UNIVERSITÉ DU QUÉBEC À MONTRÉAL

ÉTUDE DE LA LIGNE PLUIE-NEIGE DANS LA VALLÉE DE KANANASKIS, ALBERTA

MÉMOIRE

PRÉSENTÉ

COMME EXIGENCE PARTIELLE

DE LA MAÎTRISE EN SCIENCES DE L'ATMOSPHÈRE

PAR

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FÉVRIER 2017

UNIVERSITÉ DU QUÉBEC À MONTRÉAL
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REMERCIEMENTS

Je tiens à remercier le Conseil de recherches en sciences naturelles et en génie du Canada (CRSNG), les Fonds de recherche Nature et technologies du Québec (FRQNT), Environnement et Changement climatique Canada et le *Changing Cold Region Network* (CCRN) pour avoir contribué au financement de ce projet. Merci aussi à tous les participants de la campagne de terrain *Alberta Field Project.* Merci à Paul Vaquer pour m'avoir fourni des données de la campagne de terrain. Merci à mon mari Martin pour m'avoir supporté lors de mon retour aux études qui aura duré cinq ans. Merci au professeur René Laprise pour m'avoir permis de faire mes débuts en recherche lors de stages d'été ainsi qu'à Oumarou Nikiéma et Danahé Paquin-Ricard pour leur aide précieuse durant mes premiers stages. Merci tout particulièrement à ma directrice, la professeure Julie M. Thériault, pour m'avoir donné l'opportunité de participer à la campagne de terrain, pour sa confiance en moi et aussi pour sa grande disponibilité.

TABLE DES MATIÈRES

LISTE DES FIGURES

LISTE DES TABLEAUX

LISTE DES ABRÉVIATIONS, SIGLES ET ACRONYMES

ACR	Model run without heating due to accretion
ASL	Above Sea Level
BAR	Barrier Lake research station
CTL	Control run
FOR	Fortress Mountain
GPS	Global Positioning System
KES	Kananaskis Emergency Services
MAP	Mesoscale Alpine Program
MCICA	Monte Carlo Independent Column Approximation
MLT	Model run without cooling due to melting
MRR	Micro Rain Radar
NAK	Nakiska ski area
NARR	North American Regional Reanalysis
NCAR	National Center for Atmospheric Research
NCEP	National Centers for Environmental Prediction
NOGRP	Model run assuming no graupel formation
RRTMG	Rapid Radiative Transfer Model
SBL	Model run without cooling due to sublimation
UTC	Universal Time Coordinated
WRF	Weather Research and Forecasting

RÉSUMÉ

La phase de la précipitation affecte la ressource en eau et peut mener à d'importants désastres comme l'inondation de Calgary en 2013 où une ligne pluie-neige élevée en altitude a été l'un des principaux facteurs menant à cette inondation catastrophique. Les changements de phase des précipitations traversant la ligne pluie-neige peuvent avoir un impact significatif sur les conditions environnantes. Le refroidissement dû à la fonte de la neige engendre de l'air froid plus dense qui descend vers le bas, ce qui peut affecter la direction du vent. Il a été démontré que la ligne pluie-neige descend sur la pente face au vent d'une montagne durant une tempête. Une inversion de la direction du vent de montant à descendant a aussi été observée à cause du refroidissement dû à la fonte de la neige. La présente étude a focalisé sur l'importance relative de la sublimation et de la fonte en utilisant des simulations numériques basées sur un événement de précipitation particulier observé durant une campagne de mesures dans la vallée de Kananaskis en Alberta. Des études de sensibilité ont été effectuées pour examiner la relation entre l'effet de la variation de température due à la sublimation et l'écoulement dans la vallée, ainsi que l'importance de l'accrétion et l'impact de la formation de la neige roulée. Les résultats ont montré que des lignes pluie-neige peuvent être observées à des températures de l'air au-dessus de 0 °C dans la vallée de Kananaskis en Alberta à cause du climat sec. L'effet du refroidissement dû à la sublimation et à la fonte a affecté le taux de précipitation et a influencé le changement de direction de l'écoulement de montant à descendant sur la pente en amont de la montagne. Les simulations numériques ont aussi montré que la formation de la neige roulée durant un événement météorologique a influencé l'inversion de la direction du vent. Le réchauffement dû à l'accrétion en altitude durant la formation de la neige roulée a affecté le refroidissement dû à la sublimation à plus basse altitude et a aussi influencé le taux de précipitation à la surface. Dans l'ensemble, cette étude a démontré que le refroidissement dû à la sublimation et aussi la présence de neige roulée peuvent affecter l'intensité et aussi la durée de la précipitation, ainsi que la direction de l'écoulement dans la vallée.

