp} \\
1 & \theta_k \geq \theta_{tp}
\end{cases}$$

and

$$r_{v_k} = \begin{cases} 
\frac{d_r}{\sum_{n=1}^{N} \Delta z_n} & k = 1 \\
\frac{(d_r - \sum_{n=k}^{N-1} \Delta z_n) / \sum_{n=1}^{N} \Delta z_n}{k > 1}
\end{cases}$$

where $d_r$ is the rooting depth in m. The values of $r_{v_k}$ are limited to the range $0 < r_{v_k} < \Delta z_k$. The turgor-loss point $\theta_{tp}$ is the soil moisture value above which vegetation is unstressed:

$$\theta_{tp} = \theta_{wp} + (\theta_{fc} - \theta_{wp})(0.81 + 0.21 \cdot \arctan(-86400 E_{pot} - 4.75))$$

This parameter has the same function as $\theta_{fc}$ for bare ground evaporation and is only used in the SM.

The most obvious drawback of the current vegetation parameterization (see App. B, pp. 96) is the temporal dependence of the rooting depth. More physically sound approaches use a fixed rooting depth or better, a root density as described in Viterbo and Beljaars (1995). Note that root weighting has considerable impact on the computed transpiration (Desborough, 1997). The growth and decay of the vegetation is then simulated with the leaf area index (LAI), which is defined as the total leaf surface area per square meter. Moreover, the possibility exists to retrieve this parameter with remote-sensing methods, and this certainly constitutes an advantage.

Recent models use the concept of stomatal resistance to mimic the vegetation transpiration. The stomatal resistance $r_s$ (s m$^{-1}$) depends not only on soil moisture availability ($F_2$), but also on solar radiation ($F_1$), vegetation temperature ($F_3$) and vapor pressure deficit ($F_4$), and can be expressed as:

$$r_s = \frac{r_{smin}}{LAI} F_1 F_2^{-1} F_3^{-1} F_4^{-1}$$
where $r_{s\text{min}}$ is the minimal stomatal resistance, which depends on plant species. Hence, vegetation is able to respond dynamically to the external forcing and to adjust its transpiration accordingly. See Niyogi and Raman (1997) for a comparative study and several examples of $F$ functions.

The Penman-Monteith formulation is another way of computing evapotranspiration (see e.g. Penman, 1948; Monteith, 1965; De Bruin and Holtslag, 1982) and is principally used by hydrologists:

$$E_{tot} = \frac{\rho_h C_p (q_h - q_{sat}(T_h)) + r_a s_{cc}(T_h)(G - R_n)}{C_p r_a + r_s + L r_a s_{cc}(T_h)}$$

(2.26)

where $R_n$ and $G$ are the net radiative and ground heat fluxes respectively ($W m^{-2}$), $s_{cc}(T_h) = \frac{dT}{d q_{sat}|_{T=T_h}}$ is the slope of the saturation specific humidity function ($K^{-1}$) and $r_a^{-1} = c_H |u_h|$ is the aerodynamical resistance ($s m^{-1}$). This equation is obtained by eliminating the surface temperature variable $T_B$ with a linearization procedure to avoid the imprecision associated to its measure.

**Surface Humidity**

The total surface moisture flux is given by:

$$E_{tot} = (1 - \sigma_W)(E_h + \sum_{k=1}^{N} E_{v_k}) + \sigma_W E_{pot}$$

(2.27)

and leads according to Eqn. 2.16 to the “fictive” surface humidity $\tilde{q}_{sat}$, which is used as a boundary condition for further model computation (diffusion and convection processes) and ensures the conservation of moisture:

$$\tilde{q}_{sat} = q_h - \frac{E_{tot}}{\rho c_H |u_h|}$$

(2.28)

This constitutes a model limitation (i.e. $\tilde{q}_{sat}$ is different from the real value) and should be removed to allow the use of the real surface humidity (e.g. to compute the 2m relative humidity). However, estimating the real surface specific humidity remains difficult (Lee and Pielke, 1992).

### 2.2.4 The Interception Reservoir

The interception reservoir is a thin layer on top of the soil/vegetation, collecting water by interception of rain and collection of dew, and evaporating at potential rate. It obeys to the following prognostic equation:

$$\rho_w \frac{\partial W}{\partial t} = \gamma P + \sigma_W E_{pot} - T - R_W$$

(2.29)
where $W$ is the capacity of the interception reservoir in m. $P$, $T$ and $R_W$ are respectively the precipitation, throughfall and run-off fluxes in kg m$^{-2}$ s$^{-1}$. Run-off occurs when the interception reservoir is full. $\sigma_W$ is the cover fraction of the interception reservoir and is computed as follows:

$$\sigma_W = \max \left\{0.01, 1 - e^{-4 \cdot 10^{-4} W} \right\}$$

(2.30)

The upper limit for the interception reservoir depends on the vegetation cover fraction and is given by:

$$W_{\text{max}} = 5 \cdot 10^{-4} (1 + 5 \sigma)$$

(2.31)

Interception is computed as the fraction $\gamma$ of the precipitation flux that can be stored in the reservoir, the remaining part being throughfall:

$$\gamma = \sqrt{1 - \frac{W}{W_{\text{max}}}}$$

(2.32)

The parameter $\gamma$ is modified during precipitation events ($P > 0$) to avoid a fall in the reservoir level:

$$\tilde{\gamma} = \max \left\{ \gamma, \frac{(W_{\text{max}} - W') \rho_w}{P} + T \right\}$$

(2.33)

$W'$ and $T$ are defined as:

$$W' = W - \frac{2 \Delta t}{\rho_w} \sigma W E_{\text{pot}}$$

$$T = \frac{\rho_w}{\tau_w} W'$$

(2.34)

where $\tau_w = 1000$ s is a time constant controlling the rate of throughfall (interception reservoir loss). The concept of interception reservoir was originally proposed by Rutter et al. (1971) and has been generalized since (Massman, 1980). For more complicated formulations still based on the Rutter concept, see for instance Mahfouf and Jacquemin (1989); Dolman and Gregory (1992).

The soil surface infiltration $I$ is limited to $I_{\text{max}}$ according to a simplified Holtan-equation:

$$I_{\text{max}} = I_{\text{min}} + 0.002 \cdot \frac{\theta_{\text{sat}} - \theta_1}{\theta_{\text{sat}}} \max \left\{ \frac{1}{2}, \sigma \right\}$$

$$I = \min\left\{ I_{\text{max}}, T + (1 - \gamma)P \right\}$$

(2.35)

Surface run-off occurs when the soil is not able to absorb the entire incoming water flux ($T + (1 - \gamma)P > I_{\text{max}}$). Note also that the maximal infiltration rate can be derived directly from Darcy’s equation (Mahrt and Pan, 1984).
2.3 Screen-level Interpolation

Screen-level variables at 2 m (temperature and relative humidity) and 10 m (wind) are interpolated following Geleyn (1988). This method uses the ground surface and first model level (~30 m) values, as well as different transfer functions depending on the stability of the atmosphere. This interpolation procedure is known to be questionable for very stable atmospheric conditions (especially during the night).

An alternate method that uses directly the Obukhov length was proposed to circumvent this problem (Hess et al., 1995; Hess and McAvaney, 1997). Unfortunately, this scheme requires an iterative process, which takes more time to complete. This procedure was adopted by the Atmospheric Model Intercomparison Project (AMIP) as the standard for computing near-surface parameters.

Note that the current relative humidity at 2m provided by the SM is computed with a simplified interpolation method and thus, cannot be considered as reliable.
Chapter 3

Comparisons With Observations

Observation and data of diverse kinds (e.g. upper-air soundings, surface stations and satellites) are essential to NWP models for their initialization and validation. Analysis methods (e.g. optimal interpolation, 3/4D-VAR or nudging) are used to process this data in association with a forecast model to propagate the information in time and space. For example, the German Weather Service (DWD) uses the EM for its data assimilation. In principle, the assimilation cycle should supply the model with initial conditions that are as close as possible to the current state of the atmosphere. Validation procedures seek to: assess the performance of a model; evaluate objectively the quality of its forecasts; and reveal intrinsic weaknesses. The gathered knowledge is seminal for improving model parameterizations.

The DWD validates its different models over Europe for the standard set of meteorological parameters, i.e. 500 hPa geopotential, surface pressure, cloudiness, 10m wind and 2m temperature. The results are then published regularly in quarterly reports (Schrodin and al., 1988). At SMI, the SM is validated on a monthly basis, but only for the domain inside the boundaries of Switzerland. Observations derived from the Swiss automatic observing network (ANETZ) are used for this purpose. Comparisons of cloudiness, precipitation, 2m temperature, 10m wind velocity and wind direction are carried out and summarized in the form of monthly means, scores and contingency tables (Schubiger, 1995).

During the last decade, numerous field campaigns were undertaken in several regions to provide long time series of high quality detailed surface observations (Andre et al., 1986; Sellers et al., 1988; Bolle and al., 1993; Beljaars and Bosveld, 1997). Progress was achieved in the development and the understanding of SVAT’s operation from these datasets. For example, observations from the Cabaw (Netherlands) and FIFE (Amazonia) experiments were used to calibrate and validate the new ECMWF model (Viterbo, 1995). Henderson-Sellers et al. (1993) made extensive use of the Cabaw dataset for the Project of Intercomparison of Land-surface Parameterizations Schemes (PILPS). These two specific examples illustrate the importance of more detailed observations in meteorology and in particular for SVAT modeling.

In this study, the SM is compared with data from the ANETZ network and additional observations such as soil moisture and evapotranspiration, only available from the hydrological station Büel, for the months of June through to September.
The Swiss Automatic Observing Network - ANETZ

FIG. 3.1: Geographical distribution of the 72 ANETZ stations. Classified according to the altitude, 39 sites are located below 800 m (□), 13 between 800 and 1500 m (●) and 20 above 1500 m (△). The stations measuring potential evaporation are underlined. The position of the hydrological station Büel is also reported on this map.
3.1 The Swiss Automatic Observing Network

The ANETZ is composed of 72 meteorological stations spread throughout Switzerland (see Fig. 3.1). This network was developed in the early seventies by SMI and all stations have now been in operation for more than 15 years. More than half of the stations are located below 800 m (39) and a quarter at an altitude above 1500 m (20). Classified according to altitude, the stations appear not to be distributed in an homogeneous fashion. Note also the visibly stronger spatial concentration in the northern part of Switzerland. A broad range of parameters are measured with a sampling rate of 10 minutes, but with some exceptions (e.g. potential evaporation is only available hourly). Note also that a limited number of stations are equipped to measure only soil temperature at various depths (i.e. 5, 10, 20, 50 and 100 cm) and potential evaporation.

The stations at the Jungfraujoch and the Corvatsch (altitude above 3000 m) were discarded for the comparisons due to their special conditions (eternal snow). Note also that some other alpine stations (e.g. Säntis, Pilatus) are located at mountain tops and their observations may be less representative of their surroundings, in comparison with middle-land stations.

3.2 The Hydrological Station Büel

The hydrological station Büel, with several other automatic stations (e.g. Egghof, Sack, Stadel, Südhang and Bühl), belongs to the Rüetholzach experiment field. This area is located in the low Toggenburg (see Fig. 3.1), across both Mosnang and Gäwhil (SG) communes and forms a part of the Thur river system. Its surface is approximatively 3.2 km² and the altitude of the different subdomains is located between 680 and 950 m. The surface soil structure is composed mostly of a loamy, carbon-rich soil type resulting from the weathering processes. The high precipitation of this region often leads to the formation of mineral wet soils, especially in the flatter undulating parts. The land partition of the domain is representative of a large part of the Swiss pre-Alps and is composed of 25% forests with the remainder comprising pasture lands.

The annual amount of precipitation and the 7° C yearly mean temperature characterize the cold, sub-mountainous climate for this region. The general increase of precipitation with height and its location northwest of the Alps provides a yearly precipitation sum of 1500 mm, of which 53% falls during the summer period. Over the year, the observed value for run-off and total evaporation are around 1000 and 500 mm respectively. At the main climate station Büel (755 m), the 20 year mean measured evapotranspiration is 545 mm (Menzel, 1997).

Global radiation, precipitation, 2m temperature, 2m relative humidity and 2.5m wind data are available hourly. Beside these upper-soil variables, the soil temperature is measured at depths of 20 and 40 cm and the soil moisture at 5, 15, 25, 35, 55
Comparisons With Observations

and 80 cm deep as well as for the whole layer between 80–110 cm. Soil moisture values are obtained indirectly from the raw dielectric soil capacitance measure, which is then converted to a more convenient form. This procedure require a priori accurate calibration (Menzel, 1995). This station also has a lysimeter, from which evapotranspiration values can be deduced. (A lysimeter is simply a stand-alone block of soil and vegetation, whose soil water content is inferred from the continuous monitoring of its weight. Lost water (drainage) is also collected and with the knowledge of the precipitation rate, they give together the evapotranspiration from a simple balance equation.)

3.3 The Validation Method

Validation of model surface parameters in the Swiss terrain is a challenging task. The orography of the Alps, as represented by the SM (see Fig. 2.1, pp. 8), is much too coarse at the current resolution (~ 14 km) to catch the fine orographic details of this mountainous region. Accordingly, the SM may fail to simulate particular the meteorological conditions that a specific ANETZ station may be subject to. For instance, the appearance of valley winds or the shadowing of surrounding mountains might have non negligible effects on the local observing conditions. Other inherent model limitations will be discussed later. The inner-alpine station network is also less dense and thereby hinders the model validation in this area.

At SMI, the validation method applied to the SM is as follows. For each ANETZ station, a single model grid-point is picked for the comparison. First, the four nearest grid-points are identified and then, the one with the minimal height difference with the station is retained as the corresponding model reference point if the differences are greater than 100 m, otherwise the nearest (horizontal distance) is retained. Hence, with this simple procedure, the geographically nearest point is not automatically selected. Only 15 peers, mostly located on the Swiss Plateau, have an altitude difference lower than 100 m. Indeed, the criterion results in the selection of numerous peer points with values larger than a few hundred meters, and indeed in the alpine region, it is not unusual to encounter differences between 500 and 1000 m. Säntis is the least favorable case, where the model peer point is 1250 m higher. Note that taking a more sophisticated approach, such as an interpolation between the four nearest grid-points is not able to improve this situation. These considerations demonstrate clearly one of the difficulties that complicate the validation and the interpretation of the results.

A key question is the reduction of model parameters values to the altitude of the observations. For the usual standard verification, a correction is applied to the 2m temperature only, following the linear relationship:

\[ \bar{T}_{2m} = T_{2m} + \gamma_{T_{2m}} \cdot \Delta h \]  

(3.1)

where \( \gamma_{T_{2m}} = 6 \text{ K km}^{-1} \) is assumed to be the 2m temperature gradient and \( \Delta h = h_{SM} - h_{ANETZ} \) is the height difference between the ANETZ station and its
3.3 The Validation Method

Fig. 3.2: The 2m temperature versus height: gradient (a & b), amplitude (c & d) and time of the maxima (e & f) for July 1996. Uncorrected data (left) and corrected data (right) are used. Regression lines are drawn for ANETZ (solid line) and SM peer points (dotted line).
corresponding model grid-point. This correction is simply based on the linear temperature behavior of the standard atmosphere in the troposphere (γ = 6.5 K km⁻¹). An example of the corrected and uncorrected 2m temperature gradient for July 1996 is shown in Fig. 3.2a,b. The applied correction has the effect of moving grid-points to lower altitudes, but the original gradient appear to be conserved. The SM values use the grid-point height in Fig. 3.2a and the altitude of the ANETZ station in Fig. 3.2b. In this last figure, the redistribution of the SM points is clearly visible, especially in the pre-Alps. The overall effect upon the mean amplitude (Fig. 3.2c,d) and the mean time at which the 2m temperature maximum occurs (Fig. 3.2e,f), is also presented. Note that the SM is not able to provide weak 2m temperature amplitude at high altitude and the 2m temperature maxima always appear 1–2 hours before the observations, which remains difficult to explain.

A reference describing standard methods to be applied in such cases is unfortunately not available. The validation scheme used at the SMI is based on a note published by Hall (1987), which contains only general recommendations, i.e. for the meteorological variables to be compared, the verification time/periods and the statistical methods to use. Note that meteorological institutes running and validating a NWP model are not committed to following a particular validation procedure. Validation procedures at SMI diagnosed a warm bias of the 2m temperature in the SM as well as a regular underestimation of summertime cloudiness. Global radiation, 2m relative humidity, soil temperature and potential evaporation are not part of the standard validation procedure, but are used in the present study to investigate the systematic 2m temperature bias detected through the routine verification.

3.4 Comparisons With ANETZ

Three domains are considered for the comparisons: the Swiss Plateau (h < 800 m), the pre-Alps (800 ≤ h < 1500 m) and the Alps (h ≥ 1500 m). This partition according to the altitude is currently used for the standard verification at SMI to distinguish these important sub-regions in Switzerland. SM data are provided exclusively by the model integrations starting at 12 h UTC, forecast time window ranging from +12 to +36 h UTC.

3.4.1 Shortwave Radiation

During daytime, shortwave radiation is the main source of energy and induces large perturbations in the soil-atmosphere system. The ANETZ network measure continuously the global (i.e. sum of the direct and diffuse) radiation, but the determination of the amount of shortwave radiation effectively absorbed by the ground requires knowledge of the albedo, which is unfortunately not measured.

Comparisons show that for both the selected summers, the model overestimates the shortwave radiation reaching the land-surface. This is probably connected with the misprediction of cloudiness recent study (c.f. the recent study of Gall, 1998). The amount of incident shortwave radiation is higher at alpine stations, primarily due to the reduced atmosphere thickness. However, note also that some alpine
3.4 Comparisons With ANETZ

Fig. 3.3: Mean screen-level variables curves for July and August 1996. Observations (solid lines) and model (dotted lines) are displayed simultaneously for the three usual sub-regions.

stations might be affected by the localized formation of clouds at the mountain tops in convective conditions in Summer. The discussion will now be continued with reference to the Büel comparison.

3.4.2 Screen-level Variables

Results of the screen-level validation are displayed in Fig. 3.3 for two selected months (July and August), although the following discussion also applies without any significant difference to other summer months. Observed screen-level variables are measured at almost all stations (65), which is sufficient to study the effect of orography on these parameters.

Fig. 3.3a,b show that the amplitude of the 2m temperature $T_{2m}$ is overestimated in all regions, especially in the third sub-region (above 1500 m). The observations show a decrease of the amplitude with the altitude, but this is not well reproduced by the model. The forecast of the morning minimum match creditably the observations
below 800 m, but tend to be colder with the increase of height. The temperature maxima appears always 1–2 hours earlier in the model, whatever the region considered. Moreover, for both model and observations, the lower the altitude is, the later this maxima occurs. The conclusions are the same for July 22, 1996 (Fig. 3.4b). It is not an effect related to the spread with longitude of the ANETZ stations (see Fig. 3.1), as the temperatures begin to raise simultaneously and the time lag between the sunrise at the eastern and western extremities of Switzerland is only 20 minutes.

Surface air temperature takes more time to warm above 1500 m, although the air density is less at these altitudes, and therefore should be more responsive to the warming of the soil, and this suggests that it is not a dominant factor. Alpine stations are probably sensitive to the thermally induced air currents which are cooled adiabatically as they gain height and traverse the stations. The smaller temperature amplitude at elevated stations is probable, since they closer represent free tropospheric conditions than low-level ones.
3.4 Comparisons With ANETZ

Fig. 3.5: Air and soil temperatures (top panels) for the ANETZ stations (a), and SM (b). A combination of observed relative humidity, precipitation (c & d), evapotranspiration (c) and potential evaporation (d). Only stations below 800 m (18), measuring soil temperature and potential evaporation simultaneously were considered.

The data for the 2m relative humidity ($RH_{2m}$) is also informative (see Fig. 3.3c,d). The model fails completely to predict it at night-time, as its value is derived from the value of the first model level ($\sim 30$ m). The observations show that $RH_{2m}$ values are similar during the night, whatever the altitude. The atmosphere is at that time stable, and the water evaporated remains close to the surface, increasing mean values of $RH_{2m}$ to 85–90%. The daylight depletion of humidity is more pronounced and occurs later at low altitude too. This is not surprising, as $RH_{2m}$ depends on the temperature. The timing problems of the model are the same as for $T_{2m}$. The observed amplitudes are more differentiated with altitude than for the temperature. The conclusions are the same for July 22, 1996 (Fig. 3.4d).

For the rainy day July 8, 1996, $T_{2m}$ and $RH_{2m}$ modeled values are closer to the observations, because the atmospheric influences predominate over the soil that day.

Fig. 3.5 show that the air and soil model temperature are correlated. $T_{2m}$ and
Comparisons With Observations

$T_M$ have a correlation coefficient greater than 0.9. The observed soil temperature is more decoupled from the 2m temperature and is also warmer (1–3 K). It seems contradictory that the soil temperature is colder in the model, but that the 2m model temperature, on the contrary, is warmer than the observations. The observed 2m relative humidity is also highly correlated with potential evaporation ($R^2 = 0.93$). This is not the case in the model with the evapotranspiration ($R^2 \approx 0$). The observed 2m temperature is moderately correlated with the evaporation ($R^2 = -0.66$). It is also true in the SM, but to a lesser extent ($R^2 = -0.55$).

3.4.3 Potential Evaporation

Potential evaporation (PE) is available for a subset of 25 ANETZ stations (see Fig. 3.1). For most of them, the PE measuring apparatus are activated for less than 6 months of the year: from April to September (5) and from June to September (15). Only 5 stations observe this variable throughout the year. These stations are mostly located below 800 m (19) and distributed predominantly along the Swiss middle-land and the Graubünden. Davos (1590 m) and Samedan (1705 m) are the only ones above 1500 m. (Visp is known to exhibit unrealistic evaporation values that are double, compared to the other stations and was therefore discarded for the comparisons. This behavior can be explained by the strong winds in this region that might blow small quantities of water off the instrument, thus inducing wrong values of potential evaporation.)

A statistical evaluation for a decadal observational period (1984–1993) shows a strong correlation between the stations of the Swiss middle-land, in conformity with their similar meteorological conditions (Brändli, 1996). However note that the evaporation in Reckenholz is for instance, about 30 % larger than in Zürich, although the two stations are only 10 km away from each other. A possible explanation for such a discrepancy is the following: the site of Zürich is surrounded by buildings and vegetation, protecting it from the wind, and therefore reducing the evaporation. Nevertheless the challenge of deriving accurate PE measurements is formidable (Schrödter, 1985).

Monthly PE means from 18 ANETZ stations below 800 m (Figs. 3.6 a,c), and evapotranspiration means (Figs. 3.6b,d) from the SM, are displayed from June to September, for 1995 and 1996. Although it is not possible to compare directly the absolute values, the PE from ANETZ stations might be a good approximation of the evapotranspiration maximum. This is difficult to assess, as the instrument is at a height of 2 m and is protected from the direct solar radiation. In the model, if the water content is above field capacity, bare soil can evaporate at potential rate, whereas vegetated areas are limited to $0.7E_{pot}$, for a rooting depth set to its maximum (0.7 m). This lower value takes into account the additional resistance of the plants to evaporation.

The mean model evapotranspiration is always greater than the mean observed PE, irrespective of the month or the year. In 1995, the ratio evaporation/PE ranged from $\sim 1.3$ (July, August) to $\sim 1.7$ (June, September). In 1996, they were lower, between 1.1 and 1.4. The model evapotranspiration is largely overestimated. Note
3.4 Comparisons With ANETZ

Fig. 3.6: Mean PE measured by 18 ANETZ stations below 800 m (a & c), between June and September 1995/96. The station of Visp was discarded. The mean evapotranspiration of the SM for the corresponding grid-points (b & d) are shown for the same periods. Note that the model data are not corrected to the height of the ANETZ stations.

that the night-time model fluxes are close to zero, whereas a minimal PE is still observed. This is linked to the stability of the atmosphere, which is very stable in the model, giving low values for the transfer coefficient $c_H$. As a consequence, the evapotranspiration fluxes are quasi non-existent during the night. These comparisons show also that the mean model maximum occurs always at 13 h UTC, whereas the observed maxima appears later, between 14 and 15 h UTC. The model curves are temporally quite symmetrical, whereas an asymmetry is clearly visible in the observations.

The Fig. 3.7a,b show the behavior of evaporation for two selected rainy/sunny days. During the sunny day, the modeled evapotranspiration is in good agreement with PE and is slightly lower during the rainy day, whereas the observed PE is reduced drastically to the third for the same period. The origin of the evapotranspiration excess can be understood by noting that the model simulated poorly the
evapotranspiration during June 1995 (Fig. 3.6b), a particularly rainy month (c.f. the analysis). This can be seen even clearer in Fig. 3.5c,d, where PE is reduced during precipitation, whereas evapotranspiration is not. Again, the case of July 8, 1996 is interesting: the surface receives sensible heat from the atmosphere and uses this energy to sustain the night-time evapotranspiration. This behavior is not observed by the ANETZ stations and thereby exposes a model deficiency in the PBL structure.

The distribution of evaporation as a function of the ANETZ station height for two days, July 8 (a rainy day) and July 22 (a sunny day), is shown in Figs. 3.7c, d. On July 8, the model evaporates globally more than the ANETZ stations. This situation is reversed on July 22, showing more realistic values for the evapotranspiration. It is unfortunately not possible to draw firm conclusions about the dependence of the model error upon altitude, since the number of available points is too small. Nevertheless, note the decrease of evaporation with increasing height.
3.4 Comparisons With ANETZ

3.4.4 Soil Temperatures

Observed soil temperatures are generally not taken into account in the assimilation cycles of NWP models. In the ECMWF model, the land-surface temperatures results from the heating and cooling provided by the atmospheric model during first guess computations (Beljaars et al., 1996a). The ability of the soil model packages to simulate ground temperature is not well known, as it depends on local soil and vegetation conditions. Ground temperatures are not part of the common set of parameters usually validated.

The surface temperature $T_B$ determines not only the amount of longwave radiation emitted by the ground, but the stability of the surface layer of the atmosphere (via the transfer coefficient $c_B$), the sensible heat flux, and finally the 2m temperature. $T_B$ is a so-called "skin" temperature, and by design of the EFR-method, reacts quickly to the solar forcing. The middle soil and deep soil temperatures $T_M$ and $T_U$ do not have such a close interaction with the atmosphere in comparison to $T_B$, but they influence the relaxation behavior of the 2m temperature and its morning minimum.

In Fig. 3.8a, the observed temperature at 5 cm ($T_{+5cm}$) is compared with $T_B$, for the month of August 1996. $T_{+5cm}$ can be approximately considered as a skin temperature, and its amplitude is greater than $T_B$, whereas its phase shift with the altitude is barely noticeable (e.g. $T_B$). However, comparison of these two temperatures may be questionable. The middle soil temperatures respond to the solar radiation with a time lag of several hours (Fig. 3.8b), which is typical of a 10 cm deep temperature. The incomplete formulation of Eqn. 2.5, pp. 11, which lead to higher soil heat capacities might account for the colder soil model temperature $T_M$ (2–3 C).

Unfortunately, no soil analyses were undertaken during the setting of the ANETZ network. Hence, there is no soil information available. (Payerne excepted, where soil type, porosity, heat capacity and heat conductivity values were determined recently.
Comparisons With Observations

in laboratory, Mühlemann (see 1996).

3.5 Comparisons With Büel

Note that the measure of evapotranspiration with a lysimeter is limited during rainy conditions. Actually, the observed values are of the same order of magnitude as the instrumental error. Accordingly, the hourly evapotranspiration values with simultaneous recorded precipitation rate exceeding 0.1 mm h$^{-1}$ are set to zero in the evapotranspiration time series. Consequently, theses values were discarded from both Büel and SM time series to ensure consistency in the evapotranspiration comparisons.

3.5.1 Global Radiation, Precipitation and Evapotranspiration

A comparative overview of global radiation, precipitation and evaporation sums is shown in Fig. 3.9 for Büel and its SM peer point. The values of the second closest ANETZ station Tänikon are also shown to provide an additional comparison point.

The SM overestimates global radiation for all months without exception and by 10–15 % in average (Fig. 3.9a,b). This confirms the result already obtained from ANETZ comparisons. The daily global radiation differences averaged over each month are displayed in Fig. 3.12a,b. They reveal a small radiative deficit starting at sunrise until 9–10 h UTC followed by a much larger radiative excess until sunset. This is the behavior resulting from computing the radiative scheme only once per hour. The large shortwave radiation excess during the afternoon is a consequence of the insufficient model produced cloudiness.

The amount of moisture provided to the ground is given by comparisons of monthly precipitation sums (Fig. 3.9c,d). They are interesting to examine in parallel with the monthly evapotranspiration sums (Fig. 3.9e,f). For Summer 1996, Büel received each month slightly more precipitation than the SM in average, but produced less evaporation. This is the same for Summer 1995, except August where an important precipitation event occurred in the model on August 9 (Fig. 3.11f). Note that precipitation is not an exactly indicator of the water retained locally in the soil, since run-off can occur especially during heavy precipitation events.

The evapotranspiration is evidently overestimated by the SM, especially during Summer 1996. By looking at the daily evapotranspiration sums (see Fig. 3.10e,f), we see that most of the differences arise during rainy days. The conclusion is the same as for the potential evaporation comparisons from ANETZ stations.

Daily evapotranspiration curves are shown in Fig. 3.13. The evapotranspiration produced by the SM appears to be idealized, i.e. the curves are smooth and symmetrical as compared to the observations. The morning peak of July 1995 results from dew evaporation, which seems to be important in this region (Menzel, private communication). The model produces also dew from time to time, but in relatively small quantities (not shown). The evening peak for June 1995 remains unexplained.
3.5 Comparisons With Büel

![Bar charts showing comparisons of global radiation, precipitation, and evapotranspiration for the hydrological station Büel, the corresponding SM grid-point, and the ANETZ station Tänikon for Summer 1995/96.]

**Fig. 3.9:** Sums of global radiation (a & b), precipitation (c & d), and evapotranspiration (e & f) at the hydrological station Büel, the corresponding SM grid-point and the ANETZ station Tänikon for Summer 1995/96.
Fig. 3.10: Various observations at Büel (left panels) and SM peer point (right panels) for Summer 1996. Available are the 2m and soil temperatures (a & b), soil moisture (c & d), and a superposition of the 2m relative humidity, precipitation, and evapotranspiration observations (e & f).
Fig. 3.11: The same as Fig. 3.10, but for 1995. The SM soil moisture was unfortunately not available that year.
Comparisons With Observations

3.5.2 Soil Moisture

First of all Büel, and its SM peer point, possess quite different hydrological properties. The field capacity \( \theta_{fc} \) and wilting point \( \theta_{wp} \) values assigned to the station of Büel are 0.47 and 0.25 m\(^3\) m\(^{-3}\) respectively (excerpt from Menzel (1997)), whereas the SM corresponding values for loamy soil types are lower (i.e. 0.34 and 0.11 m\(^3\) m\(^{-3}\)). This is not unusual, as hydrological parameters have a great spatial variability. Büel has a greater water storage capacity than the SM grid-point, which can make a difference, especially during periods with little precipitation. On the other hand, available water capacity values \( \theta_{awa} \) are rather similar: 0.22 m\(^3\) m\(^{-3}\) for Büel, and 0.23 m\(^3\) m\(^{-3}\) for the SM. This is due to the relatively high wilting point attributed to Büel.