Mots-clés : Ligne pluie-neige, Sublimation, Accrétion, Terrain montagneux, Écoulement dans une vallée, Précipitation, Microphysique

INTRODUCTION

La précipitation a un impact majeur sur la population du sud de l'Alberta, surtout au printemps quand la neige fond à haute altitude. La phase de la précipitation affecte la ressource en eau et peut mener à d'importants désastres comme l'inondation de Calgary en 2013. En juin 2013, de fortes précipitations et la fonte rapide de la neige à haute altitude ont causé d'importantes inondations dans la partie sud de l'Alberta. Lors de cet événement, les fortes pluies ont causées du ruissellement à basse et à moyenne altitude, mais aussi du ruissellement causé par de la pluie sur neige à haute altitude dû à une accumulation tardive de neige au sol (Pomeroy et al., 2016). Les conditions à l'échelle synoptique ont aussi joué un rôle important lors de l'inondation (Milrad et al., 2015; Liu et al., 2016). La circulation atmosphérique autour d'un système de basse pression à la surface a produit des conditions de soulèvement orographique au pied des montagnes du sud de l'Alberta durant plus de 36 heures. Cet air chaud et humide a mené le niveau de gel à une plus haute altitude dans l'atmosphère (Pomeroy et al., 2016). Les tempêtes de début d'été amènent habituellement de la neige à haute altitude, mais dans le cas présent, de la pluie est survenue. La pluie sur neige causée par une ligne pluie-neige plus haute que la normale a été un des nombreux facteurs ayant mené à cette inondation catastrophique. Cet événement démontre l'importance de bien comprendre les caractéristiques et le comportement de la ligne pluie-neige en terrain montagneux.

La ligne pluie-neige ou encore région de transition est la région des tempêtes où l'on retrouve un mélange de précipitation sous les formes solide et liquide. En région montagneuse, c'est la limite entre la pluie à basse altitude et la neige à haute altitude. La largeur de la zone de transition et le type de précipitation à l'intérieur de cette limite dépendent fortement des conditions météorologiques comme la température, le vent et l'humidité relative, car ces facteurs peuvent être modifiés par la chaleur latente associée aux changements de phase (Stewart, 1992). Les changements de phase des précipitations traversant la ligne pluie-neige peuvent avoir un impact significatif sur les conditions environnantes. Une meilleure compréhension des processus microphysiques et des interactions complexes ayant lieu dans la région de ligne pluie-neige est un élément clé pour une meilleure prévision des événements météorologiques extrêmes en région montagneuse.

Un des principaux processus physiques dans la région d'une ligne pluie-neige est la fonte de la neige. La fonte de la neige extrait de l'énergie de l'environnement, ce qui amène un refroidissement de l'atmosphère. Wexler et al. (1954) ont montré que le refroidissement dû à la fonte peut affecter l'emplacement de la ligne pluie-neige et une région qui recevait initialement de la pluie pourrait éventuellement recevoir de la neige. La précipitation associée à une ligne pluie-neige est organisée en bandes alignées le long de la région de transition et les précipitations qui tombent à l'intérieur de cette région sont sujettes à une plus forte agrégation (Stewart, 1992). Ces particules sont associées à une région où la réflectivité radar est plus forte et est connue sous le nom de bande brillante.