To establish a suitable comparison base, soil moisture values are normalized as follows:

**Fig. 3.12:** 2m temperature (a) and global radiation (b) differences at Büel for Summer 1996.
3.5 Comparisons With Büel

Fig. 3.13: Evapotranspiration at Büel (left panels) and its SM peer point (right panels) for Summers 1995/96. Hourly observations from evapotranspiration with a simultaneous precipitation rate $P > 0.1$ mm h$^{-1}$ were discarded (see text for explanation).

$$\tilde{\theta} = \frac{\theta - \theta_{wp}}{\theta_{fe} - \theta_{wp}}$$ (3.2)

The explanation for the range of $\tilde{\theta}$ values can be viewed as follows: vegetation transpiration has ceased when $\tilde{\theta} < 0$, but bare soil evaporation still remains possible until the air dryness point is reached. Evapotranspiration is limited by soil moisture conditions when $0 < \tilde{\theta} < 1$ and finally, potential rate is reached and evapotranspiration is no more under soil control when $\tilde{\theta} > 1$. This normalization enables to separate the different possible stages of evapotranspiration and to know under which conditions it takes place. As it is a relative value, the water holding capacity should still be considered to know the real amount of water available for evapotranspiration.

The model stipulated climatological soil moisture $\theta_3$ changes only monthly (see Fig. 3.10d). This value is taken from the global model run at DWD. This ad-hoc
specification was removed from the new ECMWF model, since they adopted a free
drainage boundary condition to reproduce the deep soil moisture more realistically
(Viterbo, 1995).

The wet and dry periods are visible, especially in the surface soil moisture $\theta_1$
and are well reproduced by the SM. They remain dependent on the specific amount
of precipitation and run-off. Precipitation intensity and distribution are different in
the SM and at Büel and is responsible of the different amplitude and variation of
$\theta_1$. The second soil model layer $\theta_2$ has a greater inertia than the observation and
takes also more time to fill after precipitation events. The soil model seems to lose
its moisture throughout the summer, whereas the soil in Büel fills up to reach field
capacity at the end of September. Note that the soil moisture has a the greater
correlation with previous day's precipitation, as compared to the other parameters.

3.5.3 Temperatures

The previous remarks concerning ANETZ observations also apply here. The daily
mean soil temperature $T_M$ and the 2m temperature $T_{2m}$ are closely correlated (see
Fig. 3.11b and Fig. 3.10b). The mean climatological temperature $T_U$ ($z \approx 36\text{cm}$)
is decoupled but its response is less damped than the observed soil temperatures.
This behavior can be explained, given that SM data result from a collection of single
model integrations (an integration per day). The temperatures have no memory of
the previous integrations, because each integration starts with initial conditions
that are not related to the previous integration. This comparisons show that the
assimilation of soil temperature would be a benefit. Note that $T_M$ is also closer to the
ground surface ($z \approx 8\text{ cm}$) than the first measured soil temperature ($z = 20\text{ cm}$). SM
soil temperature show clearly a cold bias resulting possibly from the overestimated
heat capacity.

The daily mean 2m temperature difference, averaged for each month are provided
in Fig. 3.12a,b and show a behavior encountered in many ANETZ stations. The 2m
temperature difference increases in the morning, reaches a maxima around 12 h
UTC and decreases during the afternoon.
Chapter 4

Sensitivity Experiments

Sensitivity experiments are useful for exploring the complex interactions simulated by NWP models and to identify the relevant physical processes and variables. This is especially true for SVAT schemes where non-linear effects are significant (see e.g. Sun and Bosilovich, 1996; Franks et al., 1997; Qu and al., 1998; Rodriguez-Camino and Avissar, 1998).

The present sensitivity study is designed to improve the understanding of soil-vegetation-atmosphere model interactions and to shed light upon the associated feedback mechanisms. The main goal is to determine the significant model parameters that influence temperature and relative humidity at screen-level height.

Previous simulation studies suggest that SVAT processes control screen-level variables in clear sky conditions, rather than planetary boundary layer (PBL) processes. Nevertheless, near-surface parameters are less sensitive to the underlying surface during cloudy and rainy weather. The scope of the present studies is restricted principally to soil/vegetation parameter changes only and this can be justified by the following considerations: PBL parameterizations, which describe momentum, sensible and latent heat flux transfers, are more standardized than soil-vegetation models and unlike the latter, do not offer a broad spectrum of tuning possibilities. Thus, SVAT parameterizations require more profound testing than other parts of the code.

Note however that under some circumstances, PBL parameterizations can inflict major problems. In particular during stable conditions, the surface tends to become thermally disconnected from the atmosphere and thus, results in far more surface cooling than actually observed. This is related to the limited range of validity of the conventional stability functions (Beljaars and Holtslag, 1991).

4.1 Case Studies

Typical summer days with distinct and different weather conditions were selected ("clear sky" and "cloudy+rainy" days). Four summer days meeting these requirements were selected for this study: July 22 and August 18, 1996 are typical examples of sunny days, whereas July 8 and August 7, 1996, are representative of cloudy and rainy days. This choice leaves open the possibility of extrapolating the results to
mixed weather regimes. As the results of the sensitivity experiments show no essential differences, the rest of the discussion is focused on July 8 and July 22 only.

The general meteorological situation over Europe for these two particular days are shown in satellite images (see Figs. 4.1). On July 22, 1996, a high pressure system centered on Denmark moved toward Poland, resulting in the cessation of north-east wind (Bise) dry air flow. Over most of the country, the wind turned to south-west, except in the eastern part of the Swiss Middle-land where a weak north-west wind still dominated. This day was particularly warm and sunny with temperatures reaching a maxima of 26°C. A mean sunshine duration of 12–14 hours was measured at all ANETZ stations.

On July 8, 1996, the meteorological situation were different. A front crossed the country during the day, giving 20–40 mm precipitation to most regions. In the central and eastern Alpine area higher than 2500 m, the rain turned to snow. The sky was overcast everywhere and the temperatures, refreshed by the cold north-west wind, reached only 12°C in average.

4.2 Test Area

From the outset, it was decided to change soil/vegetation parameters for only a limited area, rather than the entire SM domain. This “test area” include mostly Switzerland (as displayed in Fig 2.1b, pp. 8) and it has a sufficient orographic variability for our purposes. Other grounds for that decision are: first, this region holds a large variety of different soil and vegetation conditions; second, averaging screen-level variables and surface energy fluxes over a too large domain makes the interpretation more difficult than for a smaller one, and third, it allows a minimization, as far as possible, of secondary effects of large scale PBL perturbations. With this approach, clear conclusions are expected from the set of sensitivity experiments.

For a finer analysis of the results, the test area is decomposed in three distinct
4.2 Test Area

Fig. 4.2: Sub-domains of the test area (a). Grid-points in regions A, B and C have an altitude below 800 m (white), between 800 and 1500 m (light gray) and above 1500 m (gray) respectively. Soil map for the same area (b). Colors ranging from dark to pale gray correspond to sand, sandy loam, loam, clay loam and clay soil textures respectively. The three water points are omitted and appear white. There is no ice or rock soil types.

Fig. 4.3: Vegetation cover fraction $\sigma$ (c) and rooting depth $d_r$ (d) maps on July 22, 1996. These external fields are similar for July 8, 1996. The limits of the three sub-regions are indicated with contouring at 800 and 1500 m.

Sub-regions (see Fig. 4.2a) according to the same altitude criterion that is used for the verification. Region A comprises all grid-points below 800 m and includes the Swiss Plateau, a part of the Doubs region in France and smaller contributions of the remaining surrounding countries. The locations between 800 m and 1500 m constitute region B, which includes the pre-Alps, the Jura chain and the Schwarzwald. The remaining regions above 1500 m (the Alps) form the third and last region C.

The overall test area is composed of 486 grid-points, sub-divided as follows. Region A: 211 points (43 %), Region B: 129 points (27 %) and Region C: 146 points (30 %). Thus, this grid-point distribution give a proportionate weight to the various regions.

4.2.1 Soil types

The attribution of a soil type/texture influences definitively the model response of each grid-point to external forcing, as the value of many static parameters depends
14 Sensitivity Experiments

<table>
<thead>
<tr>
<th>soil types</th>
<th>Region A</th>
<th>Region B</th>
<th>Region C</th>
<th>Area</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>#</td>
<td>%</td>
<td>#</td>
<td>%</td>
</tr>
<tr>
<td>sand</td>
<td>2</td>
<td>1</td>
<td>21</td>
<td>16</td>
</tr>
<tr>
<td>sandy loam</td>
<td>25</td>
<td>12</td>
<td>12</td>
<td>9</td>
</tr>
<tr>
<td>loam</td>
<td>103</td>
<td>49</td>
<td>67</td>
<td>52</td>
</tr>
<tr>
<td>clay loam</td>
<td>51</td>
<td>24</td>
<td>17</td>
<td>13</td>
</tr>
<tr>
<td>clay</td>
<td>27</td>
<td>13</td>
<td>12</td>
<td>9</td>
</tr>
</tbody>
</table>

| Area | 211 | 43 | 129 | 27 | 146 | 30 | 486 |

Table 4.1: Distribution of soil types/textures among the test area and the three sub-domains A, B and C. The number of grid-points and their percentage with respect to each region are listed in left (#) and right (%) columns respectively.

on that choice. This concerns the hydraulic conductivity and diffusivity coefficients, fixed threshold values like porosity, field capacity, wilting point and air dryness point (see Table B.1, pp. 96), as well as heat capacity, thermal conductivity and surface albedo specifications (see Table B.2, pp. 96). The soil type/texture distribution for the test area is displayed in Fig. 4.2b.

Loamy and sandy loam soils cover two thirds of the map and are the most common soil types found in regions A and B (~70 %), while sandy loam is the dominant type in the Alpine region (~60 %). Clay loam and clay soils typify region A, whereas in regions B and C, sandy soils predominate.

Note that sandy soil type is in effect a compromise in mountainous regions. It has a low heat capacity and a small water capacity (porosity) but nevertheless permits evapotranspiration. The use of rock or ice soil types may be more appropriate at high altitudes but raises some difficulties. Rocky soils cannot evaporate and hence produce an unrealistic ground temperature.

Only three points are defined as water, two for the Lac Léman (i.e. the west) and one for the Bodensee (i.e. the north-east). In effect, the current spatial resolution of the SM is not able to resolve the numerous smaller lakes. Note that water points are left unchanged in the sensitivity experiments and are not taken into account in the comparisons.

4.3 Method

All sensitivity studies were performed with an operational version of the SM, slightly modified for the purpose of the experiments. The case of July 22 serves to illustrate the methodology, which is also applied to all days selected for this study.

All model runs are started some time earlier on the previous day at 12 UTC (July 21 at 12 UTC) and integrated in the same way as the control run during the first twelve hours (i.e. without modifications). For the test runs only (not...
4.4 Result of Experiments

A number of experiments were conducted to test the sensitivity of screen-level variables. Vegetation cover fraction experiments $\sigma^-$ (bare soil everywhere) and $\sigma^+$ (vegetation everywhere), rooting depth experiments $d_r^-$ (minimum value $d_r = 0.12$ m) and $d_r^+$ (maximum value $d_r = 0.7$ m), soil moisture (first layer) experiments $\theta_k^-$ ($\theta_1$ set to the air dryness point) and $\theta_k^+$ ($\theta_1$ set to porosity), and soil moisture (second layer) experiments $\theta_k^-$ ($\theta_2$ set to $\theta_{adp}$) and $\theta_k^+$ ($\theta_2$ set to $\theta_{sol}$), soil temperature profile experiments $T^-$ ($-3$ K) and $T^+$ ($+3$ K). Several additional experiments with albedo, roughness length and interception reservoir changes were also performed. An overview over the experiments is shown in Figs. 4.4 and 4.5, in form of 2m temperature and humidity difference maps at 14 hour UTC.

The temporal evolution of the mean screen-level variables for the control run are shown in Fig. 4.6 (July 22, 1996), and Fig. 4.7 (July 8, 1996). The mean surface energy fluxes for regions A and C are shown together in Fig. 4.8 (both days). These two particular days exhibit, as expected, quite different behavior. First, the control integrations are discussed (bold lines + filled symbols in Figs. 4.6-4.8).

On July 22, the 2m temperature has an amplitude of about $\sim 13.6^\circ$ C in region A and one more degree in regions B and C. Note that the higher the altitude, the earlier the 2m temperature maxima occurs. Examination of the surface energy fluxes (Fig. 4.8) shows that the evapotranspiration flux maxima also follows this form of

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Region A</th>
<th>Region B</th>
<th>Region C</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\sigma$</td>
<td>$0.82 \pm 0.05$</td>
<td>$0.57 \pm 0.09$</td>
<td>$0.25 \pm 0.07$</td>
</tr>
<tr>
<td>$d_r$</td>
<td>$0.55 \pm 0.06$</td>
<td>$0.30 \pm 0.07$</td>
<td>$0.14 \pm 0.02$</td>
</tr>
<tr>
<td>$\theta_1$</td>
<td>$0.18 \pm 0.26$</td>
<td>$0.35 \pm 0.43$</td>
<td>$0.50 \pm 0.46$</td>
</tr>
<tr>
<td>$\theta_2$</td>
<td>$0.39 \pm 0.22$</td>
<td>$0.87 \pm 0.30$</td>
<td>$1.55 \pm 0.37$</td>
</tr>
</tbody>
</table>

Table 4.2: Mean test parameters in the three sub-regions. $\sigma$ (m$^2$ m$^{-2}$), $d_r$ (m) and $\theta_k$ (m$^3$ m$^{-5}$) are the vegetation cover fraction, rooting depth, and reduced soil moisture in the $k$th layer respectively. Soil moisture left and right column values are for July 22 and July 8 respectively. The standard deviations for both days are comparable.

The procedure is applied to the various soil/vegetation model parameter changes and this forms the framework for the sensitivity experiments. Following the discussion of the results in the next section, we will also address the relevance of local versus large-scale effects in Sec. 4.5.

4.4 Result of Experiments
Fig. 4.4: 2m temperature difference maps (experiment - control run) for the most valuable sensitivity experiments on July 22, 1996 at 14 hour UTC. Vegetation cover fraction experiments $\sigma^-$ (a) and $\sigma^+$ (b), rooting depth experiments $d_1^-$ (c) and $d_1^+$ (d), soil moisture (first layer) experiments $\theta_1^-$ (e) and $\theta_1^+$ (f), and soil moisture (second layer) experiments $\theta_2^-$ (g) and $\theta_2^+$ (h).
Fig. 4.5: The same as Fig. 4.4, but for the 2m relative humidity.

temporal evolution, but this does not apply to the sensible heat flux. The amplitude
of the daylight depletion of the 2m relative humidity is, as for the 2m temperature, less accentuated at higher altitudes, as the grid-points in this region are closer to the free atmosphere. The transition of the night-time stable atmosphere to the daily unstable atmosphere is easily identified around 8 UTC for regions B and C. In region A, the transition occurs earlier at around 6 UTC and is distinctly smoother.

On July 8, the 2m temperature decreases throughout the day, excepted for a small warming up in the afternoon. The sensible and latent heat fluxes suffer from the shortage of shortwave radiation, which accounts for only one third of the solar energy available on July 22 and this has a negative impact on all energy fluxes of the SVAT system. The period after sunset is especially interesting. The latent heat flux is negative (\(\sim 50 \text{ W m}^{-2}\)), whereas it is close to zero on July 22. This indicates that thermal energy is decreasing to the benefit of the ground surface. This energy is in turn transferred directly into latent heat, as the ground heat flux remains unchanged. Also a mechanism seems to be triggered by some rainy conditions, allowing a weak, but sustained evapotranspiration during the night. This effect is probably connected to the limitations of PBL parameterization of stable atmospheric conditions (see earlier). Clouds have a clear positive impact on the longwave radiation balance on this day.

Note that the maximum values of the evapotranspiration flux are approximately the same whatever the meteorological situation, and this merits further analysis. Screen-level difference maps for the most significant sensitivity experiments are available on Figs. 4.4 and 4.5 on 22 July. (They are not provided for July 8, as the results are less interesting.)

4.4.1 Vegetation

The vegetation parameterization, as described in App. B (see pp. 96), relates to two parameters: the vegetation cover fraction and the rooting depth. First, the vegetation fraction is the part of the grid-point surface covered by vegetation, the remaining being bare soil. This partitioning influences the evapotranspiration distribution itself. Recall that evapotranspiration is the sum of bare soil evaporation and vegetation transpiration (Eqn. 2.27, pp. 17) and that bare soil takes water from the first active soil layer exclusively, whereas all active soil layers can contribute to the vegetation transpiration processes. The experiments with the vegetation fraction is designed to allow one to distinguish bare soil evaporation from vegetation transpiration.

Second, the rooting depth experiments (Exps. \(d^\pm\)) modify the transpiration rate of the vegetation only, but do not influence directly bare soil evaporation (Figs. 4.9, 4.4.c,d, and 4.5.c,d). Removing all vegetation (Exp. \(\sigma^-\)) lowers the 2m temperature maxima and raises the 2m relative humidity. For a fully vegetated area (Exp. \(\sigma^+\)), we obtain exactly the inverse behavior. Vegetation is able to transpire up to 0.7\(E_{pot}\)), whereas bare soil can evaporate up to potential evaporation. Increasing the vegetation fraction reduces therefore the maximum evapotranspiration value.

Vegetation fraction modifications imply a change of the surface albedo (see Eqn. B.1, pp. 95). Without vegetation, the albedo is much higher than in the
4.4 Result of Experiments

Fig. 4.6: 2m mean screen-level variables for the vegetation cover fraction experiments $\sigma^-$ (left panels), and $\sigma^+$ (right panels) on July 22, 1996. In experiment $\sigma^-$, bare soil is everywhere (no vegetation at all). In experiment $\sigma^+$, vegetation is everywhere (no bare soil at all). Control run (bold lines + filled symbols) and experiments (dotted lines + empty symbols) are displayed together.

Case of full vegetation. Vegetation has an albedo of 0.15, whereas the mean bare soil albedo has values between 0.20 and 0.25, depending on the soil moisture content. In this case the albedo reduces to the unique and constant value of the vegetation albedo. This is the explanation for the changes of the shortwave radiation budget $S$ in Figs 4.8. The longwave radiation budget $R$ is only slightly affected by the changes of the surface temperature, and can be neglected for the present discussion.

Decreasing the rooting depth (Exp. $d^-_r$) raises the 2m temperature maxima and lowers the 2m relative humidity (i.e. less evapotranspiration). By increasing the rooting depth (Exp. $d^+_r$), the inverse behavior is obtained. For this particular sequence of processes, there is simply an energy transfer between the sensible and latent heat fluxes, since the ground heat flux remains unchanged.

The most striking changes appear in the sensible and latent heat fluxes. This
remark remains valid for all these sensitivity experiments. In other words, the Bowen ratio – the ratio between sensible and latent heat flux – is modified significantly, influencing screen-level variables. In contrast, the longwave radiation is not very sensitive, as it depends only on the surface temperature $T_B$ and the ground heat flux, at least if there is no feedback with the cloud cover. The latter is calculated as the remaining part of the surface energy balance and is somewhat more sensitive.

The rooting depth controls also the amount of water extracted from the soil by the plants. It is difficult to link the root parameterization to a real physical counterpart. The first active layer contribution is always greater than the second layer, and predominates when the rooting depth is low (especially in mountainous region). A short rooting depth could represent grass with roots mostly close to the surface. Higher rooting depth could represent forests, but note that trees can take water from the lowest active layers, whereas this parameterization doesn’t give a higher priority to the lowest layers for longer roots. In effect, the rooting depth is a practical means of modulating transpiration, but the associated physical interpretation
is not so obvious.

For all days selected for the sensitivity studies, both parameters attain their maximum value for the test area (see Fig. B.1, pp. 97).

The vegetation fraction and the rooting depth are obviously correlated with the orography (see Figs. 4.3). This follows directly from the design of the parameterization, which incorporate a strong dependency on height, especially for the rooting depth (see Eqn. B.2, pp. 96). Of the two parameters, the vegetation fraction has the most impact on screen-level variables changes (Figs. 4.6, 4.7, 4.4a,b, and 4.4a,b).
Fig. 4.9: 2m mean screen-level variables (a, b, c & d) and corresponding mean energy surface fluxes in Region B (e & f) for the rooting depth experiments $d_r^-$ (left panels) and $d_r^+$ (right panels) on July 22, 1996. In experiment $d_r^-$, the rooting depth is set to the minimum (0.12 m). In experiment $d_r^+$, the rooting depth is set to the maximum (0.7 m).
4.4 Result of Experiments

Fig. 4.10: Normalized soil moisture maps of the first (top panels) and second (bottom panels) layers on July 22 (left panels) and July 8 (right panels), 1996 at 0 h UTC. The soil moisture values are normalized as follows $\theta_k = 100 \cdot (\theta_k - \theta_{fc})/(\theta_{fc} - \theta_{wp})$. Negative values indicate that vegetation transpiration has ceased and values above 100 indicate areas evaporating at the maximal possible rate.

4.4.2 Soil Moisture

Changes in the soil moisture contained in the two hydrological layers have different timescales depending on their respective holding capacities. Deep soil moisture values do not differ substantially between July 8 and July 22 (see Figs. 4.10c,d) and all regions are above wilting point. Most of Region C is even above field capacity, as transpiration is not particularly active in mountainous areas. On the other hand, surface soil moisture fields can have quite different patterns over two weeks. This layer respond quickly to precipitation events and evapotranspiration (see Figs. 4.10a,b). The lowest parts of Region A are close to saturation on July 8 and already below wilting point on July 22, due to evapotranspiration activity.

Experiments examining variations of the soil moisture show the strongest responses in the screen-level domain. The signal is clearly seen on July 22, but is much weaker on July 8. On setting the soil moisture of the first layer to its air dryness point (Exp. $\theta_1^a$) the most sensitive area appears to be Region C. (see Figs. 4.11, 4.4e, and 4.5e). In mountainous region, both vegetation fraction and rooting depth are lower than in the other sub-areas, that is bare soil evaporation is the dominant mechanism, and thus it is very sensitive to soil moisture changes in first active layer. Fig. 4.13 demonstrates this effect: the latent heat flux is close to zero. Temperature differences up to 4° C reach their maxima at midday, at the same time as the sen-
Fig. 4.11: 2m mean screen-level variables for the upper soil moisture layer experiments $\theta^-_1$ (left panels) and $\theta^-_2$ (right panels) on July 22, 1996. In experiment $\theta^-_1$, the upper soil moisture value is set to the air dryness point $\theta_{adp}$. In experiment $\theta^-_2$, the soil moisture value is set to the porosity $\theta_{sat}$.

The mean surface energy fluxes for all soil moisture experiments are represented in Figs. 4.13. They show that the latent heat flux modifies the Bowen-ratio and that the intensity of sensible heat flux is adjusted consequently to the new energy surface balance. The other fluxes (ground heat flux excepted) are barely affected by soil moisture changes.

On July 8, the influence of soil moisture changes affects energy surface fluxes and screen-level variables to a much lesser extent (see Figs. 4.14). The energy fluxes are weak and the atmosphere controls mostly screen-level variables, and therefore the impacts of soil moisture changes are much smaller than in fair weather.
4.4 Result of Experiments

4.4.3 Soil Temperature

The purpose of this experiment is to examine the impact of soil energy content changes. Instead of modifying each temperature variable separately, it was decided to shift uniformly the soil temperature profiles by ±3 K (Exps. $T^\pm$). Fig. 4.15 shows that the energy fluxes are little influenced by temperature changes (ground heat flux is the exception). As a consequence, the relative humidity remains unchanged. For the 2m temperature, the differences appear clearly at +1 h UTC, just after the modification. The initial temperature difference decreases rapidly when the sun starts to shine, and reaches a minimum at midday. When the shortwave radiation intensity declines, the temperature difference grows again, but with a weaker amplitude, compared to the initial amplitude (one third). This means that the deep soil temperatures $T_M$ and $T_U$ influences the surface temperature $T_B$ mostly during the night, when the sensible and latent heat flux are weak, compared to the soil heat fluxes.

Fig. 4.16 displays the soil temperatures $T_B$ and $T_M$, the temperature of the first
atmospheric layer $T_{30M}$ ($\sim 30$ m) and the 2m temperature. Somewhat surprisingly, the differences at $30$ m are very small. This comparison reveals the dominant influence of $T_B$ with respect to $T_{30M}$, on the screen-level temperature, and this is particularly evident after $+18$ h UTC. The absence of pronounced effects at $30$ m implies that the night-time boundary layer is too stable in the mean to allow for a notable reaction of the vertical energy fluxes in response to the increased soil temperature.

### 4.4.4 Other Sensitivity Experiments

Further experiments were undertaken, with other parameters than vegetation, soil moisture content or soil temperature. For the sake of brevity, only a short overview is given here, and graphical results are not presented, as the impacts are much weaker.

The modification of the surface albedo by - 5 % or + 5 % has the immediate
4.4 Result of Experiments

Fig. 4.14: 2m mean screen-level variables (a, b, c & d) and corresponding mean energy surface fluxes in Regions C (e) and A (f) for the soil moisture experiments $\theta_1$ (left panels), and $\theta_2$ (right panels) on July 8, 1996.
58 Sensitivity Experiments

Fig. 4.15: 2m mean temperature (a & b) and corresponding mean energy surface fluxes in Region A (c & d) for the soil temperature experiments $T^-$ (left panels) and $T^+$ (right panels) on July 22, 1996. All soil temperatures ($T_B$, $T_M$, and $T_U$) are shifted by -3 C ($T^-$) and +3 C ($T^+$) at 0 h UTC.

Consequence of raising or reducing the shortwave radiation balance at the ground respectively. The gain or loss of energy is then equally distributed between the latent and sensible heat fluxes. The ground heat flux and the longwave radiation balance, as well as the 2m relative humidity remain unaffected by the albedo change. Differences in the 2m temperature are only noticeable on July 22. These differences are small (less than a third of a degree), but yield similar values among the three sub-regions. This means that the effects of an albedo change are essentially independent of the altitude. The results of July 8 are not particularly revealing, because the shortwave radiation is already small.

The modification of the roughness length for momentum $z_o$ by -0.1 m or +0.1 m has no observable effects on screen-level variables. The roughness length contributes to the computation of the sensible and latent heat fluxes (Eqn. 2.15 and 2.16), as does the roughness length $z_H$ for heat and moisture. This latter parameter has a
constant value 0.1 m. Experiments modifying $z_H$ were not attempted, but differences are expected for the following reason. The factor $\frac{h}{z_H}$ appears in Eqn. 2.17 (pp. 14), at the denominator and in the transfer function $F$. The value of $h$ is about 30 m. Increasing $z_H$ stimulates the flux transfers, and vice-versa. Both sensible and latent heat fluxes are affected, as the coefficients for heat and moisture are the same. By differencing the roughness length and the transfer functions for these processes, a finer control of these fluxes would be possible.

Changing the values of the climatological reservoir $\theta_3$ does not affect screen-level variables or surface energy fluxes at all during the specified period.

For the last experiment, the interception reservoir was filled. The interception reservoir evaporates at the potential rate and hence, the total evapotranspiration flux is increased. As a result, there is a slight decrease of the 2m temperature, as well as an increase of the 2m relative humidity on July 22. Region C is less sensitive to this effect since the reservoir capacity depends on the vegetation cover fraction (Eqn. 2.31, pp. 18), and at these altitudes, the reservoir capacity is small.
Sensitivity Experiments

Fig. 4.17: Results of the experiment $\theta_v$ for July 8, 1996. The first soil layer is set to the air dryness point at 0 UTC in the test area only (a & c), and in the entire SM domain (b & d). For each region, the 2m mean temperature (top panels) and 2m mean relative humidity (bottom panels) is displayed for the control run (bold lines) and the experiment (dotted lines).

and evaporates much more quickly. There was no change on July, 8, as the reservoir of most grid-points were already filled up with precipitation.

4.5 Local Versus Large-scale Effects

To evaluate the difference between the impact of a large scale compared to a local change in surface parameters, experiments showing the strongest response on the amplitude of screen-level variable in the test area were repeated, but this time at full scale (whole SM domain). Differences between full and test area on July 22 experiments are negligible, especially for the 2m temperature, whereas the 2m relative humidity appears to be more sensitive (not shown).

To a measure, the results of the experiment displayed in Fig 4.17 are the clearest
of the series. This is the experiment where the upper soil layer is completely dry on July 8. The differences for the 2m relative humidity are more pronounced than for the 2m temperature, but are nevertheless quite reasonable. The amplitude only is modified and the results remain homogeneous. This tends to support the contention that screen-level variables depend, above all, on local soil conditions.

In conclusion, perturbing the soil conditions on the entire SM domain or on a limited area is not of great concern and this help justify our approach.

4.6 Discussion

In mountainous regions, the evapotranspiration proceeds almost from bare soil evaporation, because vegetation is sparse. Thus great care needs to be given to the formulation of evaporation formulation at altitudes higher than 1500 m. However, the decomposition of the evapotranspiration between vegetation transpiration and bare soil evaporation is less obvious at lower altitudes. Vegetation plays a dominant role, but bare soil is still significant, the latter accounts for up to one third or even more of the total evapotranspiration. This behavior is parameterization dependent.

Note that for many soil models, evapotranspiration is more significant than bare soil evaporation.

The sensitivity studies reveal that soil moisture is by far the most important soil model parameter influencing temperature and relative humidity at screen-level height. A soil moisture change affect the evapotranspiration, then the Bowen-Ratio and finally screen-level variables. On night-time, the 2m temperature depends mainly on the ground surface temperature.

Similar conclusions for several soil models were found by Rodriguez-Camino and Avissar (1998), by using the Fourier amplitude sensitivity test. They found that soil moisture and vegetation parameters where the dominant parameters under buoyant conditions. Roughness length was mostly important under stable atmospheric conditions.
Seite Leer / Blank leaf
Chapter 5

Model’s Evapotranspiration

The previous sensitivity studies have shown that evapotranspiration is the relevant process that modifies the Bowen-ratio and hence, screen-level variables. Consequently, a more detailed analysis and discussion of the corresponding parameterization is called for.

Vegetation parameters play a crucial role in the evapotranspiration formulation. In Fig. 5.1a, vegetation cover fraction and rooting depth parameters are represented for July 22, 1996 as a function of the altitude. This figure was obtained from all grid-points within the test area, i.e. the model sub-domain used for the previous studies. The vegetation cover fraction exhibits a light variation around its main curve, as the minimum and maximum values are specific to each grid-point (see Eqn. B.2, pp. 96). There is no such variation for the rooting depth curve, as the specified minimum and maximum values are shared by all grid-points.

By design, both components of the evapotranspiration are proportional to potential evaporation. Bare soil evaporation $E_b$ and vegetation transpiration $E_v$ can therefore be expressed relative to $E_{pot}$:

$$
\tilde{E}_b = \frac{E_b}{E_{pot}} = (1 - \sigma) \cdot \beta_b^2
$$

$$
\tilde{E}_v = \frac{E_v}{E_{pot}} = \sigma \cdot \sum_{k=1}^{N} r_k \beta_{vk}^2
$$

With operational settings, $N = 2$ is the number of active hydrological layers and the corresponding rooting depth factors are $r_1 = 0.1$ and $r_2 = d_r - r_1$ (see pp. 15 for the definition of $\beta_b$, $\beta_{vk}$ and $r_k$).

5.1 Normalized Evapotranspiration

Maximum evapotranspiration values are reached when the soil moisture content of both hydrological layers is above field capacity (i.e. $\beta_b = 1$ and $\beta_{vk} = 1$, $k = 1, \ldots, N$). In Fig. 5.1b, the maximal normalized evaporation components $\tilde{E}_b$ and $\tilde{E}_v$, as defined by Eqs. 5.1, and their sum $\tilde{E}_{tot}$ are displayed as a function of the altitude. Vegetation
has clearly a major impact on the partitioning of the evapotranspiration. In July, at the height of 1400 m (and rooting depth of 0.2 m), both layers contribute to the transpiration with equal amounts, but the overall quantity is already small (10% of $E_{pot}$). At higher altitudes, the evapotranspiration is dominated by bare soil evaporation. A normalized evapotranspiration maximum of $0.88E_{pot}$ is reached at the highest altitude, where the vegetation cover fraction is minimum. In this region, the difference between $E_{tot}$ and $E_b$ results essentially from the transpiration of the upper layer.