Les premières études sur la ligne pluie-neige en région montagneuse ont été présentées par Marwitz (1983, 1987) en utilisant des observations provenant de la Sierra Nevada en Californie. Des données de radar Doppler ont démontré que la bande brillante est descendue de 400 à 600 m à l'approche de la pente de la montagne (Marwitz, 1983). Des données d'observations à bord d'avion ont aussi montré que l'isotherme 0 °C descendait de plusieurs centaines de mètres près de la pente de la montagne (Marwitz, 1987). Ceci était causé par le processus diabatique de la fonte et par la dynamique associée à l'écoulement orographique. Une autre étude a montré que la bande brillante est descendue à une plus basse altitude sur la pente face au vent dans les Alpes et dans la Chaîne des Cascades en Oregon durant trois tempêtes différentes (Medina et al., 2005). Plus récemment, des simulations numériques ont aussi été utilisées pour étudier l'abaissement de la ligne pluie-neige sur la pente face au vent d'une montagne (Minder et al., 2011). Le refroidissement de l'air dû à la fonte des hydrométéores gelés, le refroidissement adiabatique dû au soulèvement de l'air et la distance de fonte des hydrométéores sont trois mécanismes physiques qui ont été identifiés dans cette dernière étude pour leur influence sur la position de la ligne pluie-neige sur la pente d'une montagne.

En plus des effets thermodynamiques de la fonte, des effets dynamiques ont aussi été observés. Steiner et al. (2003) ont démontré avec des mesures de radar Doppler que le mouvement de l'air dans la vallée de la rivière Toce dans les Alpes italiennes s'est inversé de montant à descendant après quelques heures de pluie durant le *Mesoscale Alpine Program* (MAP). Cette étude a suggéré que le refroidissement dû à la fonte a joué un rôle important dans la production de l'écoulement vers le bas de la vallée et que l'évaporation de la précipitation était de plus faible importance. Medina et al. (2005) ont aussi observé cet effet dynamique avec des données radar aériennes montrant un faible écoulement inversé descendant la pente dans les vallées profondes des Alpes durant le programme MAP. Asencio et Stein (2006) ont effectué des études à l'aide d'un modèle qui étaient en accord avec l'analyse de Steiner et al. (2003), montrant que le refroidissement diabatique associé à la fonte des hydrométéores était fondamental dans la production de l'écoulement descendant durant le programme MAP. Cependant, Zängl (2007) a suggéré avec des simulations numériques que le refroidissement par la fonte était de plus faible importance dans la création de l'écoulement vers le bas de la vallée pour le même événement. Plus récemment, Thériault et al. (2015) ont effectué des simulations numériques et observé un changement dans la direction de l'écoulement près d'une montagne de la région de Whistler, CB, quand l'effet diabatique de la fonte de la neige était considéré. Le temps avant l'inversion de l'écoulement était beaucoup plus long pour des plus petits taux de précipitation. La vitesse de descente de la ligne pluie-neige vers le bas de la montagne augmentait avec un plus fort taux de précipitation. Cette étude a démontré que la ligne pluie-neige atteignait généralement la base de la montagne avant que l'inversion de l'écoulement ne remplisse complètement la profondeur initiale de la couche de fonte.

Dans un environnement saturé, la fonte de la neige commence quand l'environnement entourant les flocons de neige atteint la température de $0\,^{\circ}\mathrm{C}$. Dans des conditions sous-saturées, la sublimation de la neige prend plus d'importance et ralentit le processus de fonte. Dans cet environnement sous-saturé, la fonte de la neige ne débute pas quand les flocons de neige atteignent l'isotherme $0\,^{\circ}\mathrm{C}$, mais quand $T_w = 0\,^{\circ}\mathrm{C}$. Ceci est la température du thermomètre mouillé et tant qu'elle est égale à $0\,^{\circ}\mathrm{C}$ ou au-dessous, de la neige va tomber. Dans les régions où l'humidité relative est basse, de la neige peut être observée même à de hautes températures de l'air à la surface de 4 à 6 °C (Matsuo et Sasyo, 1981 ; Harder et Pomeroy, 2013).