The contribution of the upper layer to the transpiration is the same for every grid-point, because the layer is 0.1 m thick and rooting depth values below 0.12 m are not allowed. Potential evaporation cannot be reached, because the vegetation cover fraction has still a minimum value of 0.12 m$^{-2}$ at 3000 m. In the region between 800 and 1500 m, the normalized maximum evapotranspiration curve has a minimum at about $0.6E_{tot}$, which originates in the rapid decrease of transpiration associated with the concomitant slower increase of bare soil evaporation with height. At around 800 m, the two components have the same weight, with a value of about $0.3E_{tot}$. This happens approximatively in the same region as the previously noted minimum. At lower altitudes, the evapotranspiration is dominated by the transpiration of plants, but the latter can provide only a maximum value of two third of the potential evaporation. For its part, bare soil evaporation still plays a significant role, even at the lowest altitudes. A consequence of the vegetation cover fraction parameterization is that $E_{tot}$ decreases with height in this region.

It should be noted that the variation of potential evaporation (PE) with height is not taken into account in this discussion. Measurements of potential evaporation from ANETZ stations suggest a light decrease of PE with height. Nevertheless, it is difficult to draw a firm conclusion, as the number of stations located in mountainous
5.1 Normalized Evapotranspiration

<table>
<thead>
<tr>
<th>station</th>
<th>( h )</th>
<th>( \sigma )</th>
<th>( d_r )</th>
<th>( \overline{E}_h )</th>
<th>( \overline{E}_v )</th>
<th>( \overline{E}_{tot} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Stabio</td>
<td>339</td>
<td>0.88</td>
<td>0.64</td>
<td>0.12</td>
<td>0.56</td>
<td>0.68</td>
</tr>
<tr>
<td>Biel</td>
<td>748</td>
<td>0.73</td>
<td>0.46</td>
<td>0.27</td>
<td>0.34</td>
<td>0.61</td>
</tr>
<tr>
<td>La Chaux-de-Fonds</td>
<td>1045</td>
<td>0.60</td>
<td>0.32</td>
<td>0.40</td>
<td>0.19</td>
<td>0.59</td>
</tr>
<tr>
<td>Gütsch</td>
<td>1679</td>
<td>0.33</td>
<td>0.16</td>
<td>0.67</td>
<td>0.05</td>
<td>0.72</td>
</tr>
</tbody>
</table>

Table 5.1: Vegetation parameters and maximum normalized evapotranspiration values for the specific grid-points used in Figs. 5.2. \( h \) is the height of the grid-points (m).

regions is sparse and the observed PE exhibits a large variation. Even with a PE correction (to a reference value at 500 m for example), the normalized evapotranspiration minimum at 950 m would still be present. Actually, this minimum is unlikely to have a real world counterpart, but without additional observations, a definitive answer is not possible.

The maximum normalized evapotranspiration was the topic of the previous discussion, but for a typical summer case, these values are rarely reached, as the surface soil moisture is lower than field capacity, especially on the Swiss Plateau. Four grid-points selected for the verification, including the comparison point for the station Biel, are used to explore the dependence of evapotranspiration on soil moisture and altitude further. These grid-points have only different heights and soil types, but otherwise no other special significance. (Different points could have been used as well to illustrate the discussion.)

The normalized evapotranspiration maps as a function of the water content are shown in Figs. 5.2 for the specified locations. The distinct contribution of bare soil evaporation and vegetation transpiration are also displayed. The area, delimited by both wilting point and field capacity threshold values, have different sizes and locations, as the specified soil type is different for each grid-point. It follows directly from its definition, that bare soil evaporation is independent of the second soil moisture layer. As a consequence, bare soil evaporation is only represented with straight vertical lines, whose density increase with height. Transpiration is affected by both layers, but since most of the roots are contained in the second layer, the contribution of the latter is decisive. This is not true anymore, when the moisture layer \( \theta_2 \) is close to the wilting point, or in the Alpine region, where the rooting depth is minimum. For these special cases, the transpiration originates essentially from the upper layer, but with a very weak value.

From these figures, it is obvious that more than one soil moisture configuration will produce the same evapotranspiration value. This remark still remains valid for other SVAT schemes. Outside the box, the evapotranspiration depends merely on only one of the two evaporation sources. In the lower left corner, the soil is completely dry and does not evaporate: in this region, the evapotranspiration is very sensitive to small changes in soil moisture. In the upper right corner, soil moisture is above field capacity and the soil evaporates at potential rate: in this region, soil moisture changes will not affect evapotranspiration. The size of these two regions depends on the soil type. For sandy soils, the soil moisture level should be lower than
Fig. 5.2: Normalized evapotranspiration maps for the corresponding grid-points: Stabio (a), Büel (b), La Chaux-de-Fonds (c) and Gutsch (d) as a function of the normalized water content. At the bottom, a spatial representation is reproduced for figures b and c. All specific values are grouped in Table 5.1.
5.2 Further Analysis of Case Studies

The sensitivity experiments carried out in the previous chapter were focused on changes of screen-level variables and energy surface fluxes, and the results were classified in relation to three predefined regions. Additional information about soil moisture variation, evapotranspiration and run-off can also be extracted from numerous runs, especially those involving radical changes in soil moisture content. Indeed, water content was found to be a key parameter and a better understanding
of the soil moisture transfer appears of primary importance. The cases of July 8 and July 22, 1996 are used for this purpose.

In Fig. 5.3a,c,d, soil moisture differences of the three soil model layers are displayed as function of the altitude for July 8, 1996 at 0 h UTC. Each layer has its own distinctive distribution. For July, the soil moisture values of the climatological layer $\theta_1$ are all below field capacity. A somewhat artificial upper limit is visible at about 80% of the available water capacity. There seems to be a slight tendency towards higher values with the increase in altitude. It is difficult to be more specific about it, as this field is purely static for a whole month.

The case of the middle soil moisture layer $\theta_2$ is more interesting. Note that the apparent linear relationship is a pure artifact, as the reduced soil moisture maximum depends on the soil type. The distribution indicates that the water content is below field capacity at low altitudes and consequently, the transpiration is under soil control. In the transition region, the majority of grid-points is still below field capacity, but a significant part is able to transpire at potential rate. In mountainous regions, almost all points are above field capacity and transpiration is therefore not limited by soil conditions. Note the relative compactness of low altitude points, consequence of the greater efficiency of vegetation transpiration in this region. $\theta_2$ is the only soil moisture layer showing an obvious dependence on altitude. This is related to the strong coupling of both vegetation cover fraction and rooting depth with height.

The soil moisture distribution of the surface layer $\theta_1$ exhibits a larger variation. It should be emphasized that the distribution is a snapshot, representative for a short time window only. This is obvious, when the distribution of $\theta_1$ for July 8 and July 22 are put side by side. The larger variation is encountered in the pre-Alps and the Alps. A single point (~2800 m) has a high moisture value, which is explained by the presence of a snow cover in this grid-point. On July 22, the moisture values of most high altitude points are within water holding capacity limits. This can be explained by the bare soil evaporation activity in these regions. The surface layer can also react very quickly, especially during rainy events.

The sensitivity experiments $\theta_1^\pm$ of July 22, 1996 are now further analyzed, but this time from an hydrological point of view. In these experiments, the two active soil layers are independently saturated, or dried. They show how run-off processes are triggered, evapotranspiration modified and soil moisture transfer affected by a sudden moisture supply or removal.

In Figs. 5.4, the soil moisture changes are displayed for this subset of experiments. The total soil moisture content $\theta_1 + \theta_2$ (Figs. 5.4c,f) reveals a net loss of soil moisture, from a minimum value of ~4 mm at low altitudes and up to 12 mm around 2000 m for experiment $\theta_1^+$. These values are lower for the experiment $\theta_2^+$. If the layers are considered separately (Figs. 5.4b,d), the situation of the two active layers are opposite to each other. For experiment $\theta_1^+$, the upper layer can only loose water, because it starts initially from a saturated state, and this is true for all points without exception. The second layer receives one part of this water from the upper layer by percolation. Nevertheless, some of the points have a negative balance, especially grid-points higher than 2000 m. At the latter altitudes, the percolation
Fig. 5.4: Daily soil moisture changes within the test area for the four experiments $\theta_{1,2}$ on July 22, 1996 as a function of the altitude (see chap. 4, pp. 41 for a description of $\theta_{1,2}$).
FIG. 3.5: Daily evapotranspiration sum within the test area for the four experiments $\theta_{1,2}$ on July 22, 1996 as a function of the altitude (see chap. 4, pp. 41 for a description of $\theta_{1,2}$).
Fig. 5.6: Soil moisture changes, evapotranspiration and run-off sums within the test area for the control run on July 8, 1996 as a function of the altitude.
process is not efficient, because $\theta_2$ is already above field capacity, thus reducing the hydrological conductivity and diffusivity (e.g. Eqn. 2.9, pp. 12). Another part of the water excess leaves the system with ground run-off (Fig. 5.4e), but this process is only relevant starting at at height of about 1000 m, where $\theta_2$ is higher than 80% of its reduced soil moisture value. Below this point, the second layer has enough room to absorb the entire percolation contribution from the surface layer. The more the second layer is saturated, the more run-off occurs. Note also the effect of the soil type on the structure (bands) of the soil moisture differences (Figs. 5.4b,d).

On July 8, the incoming solar radiation is weaker and precipitation is the additional external forcing, which has major consequences on the hydrological processes (see Figs. 5.6). We see that the major part of precipitation is mostly lost by surface run-off and to a lesser extent by ground run-off. As a result, there is a net gain for the second soil moisture layer, but the situation for the first layer is more confused. It seems that the grid-points where evapotranspiration has reached abnormal values
have lost water in the first layers, whereas the other ones have gained moisture. Therefore, it should be checked if the soil moisture refill is correctly modeled during rainy events, but this is beyond the scope of this study.

5.3 The 2m Temperature Maximum

On July 22, 1996, the 2m temperature maximum differences are well correlated to the daily evapotranspiration differences for the four experiments $\theta_{1,2}$ (see Figs. 5.7). There is clearly a linear relationship between these two quantities ($\Delta E \propto \Delta T_{2m}^{\text{max}}$), which cannot be considered as obvious at first sight. This information, although interesting, does not say more about the nature of the dependency of evapotranspiration and soil moisture. The fact is that evapotranspiration changes depend on both potential evaporation and soil moisture changes. This is the reason why it is not possible to derive a simple relationship between these two quantities. This information gives only the water quantity which can be added or removed from the soil for a given difference of the 2m temperature maximum.

A relationship between $\Delta E$ and $\Delta E$ exists but is not linear, and the grid-points above 1500 m exhibits a large variation (not shown). One influence is that the non linear evapotranspiration processes invite consideration of this problem from a different standpoint (e.g. for the variational technique). This forms the subject of the next chapter.
Chapter 6

Variational Studies

The potential usefulness of variational methods for meteorological problems was pointed out very early by Sasaki (1958). Variational techniques have been applied primarily to analysis, usually of mesoscale fields (Sasaki and Goerss, 1982), and to the initialization of NWP models, e.g. suppression of unrealistic high frequency gravity oscillations from the initial conditions of a numerical forecast (see Daley, 1978). Although these techniques have been known for a long time, they have caught on only recently in the operational area, due to their extremely high computational demands. The underlying mathematical concepts behind variational methods are set out later.

The assimilation of soil moisture in particular, has become an active research theme during the last few years. In NWP models, this field is the key component of the hydrological cycle that controls the evapotranspiration. The great sensitivity of NWP models to the soil wetness specification has been confirmed in many studies, our own sensitivity experiments included.

Mahfouf (1991) has studied the feasibility of retrieving soil moisture content with two different methods using screen-level information: a one-dimensional variational technique and sequential assimilation. The latter, based on statistical information gathered from numerous model runs, was preferred due to its simplicity and speed. Sensitivity and calibration studies were then carried out by Bouttier et al. (1993b), and sequential assimilation was finally implemented in a mesoscale model (Bouttier et al., 1993a). Recently, Calvet et al. (1998) studied the possibility of retrieving root-zone soil moisture from surface soil wetness or temperature estimates. The feasibility of assimilating soil moisture fields in rather dry areas using satellite data and a variational method, was demonstrated by van den Hurk et al. (1997).

A case study on soil moisture assimilation from atmospheric observations and based on the work of Mahfouf (1991) was already conducted by Callies et al. (1998) with a one-dimensional version of the EM. It has been extended later to the whole EM domain (Rhodin et al., 1998). But the procedure was only tested for Spring, where the soil moisture conditions differ considerably from the ones found in Summer.

Thus, the assimilation of soil moisture is now regarded as a necessary part of NWP. For example, the location and intensity of precipitation, as well as the evapo-
transpiration are subject to errors, creating anomalies in soil moisture fields. Model drifts (e.g. 2m temperature, precipitation) might originate from these anomalies if no correction is applied.

### 6.1 The Variational Method

The variational calculus is a mathematical tool used to solve a category of problems requiring the computation of extrema. The classical illustration of the variational principle is the following, i.e. find the function \( u(x) \) that minimizes the given functional \( I \):

\[
I(u) = \int_{x_a}^{x_b} F(x, u, \dot{u}) \, dx
\]

subject to \( u(x_a) = \alpha, \ u(x_b) = \beta \), where \( \alpha \) and \( \beta \) are specified and \( u \) and its derivatives up to the second order are continuous \( (\dot{u} = d_x u) \). Letting \( \dot{u} = u + \epsilon \eta \) with \( \eta \) arbitrary and such that \( \eta(x_a) = \eta(x_b) = 0, \epsilon \) being a small parameter, it can be shown that \( u(x) \) is a stationary point (i.e. for which \( \delta I(u) = 0 \)) only if:

\[
\frac{\partial F}{\partial u} - \frac{d}{dx} \left( \frac{\partial F}{\partial \dot{u}} \right) = 0
\]

Eqn. 6.2 is the famous result of the variational principle known as the Euler-Lagrange equation. In meteorology, the functional \( I \) is replaced with a discrete cost function \( J \), or penalty function, which measures the misfit between the model and the observations at a given time:

\[
J(x) = \frac{1}{2} \left\{ (y(x) - \hat{y})^T \cdot (O + F)^{-1} \cdot (y(x) - \hat{y}) + (x - x_b)^T \cdot B^{-1} \cdot (x - x_b) + J_c \right\}
\]

where \( x \) is the state vector of model variables to be optimized, \( \hat{y} \) and \( y(x) \) are the observations and the observation operator applied to the model state vector, respectively, and \( x_b \) is the background term, or first guess vector of model variables. \( O, F \) and \( B \) are the expected error covariance matrices for observation, model and background terms respectively. Additional constraints, if any, are relegated in the last term \( J_c \). The derivation of Eqn. 6.3, the definition of matrices \( O, F \) and \( B \), as well as an illustration of the concrete use of the variational equation is set out in App. C. Eqn. 6.3 is used in 3D-VAR data assimilation procedures (the 3 space dimensions) and can be extended to 4D-VAR data assimilation by adding the time dimension:

\[
J(x(t_0)) = \frac{1}{2} (x_a - x_b) \cdot B^{-1} \cdot (x_a - x_b) + \\
\frac{1}{2} \sum_{t=0}^{N} (y_t(x) - \hat{y}_t) \cdot (O + F)^{-1} \cdot (y_t(x) - \hat{y}_t)
\]
where the observation $\tilde{y}$ and peer model $y(x)$ vectors are sampled at discrete times $t_i, i = 1, .., N$ over a predefined time window (usually one hour). Note the similarity of Eqs. 6.3 and 6.4. In this case, the background term $x_b$ is valid at time $t_0$ and summarizes the prior information. 4D-VAR methods can account for the evolution of the non-linear dynamic of NWP models. The value $x_n$ at time $t_0$ that minimizes $J$ in Eqn. 6.4 (without $J_c$) is given by:

$$0 = B^{-1} \cdot (x_n - x_b) + \sum_{t_i=1}^{N} \nabla y_{t_i}(x_n) \cdot (O + F)^{-1} \cdot (y_{t_i}(x_n) - \tilde{y}_{t_i})$$

(6.5)

The aim of the variational method is to minimize the penalty function $J$ given by Eqs. 6.3 and 6.4, i.e to minimize the differences between the observations and their model counterparts. The iterative procedure used to accomplish this task is now presented.

### 6.2 The Minimization Procedure

The major difficulty of the variational method resides in computing the gradient $\nabla y(x)$ needed by the minimization procedure. The dimension of the problem is usually large and the only way to solve it efficiently requires the adjoint equations of the NWP model (see le Dimet and Talagrand, 1986, for a simple example). The derivation of adjoint equations can be partly automated, but remains a tedious and non-trivial task, especially for “on-off” processes like convection (Zou, 1997).

On the other hand, the physical interpretation of the adjoint equations proved to be useful for the sensitivity analysis of atmospheric models (Hall and Cacuci, 1983) and to adjust dynamically the local characteristic constants of one-dimensional models (Marais and Musson-Genon, 1992).

The gradient $\nabla y(x)$ can also be obtained directly in finite differences by a perturbation of the initial state of the control variables. This is the method that is in effect used in this study. The derivative of the cost function $J$ (Eqn. 6.4, again without $J_c$) with respect to the component $x_j$ is given by:

$$\frac{\partial J(x(t_0))}{\partial x_j} = B^{-1} \cdot (x - x_b) + \sum_{t_i=1}^{N} \frac{\partial y_{t_i}(x)}{\partial x_j} \cdot (O + F)^{-1} \cdot (y_{t_i}(x) - \tilde{y}_{t_i})$$

(6.6)

The component $\partial x_j y(x)$ of the gradient is computed with help of a perturbation method. More precisely, the vector $x$ is slightly modified at time $t_0$ (the time at which the model is started) and then, the model is integrated to get the new values of $y_{t_i}(x), i = 1, .., N$ for the given perturbed initial conditions:

$$\sum_{i=1}^{N} \frac{\partial y_{t_i}(x(t_0))}{\partial x_j} = \sum_{i=1}^{N} \frac{y_{t_i}(x(t_0) + \Delta x e_j) - y_{t_i}(x(t_0))}{\Delta x}$$

(6.7)
\( N + 1 \) model integrations are necessary to compute the gradient effectively, \( N \) being the dimension of \( \mathbf{x} \): one computation for \( y(x(t_0)) \) and \( N \) for each component \( y(x(t_0) + \Delta x e_j) \). With Eqn. 6.7, the error on the derivative is of order \( O(\Delta x) \) and experimentation shows that this precision is sufficient. It is possible to compute the derivative with a lower error (of order \( O((\Delta x)^2) \)) by computing the additional value \( y(x(t_0) - \Delta x e_j) \), but this requires \( 2N + 1 \) computations and hence, increases the computational time.

### 6.2.1 The Quasi-Newton Method

Here, the minimization method used to compute the minimum of the penalty function is briefly described. Imagine that \( J \) can be locally approximated by the quadratic form:

\[
J(x) \approx J(x_i) + (x - x_i) \cdot \nabla J(x_i) + \frac{1}{2} (x - x_i) \cdot H \cdot (x - x_i)
\]

and its gradient

\[
\nabla J(x) = \nabla J(x_i) + H \cdot (x - x_i)
\]

where \( H \) is the Hessian matrix containing the second derivatives of \( J \). In Newton’s method, we set \( \nabla J(x) = 0 \) in Eqn. 6.9 (it is assumed that we are already at the minimum) to determine the next iteration point:

\[
x - x_i = -H^{-1} \cdot \nabla J(x_i)
\]

The left-hand side is the finite step we need to take to get the exact minimum. A minimization algorithm defined by Eqn. 6.10 is termed a Newton’s Method and its solution is called the Newton direction.

The quasi-Newton method is a subtle extension to Newton’s one. The idea behind it is to start with a positive definite approximation to \( H \) (usually the identity matrix) and build up an approximation of the Hessian matrix that remains positive definite and symmetric along the minimization procedure (Gill, 1980). Far from the minimum, this guarantees that we always move in a downhill direction and close to the minimum, the updating formula approaches the true Hessian (Press et al., 1989). Moreover, the inverse of the Hessian matrix is the covariance matrix of the recovered variables (Thacker, 1989).

For each minimization step, a line search and backtracking algorithm is performed along the Newtonian direction, because the full Newton step is not always adequate and has to be corrected accordingly.

Finally, the convergence of the minimization procedure is achieved if the following curvature criteria is satisfied:

\[
\nabla J \cdot \nabla J \leq \varepsilon^2 \cdot \max \{1, x \cdot x\}
\]
6.3 Method and Results

For the present set of experiments, a value $\epsilon = 0.8$ was found to be a reasonable setting. Here, use is made of the quasi-Newton algorithm with line minimization implemented by Shanno and Phua (1976, 1980).

6.3 Method and Results

All experiments with the penalty function were obtained with a simplified one-dimensional version of the SM, where the atmospheric forcing (e.g. pressure, wind, temperature and specific humidity) are specified with the output of the SM. All runs are started at 0 h UTC and are integrated for one day.

The simple penalty function following Mahfouf (1991) is a simplified version of Eqn. 6.4 and constitutes the starting point for our discussion:

$$J(x(t_0)) = \frac{1}{2} \sum_{t_i=1}^{N} \left[ \left( \frac{T_{t_i}(x) - \tilde{T}_{t_i}}{\sigma_T} \right)^2 + \left( \frac{RH_{t_i}(x) - \tilde{RH}_{t_i}}{\sigma_{RH}} \right)^2 \right]$$  \hfill (6.12)

where $T_{t_i}$ and $RH_{t_i}$ are the 2m temperature and 2m relative humidity at a given time $t_i$ respectively. $\sigma_T$ and $\sigma_{RH}$ are the standard deviations of the observations, typically 1 K and 10 % respectively. The soil moisture content of the two active layers was used as control variables in the experiments, i.e. $x(t_0) = (\theta_1(t_0), \theta_2(t_0))$, $t_0 = 0$ h UTC.

The 2m relative humidity is not an appropriate parameter, as it is already a combination of the 2m temperature and 2m dew point. It means that a part of the information, which is already contained in the 2m temperature, is duplicated in the relative humidity. Dew point temperature is also the variable that is usually observed by surface stations (ANETZ network included) and transformed afterwards into relative humidity values. Note that the method for obtaining a relative humidity from dew point and temperature is not unique and the same formula should be used to convert observation and model data. Dew point temperature has also the advantage of being directly connected to the specific humidity contained in the air and hence, is closer to evapotranspiration information.

6.3.1 Limitations

The 2m dew point temperature produced by the SM should be used with great care. Recall that this parameter is not computed with the interpolation scheme from Geleyn (1988), but with a simplified procedure, which use the specific humidity at 30 m and the 2m temperature, and unfortunately gives poor results. Hence this variable is not verified with observation either at SMI nor at DWD. In Fig. 6.1, an example of the screen-level variables at the grid-point Büel, as represented by the SM (output every time step) on July 22, and July 8, 1996. It was already noticed, that the radiative scheme is only computed once per hour during the integration of the SM, and this has some consequence on the energy balance at the ground (two of the energy sources are fixed for an hour, i.e. shortwave and longwave radiation).
This in turn adds uncertainty in the behavior of screen-level variables, because the ground temperature $T_B$ is never in equilibrium between two radiation changes. This is the origin of the spikes that are seen at midday on the 2m temperature curve.

There are two curves for the dew point temperature: one computed with the default method, the other one following Geleyn (1988). When the atmosphere has the most influence on screen-level variables (July 8), the two methods give similar answers. But when the soil is driving dew point temperature, these methods give distinctively different results, especially during the night. The non-standard method also uses the fictitious surface specific humidity as a bottom boundary condition, and this depends directly on the computed evapotranspiration flux (see Eqn. 2.27, pp. 17). The difficulty is to define the surface specific humidity in such a way as to obtain optimum results. Hence, these observations question the usefulness of the 2m dew point temperature in our penalty function.

### 6.3.2 2m Temperature Maximum

As a consequence, the second parameter was dropped from the formulation of our penalty function. To account also for possible shifts between modeled and observed temperature curves, we kept only the 2m temperature maximum, which is well defined in fair weather. Consequently the simplest form of our penalty function (PF) reads:

$$ J_0(x(t_0)) = \frac{1}{2} \left( \frac{\max T_{2m}(x) - \max \hat{T}_{2m}}{\sigma_{T2m}} \right)^2 $$  \hspace{1cm} (6.13)

The minimization of the penalty function $J$ will lead to new soil moisture initial conditions $\theta_1$ and $\theta_2$ at 0 h UTC designed to improve the 2m temperature maximum.
6.3 Method and Results

Fig. 6.2: Normalized penalty functions maps \( J / \max \{ J \} \) for the function \( J_0 \) (a) and \( J_0 + J_b \) with the \( \sigma_p^2 \) values set to 0.25 (c & d), 0.09 (e & f) and 0.01 (b). The trajectories of the minimization procedure for three arbitrary points are also shown.
Fig. 6.3: Normalized penalty functions maps ($J / \max \{J\}$) for the function $J_1$ (a) and $J_1 + J_2$ with the $\sigma_J^2$ values set to 0.25 (c), 0.09 (e & f) and 0.01 (d). The trajectories of the minimization procedure for three arbitrary points are also shown.
6.3 Method and Results

forecast. (Note that \( \sigma_T = 1 \text{ K.} \))

An example of this function is represented in Fig. 6.2a. The overall function shape is a curved valley without a clear minimum. This means that the lowest part of this valley is the common location in the parameter space of soil moisture initial conditions that minimize \( J \). This behavior is directly connected to the evapotranspiration scheme, as described in the previous chapter. The minimization procedure was started from three distinct initial conditions resulting in three different solutions. This formulation of the PF is therefore very sensitive to the initial conditions. The retrieved soil moisture configurations reduce the forecast errors, but are sometimes completely artificial (e.g. \( \theta_2 < \theta_{wp} \) and \( \theta_1 \) close to saturation). This is clearly not acceptable. Further experiments have been done with the addition of the relative humidity part and give analogous results (not shown).

This is the main reason why a background term \( J_b \) is necessary to the penalty function \( J \):

\[
J(x(t_0)) = J_0(x) + J_b(x)
\]

\[
J_b(x(t_0)) = \frac{1}{2} (x - x_b) \cdot B^{-1} \cdot (x - x_b)
\]  

(6.14)

The benefit of this additional term is to give the PE a unique minimum and to retrieve a reasonable configuration of the soil water content. The penalty is increased with soil moisture conditions that are too distant from the initial first guess. The major difficulty reside in the setting of adequate values of the covariance matrix \( B \). We used a diagonal matrix with constant elements:

\[
B = \sigma_{\theta'}^2 \cdot \begin{pmatrix} 1 & 0 \\ 0 & 1 \end{pmatrix}
\]  

(6.15)

Several samples of the PE with different values are shown on Figs. 6.2c,e,b, with \( \sigma_{\theta'}^2 \) values of 0.25, 0.09 and 0.01 respectively. For the minimization algorithm, the normalized soil moisture \( \theta' = \theta/\theta_{sat} \) is used instead of \( \theta \), to have a symmetric background term \( J_b \). Otherwise, without this normalization, the penalty function shape looks like a disymmetric valley, due to the different thicknesses of soil moisture layers. This definitively non optimal for the minimization algorithm. Experiments with \( \sigma_{\theta'}^2 \) values of 0.25 and 0.09 give the best results and produce a new 2m temperature curve maxima that agree with observations (see Figs. 6.2d,f). The experiment with \( \sigma_{\theta'}^2 = 0.01 \) give poor results, because the background term restrain the soil moisture changes too strongly and improvements on the 2m temperature are barely noticeable (not shown).

This penalty function is only able to correct the temperature maximum. The temperature minima is also indirectly modified, but is affected to a much lesser extent. This is related to the constancy of the heat capacity and conductivity during the integration. To have a control over the 2m temperature minima, the deep soil temperature should be modified also. This experiment has been attempted but gave
poor results (not shown). This was already obvious from the sensitivity experiments, where a change of the soil temperature profile at the beginning of the integration affect the screen-level variables only before sunrise and after sunset, but not the 2m temperature maximum.

The number of steps required to minimize the penalty function ranges from 2 to 7 for these experiments, but the number of times the gradient is evaluated can be higher, as the line search algorithm tries sometimes several points to find the adequate minimization step. In these experiments, the maximal number of evaluations was 10, but this value is function of the requested precision (parameter \( \epsilon \) in Eqn. 6.11). A value \( \Delta x = 0.01 \) was also used to compute the gradient of the penalty function. The number of minimization step can be reduced by starting from the first guess value, which is already close to the function’s minimum.

6.3.3 Daily Evapotranspiration Sum

With the availability of supplementary data at Büel, another way for determining the consistency of the land-surface scheme used in the SM is possible. The daily evapotranspiration sum can be used for the penalty function, instead of the 2m temperature maximum:

\[
J_1(x(t_0)) = \frac{1}{2} \left( \mu \sum_{i=1}^{24} E_i(x) - \bar{E}_i \right)^2
\]  

(6.16)

The results of the experiments conducted with \( J_1 \) are shown in Figs. 6.3. Without a background term, the shape of the penalty function is a valley (Fig.6.3a) and is similar to the results obtained with the penalty function using temperature information. This demonstrates that the origin of that shape is strongly connected to the evapotranspiration patterns. This is confirmed by the analysis of the evapotranspiration as a function of the soil moisture content (previous chapter) that has shown similar results.

By taking the daily evapotranspiration sum, the cost function is independent of the possible temporal shift between the observed and modeled evapotranspiration curves. This argument is supported by the comparisons with the observations (c.f. Chap 3). The true minimum of functions \( J_0 \) and \( J_1 \) are quite different: a different soil initial condition will result from the minimization procedure. This has different consequences on the final 2m temperature curve (see Figs.6.3b,f). Although the daily evapotranspiration sum has been corrected, the improvements on the 2m temperature are weak, but are in the right direction (i.e. reduction of the temperature maxima). This indicates that there is a soil moisture deficit at this grid-point. The analysis is difficult since the 2m temperature amplitude between model and observation curves differ of approximately 2° C. Consequently, the penalty function \( J_0 \) is more indicated to correct screen-level variables.

A one-dimensional model tuned for each specific grid-point would surely help to refine the assimilation method. This is the way to obtain better results, especially in the Alps, where the model 2m temperature amplitude is much too large compared to
the observations and should be corrected accordingly, before applying the variational scheme.

6.4 Discussion

In the present scheme, the first guess, or background term $x_b$, is the soil moisture value of the initial conditions of the SM, and its intrinsic accuracy is rather unknown. With the variational method, both soil moisture layers values are changed, but we have no prior knowledge to decide, which layer has the larger bias. This distinction between layers is essential: the capacity in mm of $\theta_2$ is about ten times larger than $\theta_1$, and a change of 10 mm has a quite different effects according to which layer this quantity is added or removed. The size of layers defines also the timescale, upon which a change in soil moisture affects the screen-level variables. With these considerations, it would be sufficient to correct the second soil moisture only weekly, and to add minor corrections to the first soil moisture on a day to day basis. It would be interesting to study more cases of sunny days and build a statistic with sensitivity studies, but with small variations of soil moisture. Statistical information could be set up this way, depending for example on height and soil type. This information could be used to correct the initial the first guess, before applying the variational method.

For July 8, the variational method with a penalty function specified with Eqn. 6.13 is unusable, as there is no clear 2m temperature maximum. A misprediction of cloudiness in the model could also lead to a non-realistic soil moisture retrieval, because the shortwave radiation has a great impact on the intensity of sensible and latent heat fluxes. To remedy to this problem, the observed shortwave radiation should be specified as the radiative forcing in the one-dimensional model.

The recent soil moisture assimilation schemes are based on sequential assimilation, because they are much more economic. At ECMWF, the relative humidity increments during the normal assimilation cycle are then converted into soil moisture equivalents (Mahfouf and Viterbo, 1996). This ad-hoc method has been replaced with an improved sequential assimilation scheme.
Chapter 7

Conclusion

Close inspection of the SVAT scheme utilized in the SM revealed some inherent limitations of that parameterization scheme. For computational speed and computer memory storage efficiency, the EFR-method is used to compute ground heat fluxes, instead of a multiple soil layer scheme. Unfortunately, this approach requires constant heat conductivity and capacity. Grid-points with identical soil types share consequently the same heat conductivity and capacity, and this constitutes a constrain for the ground temperature response to radiative forcing. Thus, the model is not able to differentiate between dry or saturated soil conditions and as a consequence, the temperature amplitude and minima are affected. In addition to this major limitation, the analysis of the soil model shows that heat capacity is overestimated and that vegetation parameterization is somewhat artificial, the rooting depth being dependent on time. Also for example, no vegetation type is defined to distinguish forests from grasslands.

Comparisons of screen-level variables, potential evaporation and soil temperature were performed with observations from the ANETZ network for Summer 1995/96. Additional comparisons of soil moisture and evapotranspiration from the hydrological station Biel are available for the same period. It is difficult to compare the 2m relative humidity provided by the SM, as the scheme used to compute this variable is much too simple and produces poor results, especially during the night. In relation to the 2m temperature, comparisons show that its maxima occurs 1–2 hours earlier than for observations and that the 2m temperature amplitude is largely overestimated at altitudes higher than 1500 m. The comparison and reduction of screen-level variables to the height of the observing stations appear difficult in mountainous regions, as it not unusual to find differences of more than a few hundred meters between stations and SM peer points.

Sensitivity studies were carried out in a small domain comprising Switzerland to explore the impacts of distinct soil model parameter changes upon soil-atmosphere energy fluxes and screen-level variables. Soil moisture content, vegetation cover fraction and rooting depth are the relevant model parameters controlling screen-level variables during fair weather. These parameters are able to modify the surface evapotranspiration and consequently the Bowen-ratio. In contrast, vegetation parameters modify the partitioning between bare soil evaporation and transpiration of
plants. Nevertheless, the evapotranspiration rate is affected, as bare soil can evaporate at a lower value determined by the rooting depth. Experiments were undertaken with albedo, roughness length and temperature profile changes, and it was shown that they affect screen-level variables to a much lesser extent. During cloudy and rainy conditions, the surface fluxes are weak and screen-level variables are mostly controlled by the atmosphere and are not very sensitive to soil parameter changes.

The analysis of the evapotranspiration with respect to altitude shows that the surface soil moisture and the parameterization of bare soil evaporation appear to be decisive at altitudes higher than 1500 m where vegetation is sparse. At lower altitudes, deep soil moisture and vegetation transpiration gain importance, but bare soil evaporation still remains relevant to evapotranspiration. It was also shown that run-off becomes a relevant process during precipitation events and when the surface soil moisture is close to saturation. Especially during strong rain events, a large part of the precipitation is lost by the model in form of run-off, making the moisture refill of the soil layers difficult. Otherwise, run-off is negligible in the hydrological processes during fair weather.

It was pointed out that the quality of the 2m relative humidity in the SM is poor, especially during the night. Moreover, this variable is computed from the 2m temperature and dew point and thus, duplicates the temperature information. Based on this consideration, the use of dew temperature was considered for the variational method, instead of relative humidity. But the computation of relative humidity following two different methods revealed again important discrepancies. Finally, for the formulation of the penalty function used by the variational method, only the 2m temperature maximum was used, to avoid problems caused by temporal shifts between modeled and observed temperature curves. Unfortunately, there is no clear minimum, and different initial conditions for the minimization algorithm produce different results and moreover, some of them are not realistic. The addition of a background term, i.e. the soil moisture first guess information, proved necessary to obtain a distinct function minimum (this is consistent with different soil moisture configurations producing similar evapotranspiration rates). The correction of the temperature minimum remains difficult, due to the limitation of the current soil model, and the adopted method is limited to fair weather situations.

7.1 Perspectives

Improvement of screen-level forecasts are inseparable from the refinement of SVAT schemes and PBL parameterizations. For instance, DWD has planned for its next generation “Lokal Model” (LM), a high resolution non-hydrostatic LAM, to replace the old soil model with the more sophisticated AMBETI scheme developed by Braden (1995). The latter derives from the field of agricultural research and has been widely tested with observations, in contrast to the current SVAT scheme. Nevertheless, the parameterizations contained in AMBETI are far too complex for current NWP models and need simplifications before an integration in the LM. Use of turbulent kinetic energy equations is also planned for the PBL.
Although the resolution of NWP models is increasing (the LM is expected to run at $\sim 2$ km), it has not been proven yet that a higher resolution will improve screen-level variables, as well as sensible and latent heat fluxes in mountainous regions. High resolution external parameter datasets are also necessary to describe surface characteristics with more details.
Appendix A

The Force-restore Method

The force-restore (FR) method was independently proposed by Bhumralkar (1975) and Blackadar (1976) for predicting ground surface temperature and has been subsequently generalized (Dickinson, 1988). The performance of the FR-method, compared with multiple soil layer models or simpler schemes, was reviewed in Deardorff (1978) and Jacobsen and Heise (1982).

A.1 Derivation

Assuming that there are no other sources or sinks of heat in the soil, the second law of thermodynamics yields the simple prognostic equation for the soil heat transfer:

\[
\frac{\partial T}{\partial t} = \frac{\partial G}{\partial z}
\]  

(A.1)

Since molecular conduction is the primary transport process, we can write the ground heat flux \( G \) at any depth \( z \) as:

\[
G = -\lambda \frac{\partial T}{\partial z}
\]  

(A.2)

As a representation of the diurnal thermal cycle, we consider a single harmonic wave function of frequency \( \omega \) and amplitude \( \Delta T \) for the ground temperature \( T \):

\[
T = \bar{T} + \Delta T e^{i(kz - \omega t)}
\]  

(A.3)

Now assume a constant heat capacity \( (\rho C) \) and a constant heat conductivity \( \lambda \). Then, on substituting Eqs. A.2 and A.3 into Eqn. A.1, we obtain an expression for the wave number \( k \) which satisfies the heat conduction equation. Finally, replacing \( k \) into Eqn. A.3 and taking the real part of the solution yields:

\[
T = \bar{T} + \Delta T e^{-z/d} \cos \left( \frac{z}{d} - \omega t \right)
\]  

(A.4)
This solution is a periodic temperature variation that decreases in amplitude and is increasing phase lag with depth. Here, \( d = \sqrt{2\lambda/(\rho C \omega)} \) is the e-folding depth for which the amplitude of the temperature wave penetration into the ground is damped by a factor \( e^{-1} \). The preliminary computation of \( \partial_t T \) (with \( T \) given by Eqn. A.4) is helpful to find that substituting Eqn. A.4 into Eqn. A.2 yields finally the FR-approximation of the heat diffusion equation:

\[
\frac{\partial T}{\partial t} = \frac{2}{(\rho C)d} G - \omega(T - \overline{T})
\]

The forcing term \( 2G/(\rho Cd) \) is driven by the surface energy fluxes \( G \), whereas the restore term \( \omega(T - \overline{T}) \) is controlled by the mean soil temperature \( \overline{T} \). Often, a second restore-like prognostic equation for \( T \) is coupled with Eqn. A.5 to simulate the annual variation of the deep soil temperature in GCMs. A first attempt of this kind was made by Deardorff (1978), who also extended this method to soil moisture exchanges.

A.2 Outline of the Extended FR-method

A better approximation of Eqn. A.1 than the FR-method can be achieved by increasing the number of tunable coefficients and temperature variables. This is effectively the option considered for extending the FR-method, hence its name.

The diurnal forcing is now represented by two harmonic wave functions of frequency \( \omega_1 \) and \( \omega_2 \). The ground temperature variables are two in number as well and their evolution is given by FR-like prognostic equations:

\[
\begin{align*}
\frac{\partial T_B}{\partial t} &= -\alpha_1 T_B + \beta_1 T_M + \gamma G \\
\frac{\partial T_M}{\partial t} &= -\alpha_2 T_M + \beta_2 T_B
\end{align*}
\]

where \( \alpha, \beta \) and \( \gamma \) are tunable coefficients to be determined and \( G \) is the term including the surface energy fluxes. Note that both equations now contain a coupling term \( \beta T \).

For a periodic forcing \( G = \Delta Ge^{-i\omega t} \), the solution of Eqs. A.6 is of the form \( T = \Delta T' e^{i(\psi' - \omega t)} \) and its accuracy can be estimated by comparing the ratio of the amplitude reactions \( A = |\Delta T'|/|\Delta T| \) and phase differences \( \tan(\psi' - \psi) \) as a function of the frequency \( \omega \). This analysis method provides four of the five conditions that are necessary to determine all coefficients. The conditions to be fulfilled are \( A = 1 \) and \( \psi' = \psi \) for both forcing frequencies \( \omega = \omega_1 \) and \( \omega = \omega_2 \). With such prerequisite, Eqs. A.6 give an optimal answer when the external forcing match these frequencies. The complete description of the analysis method can be found in Jacobsen and Heise (1982).

Consider now the two-layer approximation of the heat conduction equation:
A.2 Outline of the Extended FR-method

\[
\begin{align*}
(rC) \Delta z_B \frac{\partial T_1}{\partial t} & = G - G_M \\
(rC) \Delta z_M \frac{\partial T_2}{\partial t} & = G_M - G_U
\end{align*}
\]  

(A.7)

where $\Delta z_{B,M}$ are the thicknesses of the two soil layers and $G_{M,U}$ are the ground heat fluxes at their boundaries (see Fig. 2.2, pp. 9). Assuming linear temperature profiles for the mean soil layer temperatures, i.e. $T_1 = \frac{1}{2}(T_B + T_M)$ and $T_2 = \frac{1}{2}(T_M + T_U)$, it can be shown that Eqs. A.7 are fully equivalent to Eqs. A.6 if $G_{M,U}$ and $z_{B,M}$ are defined in a consistent manner:

\[
\begin{align*}
G_M &= \frac{\lambda}{D_1^2} \Delta z_2 \left( (1 + x + x^2) \frac{\Delta z_1}{\Delta z_1 + \Delta z_2} (T_B - T_U) - x(T_M - T_U) \right) - G_U \\
G_U &= \frac{\lambda}{D_2^2} \frac{\Delta z_1 + \Delta z_2}{1 + x + x^2} (T_M - T_U)
\end{align*}
\]  

(A.8)

and

\[
\begin{align*}
\Delta z_1 &= \frac{D_1}{1 + x} \\
\Delta z_2 &= \Delta z_1 \left( \frac{1 + x + x^2}{x\sqrt{x + x^2} e^{\frac{x}{x + x^2}} - 1} \right)
\end{align*}
\]  

(A.9)

where $D_{1,2} = \sqrt{2\lambda/(rC\omega_{1,2})}$ and $x^2 = \omega_2/\omega_1$. As a fifth condition, we required that the amplitude of $T_M$ should be identical with that of $T$ at a depth of $z = -\Delta z_1$ with respect to the frequency $\omega_2$. This condition has no influence on the frequency dependence of amplitude and phase error and establishes $T_M$ as a good approximation of the exact temperature $T$ at that depth. The final form of Eqs. A.8 (see Eqs. 2.4, pp. 11) are obtained with the settings $\tau_1 = 2\pi/\omega_1 = 1$ day and $\tau_2 = 5\tau_1$. 
Appendix B

External Parameters

External parameters such as surface types, soil textures, albedo and vegetation are central to every SVAT scheme for describing the characteristics of the earth's surface. Each surface type and soil texture has distinct thermal and hydrological properties and thus, behaves differently in response to external forcing. The description of vegetation is essential to the parameterization of transpiration. Important hydrological terms used throughout this study are also defined and co-located here for convenience. This information can also be found in Schrodin and al. (1996).

B.1 Soil Types and Textures

Three surface types (i.e. rock, ice and water) and six soil textures (i.e. sand, sandy loam, loam, clay loam, clay and peat) are available to specify the surface land form in the SM. Note that ice and rock surface types are currently not used, although they correspond quite well to the description of the highest Alpine regions. The hydrological and thermal properties for each type and texture are listed in Tables B.1 and B.2 respectively.

B.2 Albedo

Surface albedo is specified as a combination of a constant vegetation albedo $\alpha_v = 0.15$, a dry soil albedo $\alpha_o$ and a linear albedo term $\Delta\alpha$ related to the soil surface moisture $\theta_1$:

$$\alpha = \sigma \alpha_v + (1 - \sigma)(\alpha_o - \Delta\alpha \cdot \theta_1) \quad (B.1)$$

The albedo components are weighted by the vegetation cover fraction $\sigma$. Both $\alpha_o$ and $\Delta\alpha$ values are soil type/texture dependent parameters (see Table B.2). Note that the linear relationship between albedo and soil moisture applies only to the first millimeters of soil (Idso et al., 1975), whereas $\theta_1$ accounts for the soil moisture contained in the upper soil layer, which is much deeper ($\Delta z_1 = 10 \text{ cm}$).

The albedo is larger when the sun is lower in the sky and thus, Eqn. B.1 should include an extra dependence on the sun zenith angle (Pielke, 1984).
TABLE B.1: Soil hydrological properties. $\theta_{sat}$ is the porosity, $\theta_{fc}$ is the field capacity, $\theta_{wp}$ is the wilting point and $\theta_{adp}$ is the air dryness point. All have units of $10^{-2}$ m$^3$ m$^{-3}$. $D_{sat}$ and $K_{sat}$ are the hydraulic diffusivity ($10^{-9}$ m$^2$ s$^{-1}$) and conductivity ($10^{-9}$ m s$^{-1}$) at saturation respectively. $D_1$ and $K_1$ are dimensionless hydraulic diffusion and conduction coefficients respectively. $I_{min}$ is the minimal soil infiltration rate ($10^{-4}$ kg m$^{-2}$ s$^{-1}$).

<table>
<thead>
<tr>
<th>soil types</th>
<th>$\theta_{sat}$</th>
<th>$\theta_{fc}$</th>
<th>$\theta_{wp}$</th>
<th>$\theta_{adp}$</th>
<th>$D_{sat}$</th>
<th>$D_1$</th>
<th>$K_{sat}$</th>
<th>$K_1$</th>
<th>$I_{min}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>sand</td>
<td>36.4</td>
<td>19.6</td>
<td>4.2</td>
<td>1.2</td>
<td>18400</td>
<td>-8.45</td>
<td>47900</td>
<td>-19.27</td>
<td>35</td>
</tr>
<tr>
<td>sandy loam</td>
<td>44.5</td>
<td>26.0</td>
<td>10.0</td>
<td>3.0</td>
<td>34600</td>
<td>-9.47</td>
<td>94300</td>
<td>-20.86</td>
<td>23</td>
</tr>
<tr>
<td>loam</td>
<td>45.5</td>
<td>34.0</td>
<td>11.0</td>
<td>3.5</td>
<td>35700</td>
<td>-7.44</td>
<td>53100</td>
<td>-19.66</td>
<td>10</td>
</tr>
<tr>
<td>clay loam</td>
<td>47.5</td>
<td>37.0</td>
<td>18.5</td>
<td>6.0</td>
<td>11800</td>
<td>-7.76</td>
<td>76400</td>
<td>-18.52</td>
<td>6</td>
</tr>
<tr>
<td>clay</td>
<td>50.7</td>
<td>46.3</td>
<td>25.7</td>
<td>6.5</td>
<td>442</td>
<td>-6.74</td>
<td>17</td>
<td>-16.32</td>
<td>1</td>
</tr>
<tr>
<td>peat</td>
<td>86.3</td>
<td>76.3</td>
<td>26.5</td>
<td>9.8</td>
<td>106</td>
<td>-5.97</td>
<td>58</td>
<td>-16.48</td>
<td>2</td>
</tr>
</tbody>
</table>

TABLE B.2: Soil thermal properties. $(\rho C)_0$ is the dry heat capacity ($10^6$ J K$^{-1}$ m$^{-3}$). $\lambda_0$ is the dry thermal conductivity and $\Delta \lambda$ is a thermal conductivity coefficient (W K$^{-1}$ m$^{-1}$). $\alpha_0$ is the dry bare soil albedo and $\Delta \alpha$ is an albedo coefficient (%). $\Delta z_B$ and $\Delta z_M$ are the thicknesses of the two soil layers used by the EFR-method (cm). Note that the water temperature is constant in the SM.

<table>
<thead>
<tr>
<th>soil types</th>
<th>$(\rho C)_0$</th>
<th>$\lambda_0$</th>
<th>$\Delta \lambda$</th>
<th>$\alpha_0$</th>
<th>$\Delta \alpha$</th>
<th>$\Delta z_B$</th>
<th>$\Delta z_M$</th>
</tr>
</thead>
<tbody>
<tr>
<td>sand</td>
<td>1.28</td>
<td>0.30</td>
<td>2.40</td>
<td>30</td>
<td>44</td>
<td>10.34</td>
<td>37.02</td>
</tr>
<tr>
<td>sandy loam</td>
<td>1.35</td>
<td>0.28</td>
<td>2.40</td>
<td>25</td>
<td>27</td>
<td>9.84</td>
<td>35.23</td>
</tr>
<tr>
<td>loam</td>
<td>1.42</td>
<td>0.25</td>
<td>1.58</td>
<td>25</td>
<td>24</td>
<td>7.78</td>
<td>27.83</td>
</tr>
<tr>
<td>clay loam</td>
<td>1.50</td>
<td>0.21</td>
<td>1.55</td>
<td>25</td>
<td>23</td>
<td>7.51</td>
<td>26.88</td>
</tr>
<tr>
<td>clay</td>
<td>1.63</td>
<td>0.18</td>
<td>1.50</td>
<td>25</td>
<td>22</td>
<td>7.15</td>
<td>25.58</td>
</tr>
<tr>
<td>peat</td>
<td>0.58</td>
<td>0.06</td>
<td>0.50</td>
<td>20</td>
<td>10</td>
<td>4.08</td>
<td>14.60</td>
</tr>
<tr>
<td>rock</td>
<td>2.10</td>
<td>2.41</td>
<td>-</td>
<td>30</td>
<td>-</td>
<td>12.28</td>
<td>43.95</td>
</tr>
<tr>
<td>ice</td>
<td>1.92</td>
<td>2.26</td>
<td>-</td>
<td>70</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>water</td>
<td>4.18</td>
<td>1.00</td>
<td>-</td>
<td>10</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

B.3 Vegetation

Vegetation is described by a rooting depth $d_r$ and a vegetation cover fraction $\sigma$. Both fields are parameterized with a generic function $X$ depending on the latitude $\varphi$, the day of the year $D$ and the geopotential $\Phi = g \cdot h$:

$$X(\varphi, D, \Phi) = X_{\min} + (f_v(\varphi, D) \cdot f_h(\Phi))^\alpha \cdot (X_{\max} - X_{\min})$$  \hspace{1cm} (B.2)

where $X_{\min}$ and $X_{\max}$ are constant values, $g$ is the acceleration due to gravity and $h$ is the orography. For the vegetation cover fraction, $\alpha = 1$ and $X_{\min,\max}$ are set on a grid-point to grid-point basis according to vegetation maps. Minimum and maximum values correspond to the state of vegetation at rest (Winter) and
Fig. B.1: Vegetation parameters. Yearly evolution of the vegetation cover fraction \( \sigma \) (a) and rooting depth \( d_r \) (b) parameters for the following values: \( \varphi = 47^\circ \), \( \sigma_{\text{min}} = 0.1 \, \text{m}^2 \, \text{m}^{-2} \), \( \sigma_{\text{max}} = 0.9 \, \text{m}^2 \, \text{m}^{-2} \), \( d_{r,\text{min}} = 0.12 \, \text{m} \) and \( d_{r,\text{max}} = 0.7 \, \text{m} \). The specified altitudes are \( z = 800 \, \text{m} \) (solid line) and \( z = 1500 \, \text{m} \) (dotted line). Bold lines emphasize the summertime period.

At full growth (Summer) respectively. Note that no vegetation type is defined in the SM. For the rooting depth, \( \alpha = 2 \) and the parameters \( X_{\text{min}} = 0.12 \, \text{m} \) and \( X_{\text{max}} = 0.7 \, \text{m} \) are shared by all grid-points. With such a minimum value, the first soil layer (\( \Delta z_1 = 10 \, \text{cm} \)) contains always the same rooting length (10 cm), whereas the second layer (\( \Delta z_2 = 90 \, \text{cm} \)) contains rooting lengths ranging from 2 to 60 cm. This operational configuration limits the vegetation transpiration to \( 0.7 E_{\text{pot}} \) (c.f Eqn. 2.21, pp. 16). \( f_v \) and \( f_h \) take the following form:

\[
\begin{align*}
  f_v(\varphi, D) & = \max\left\{ 0, \min\left\{ 1, c \cdot \sin \left( \pi \cdot \max\left\{ 0, \frac{D - D_v(\varphi)}{D_r(\varphi)} \right\} \right) \right\} \right\} \\
  f_h(\Phi) & = e^{-5 \cdot 10^{-5} \Phi^2}
\end{align*}
\]  

(B.3)

With these settings, the value of \( f_h \) is already reduced by half at the altitude \( z = 1200 \, \text{m} \). Note that the rooting depth (corresponding to a higher \( \alpha \) value) is more sensitive to height than the vegetation cover fraction. \( f_v \) depends in turn on the starting day \( D_v \) and the length \( D_r \) in days of the active vegetation period:

\[
\begin{align*}
  D_v(\varphi) & = \max\{1, 3(|\varphi| - 20^\circ)\} \\
  D_r(\varphi) & = \min\{365, 345 - 4.5(|\varphi| - 20^\circ)\}
\end{align*}
\]  

(B.4)

The coefficient \( c = 1.12 \) is set in such a manner as to obtain \( f_v = 1 \) for 30% of the active vegetation period. At the equator, \( D_v = 1 \) day and \( D_r = 365 \) days, at the mean latitude of Switzerland (\( \varphi = 47^\circ \)), \( D_v = 81 \) days and \( D_r = 223.5 \) days, and at south/north poles, \( D_v = 210 \) days and \( D_r = 30 \) days. Fig. B.1 provides a concrete graphical example.
With this parameterization, the start and length of the vegetation period are, unlike observations, independent of the altitude. Moreover, the advance or delay of the vegetation growth from year to year is not taken into account. The rooting depth also has a spurious time dependence and is only designed as a simple transpiration regulator throughout the year. Clearly, this unaesthetic simple solution is a poor substitute for physically sound parameterizations using LAI information.

Satellite based imagery could possibly provide a remedy for these problems (e.g. with LAI and vegetation cover fraction fields on a regular basis) and thereby improve the parameterization for NWP models. However, the specification of such a simple seasonal cycle is somewhat questionable (Heck et al., 1998).

B.4 Hydrological Terms

A number of threshold values corresponding to specific soil hydrological states are used for the parameterization of evapotranspiration and are defined below.

The porosity $\theta_{\text{sat}}$ is the fraction of soil occupied by the soil pores, or volume of air trapped over total volume. Indeed, porosity coincides with the total amount of water that can be held by the soil in its pores (i.e. the soil saturation wetness). The field capacity $\theta_{\text{fc}}$ is the maximum amount of water an entire column of soil can hold against gravity, or the equilibrium mean value of a column soil water 24–48 hours after wetting the soil (Hillel, 1980). The permanent wilting point $\theta_{\text{wp}}$ is the value at which plants can no longer recover turgidity (i.e. their normal state), even when placed in a saturated environment. The air dryness point $\theta_{\text{adp}}$ is the value of soil water at which direct evaporation ceases. Note that this last parameter is seldom used in soil model formulations. The volumetric available water holding capacity, or availability, is defined as $\theta_{\text{ava}} = \theta_{\text{fc}} - \theta_{\text{wp}}$. This value determines the size of the reservoir and is crucial for long drying periods.
Appendix C

The Variational Analysis Equation

The variational analysis equation is that used in 3/4D-VAR assimilation methods and can be derived using Bayesian probabilistic arguments. This is one of a multiple of approaches for finding the best analysis schemes for NWP models (Lorenc, 1986).

C.1 Derivation

Baye's theorem states that the posterior probability of an event $A$ occurring given that event $B$ is known to have occurred can be written as:

$$P(A|B) \propto P(A) \cdot P(B|A)$$  \hspace{1cm} (C.1)

where $P(B|A)$ is the probability of $B$ occurring given that $A$ is known to have occurred. For meteorological purposes, we replace $A$ and $B$ in Eqn. C.1 with the model state vector $x$ (dim $N_x$) and observed data vector $y$ (dim $N_y$) respectively:

$$P(x = x_t|y = y_o) \propto P(x = x_t) \cdot P(y = y_o|x = x_t)$$  \hspace{1cm} (C.2)

where the indices $t$ and $o$ denote the true and observed value respectively. The prior probability $P(x = x_t)$ encapsulates the knowledge about $x$ before the observation are taken. We assume that it can be written in terms of deviations from some known background term $x_b$:

$$P(x = x_t) = P_b(x - x_b)$$  \hspace{1cm} (C.3)

Any physical observation is subject to random observational errors. If the true values of $y$ are given by $y = y_1$, we assume that these errors can be expressed as:

$$P(y = y_o|y_1 = y_t) = P_o(y_o - y_t)$$  \hspace{1cm} (C.4)

We need to know $P(y = y_o|x = x_0)$, which we shall write as:
\[ P_{of}(y_o - y_f) = P_o(y_o - y_f) \]  

where the index \( f \) denotes the interpolated value. For the rest of the discussion, we assume for simplification that \( y_f = y_t \) (i.e. exact interpolation). With \( P(A) = \int P(A|B) \cdot P(B) \, dB \), where \( B \) is the event that some parameter has a value between \( b \) and \( b + db \), we have:

\[ P(y = y_o| x = x_t) = \int P(y = y_o| x = x_t \text{ and } y_1 = y_t) \cdot P(y_1 = y_t| x = x_t) \, dy_1 \]

\[ P_{of}(y_o - y_f) = \int P_o(y_1 - y_o) \cdot P_f(y_1 - y_f) \, dy_1 \]  

Substitution into Eqn. C.2 gives:

\[ P(x) \propto P_b(x - x_b) \cdot \int P_o(y_1 - y_o) \cdot P_f(y_1 - y_f) \, dy_1 \]  

We assumed that \( P_b(x - x_b) \) and \( P_o(y_o - y_f) \) are independent, i.e. that background errors and observational errors are uncorrelated. The use of Gaussian functions to describe the probability distribution functions (PDFs) is a common assumption and simplifies the solution:

\[ P_b(x - x_b) \propto e^{-\frac{1}{2}(x-x_b)^T B^{-1} (x-x_b)} \]

\[ P_o(y - y_o) \propto e^{-\frac{1}{2}(y-y_o)^T O^{-1} (y-y_o)} \]

\[ P_f(y - y_f) \propto e^{-\frac{1}{2}(y-y_f)^T F^{-1} (y-y_f)} \]  

where the matrices \( O, F \) and \( B \) can be shown to be covariance matrices for Gaussian PDFs:

\[ O = (y_t - y_o) \cdot (y_t - y_o)^T \]

\[ F = (y_t - y_f) \cdot (y_t - y_f)^T \]

\[ B = (x_t - x_b) \cdot (x_t - x_b)^T \]  

Replacing \( O \) and \( F \) given by Eqs. C.8 into Eqn. C.6 results in:

\[ P_{of}(y_o - y_f) \propto \cdot e^{-\frac{1}{2} \{(y_f - y_o)^T (O + F)^{-1} (y_f - y_o)\}} \]  

Finally, for the maximum likelihood estimates, we want to maximize \( P \) given by Eqn. C.7, which is equivalent to minimizing \(-\ln(P)\):

\[ J(x) = \frac{1}{2} \left\{ (y(x) - y_o)^T (O + F)^{-1} (y(x) - y_o) + (x - x_b)^T B^{-1} (x - x_b) \right\} \]  

Eqn. C.11 is called cost or penalty function. A comprehensive review of analysis methods in NWP can be found in Lorenc (1986).
C.2 Illustration

The 1D-VAR variational method used by Eyre et al. (1993) for the assimilation of satellite radiance is given here for illustrative and comparative purposes.

The model state vector $x$ is the vertical atmospheric profile containing temperature at 40 pressure levels (1000 to 0.1 hPa), humidity at 15 pressure levels (1000 to 300 hPa, expressed as natural logarithms of specific humidity), as well as surface air and skin temperature. The observation vector $y_o$ is composed of 22 satellite radiance channels expressed in terms of brightness temperatures. A fast radiative transfer model is then used to compute the corresponding model vector $y(x)$.

The background profile $x_b$ is obtained from a NWP model. The background covariance matrix $B$ is generated from statistics of radiosonde-forecast differences with inter-level correlations of error retained. Finally, the measurement-error covariance $O + F$ is a fixed diagonal matrix in brightness-temperature space corresponding to specified standard deviations.

The minimization method uses a Newtonian iteration scheme, the penalty function gradient being computed with the one-dimensional adjoint model. The iterative minimization proceeds for a maximum of five iterations and is considered to have converged when the increment is less than 0.4 the standard deviation of the background error at every level. Any sounding that has not converged by this point is rejected.
References


Beljaars, A., J.-F. Mahfouf, J. Teixeira, and P. Viterbo, 1996a: Improvements to the 2m temperature forecasts. ECMWF Newsletter Number 73.


Chen, T., and al., 1997: Cabauw experimental results from the project for intercomparison of land-surface parameterization schemes. J. Climate, 10, 1194-1215.


References


References


Richardson, L., 1922: *Weather prediction by numerical process*. Cambridge Univ. Press.


Rontu, L., 1995: One-dimensional study of a clear night temperature. HIRLAM Newsletter Number 22.


Saas, B., and A. McDonald, 1995: A possible contributor to the problem of the minimum 2m temperatures being too high on calm clear nights. HIRLAM Newsletter Number 22.


Schubiger, F., 1995: Verifikation mit Bodenbeobachtungen an der SMA. DWD-SMA Rundbrief Nr. 10, Zürich, Switzerland.


Acknowledgments

Je tiens à remercier Jean Quiby en premier lieu, pour m'avoir donné l'occasion de découvrir le monde passionnant de la prévision numérique durant cette thèse. I'm also thankful to Prof. Hew Davies for his guidance, his suggestions and patience which improved greatly the quality of this thesis. Ich möchte noch Prof. Christoph Schär für seine hilfreiche Bemerkungen danken.

Toute ma reconnaissance aux membres du processus "Modell", ou dit autrement, de l'ex-section NUM de l'institut suisse de météorologie à Zürich. En particulier, Marco Arpagaus, Jean-Marie Bettems, Peter Binder, Pirmin Kaufman, Guy de Mor-sier, Andrea Rossa, Francis Schubiger et Emanuele Zala. Un grand merci pour votre sympathie et votre disponibilité qui ont fait que l'ambiance au sein du groupe fût agréable et stimulante pendant les trois années passées en votre compagnie.

Tous mes remerciements à Laurent Bourqui, Anne Renaud, Laurent Gasser et ma famille, entre autres, pour leur amitié et leur soutien moral pendant les moments difficiles qui sont le lot de tout doctorant.
Seite Leer / Blank leaf
Curriculum Vitae

Jean-Marc Beroud  
born on January 24, 1971 in Uster (ZH), Switzerland  
citizen of Ecoteaux (VD)

Education

1977–1983  Ecole primaire, Chexbres (VD)  
1983–1986  Ecole secondaire, Vevey (VD)  
1986–1989  Gymnase, Burier (VD)  

Certificat de maturité — type C

1990–1995  Study of physics at the University of Lausanne. Specialization in the field of astronomy and astrophysics. Diploma thesis "Les étoiles à Baryum" (The Barium Stars) under the guidance of Prof. B. Hauck and P. North  

Dipl. Phys.

1995–1998  Ph.D. student of Prof. H.C. Davies at the Swiss Meteorological Institute  

Dr. sc. nat.

International Conferences and Seminars

1996  DWD-SMA Treffen, Bad Säckingen  
1997  EGS, Wien  
Seminar on Atmosphere-surface Interaction, Reading  
DWD-SMA Treffen, Bad Säckingen  
1998  EGS, Nice