Contrairement au refroidissement dû à la fonte, il y a eu très peu d'études regardant les effets du refroidissement dû à la sublimation dans les tempêtes hivernales, plus particulièrement en région montagneuse. Clough et Franks (1991) ont examiné les processus d'évaporation dans des régions de précipitation frontale et stratiforme. Ils ont démontré que la sublimation des particules de glace était un processus thermodynamique efficace et que même dans des conditions de faible sous-saturation de 5 à 10 %, une sublimation appréciable avait tout de même lieu. Parker et Thorpe (1995) ont étudié le rôle de la sublimation de la neige sur la frontogenèse et ont démontré que l'écoulement transversal frontal dans la région de sublimation était fortement modifié. Un courant descendant à méso-échelle était produit sous la surface synoptique frontale. La sublimation de la neige et de la neige roulée a joué un rôle important dans l'évolution de la bande de pluie associée à un front froid qui a été modélisée dans l'étude de Barth et Parsons (1996). L'effet de la sublimation a augmenté l'intensité et la profondeur de la masse d'air froid. Ils ont aussi trouvé que le processus de sublimation contribuait plus au refroidissement que le processus de fonte.

Le type de précipitation qui tombe a un impact sur la façon dont les particules de glace subliment. Quand une gouttelette d'eau liquide surfondue entre en contact avec un hydrométéore gelé, la gouttelette gèle à l'impact et produit un réchauffement de l'environnement. La particule solide changera ensuite de forme à cause du givrage. Ces particules de neige roulée ont une plus grande densité que la neige et donc subliment plus lentement (Clough et Franks, 1991 ; Burford et Stewart, 1998). Aussi, une plus grande vitesse terminale donne aux particules de neige roulée moins de temps pour sublimer durant la descente par rapport à la neige. La neige roulée a par conséquence plus de chance de survie en traversant la ligne pluie-neige que les dendrites par exemple.

La région d'intérêt de la présente étude est la vallée de Kananaskis qui est située sur la pente est des montagnes Rocheuses canadiennes, 60 km à l'ouest de Calgary en Alberta (Fig. 0.1). Comme la plupart des régions montagneuses continentales, le climat dans la vallée de Kananaskis varie beaucoup. Durant l'hiver, le climat alterne entre des périodes froides et sèches, et des périodes plus chaudes et sèches où le Chinook est présent, ce qui amène une large plage de températures pour cette région du sud de l'Alberta (Whitfield, 2014). Le Chinook est caractérisé par un fort écoulement provenant de l'ouest qui passe par-dessus les montagnes et par de l'air sec qui descend la pente est en amenant de hautes températures et une faible humidité relative dans cette région (Whitfield, 2014). L'hiver 2015 a reçu moins de précipitation que la normale, ce qui a mené à de plus faibles accumulations

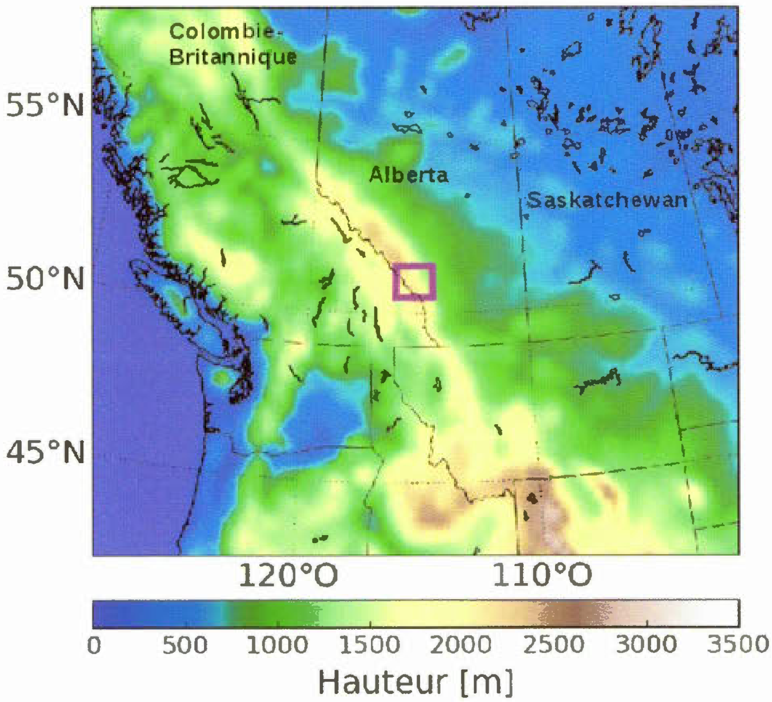

Figure 0.1 Emplacement de la vallée de Kananaskis en Alberta.

de neige dans les Rocheuses canadiennes (Derworiz, 2015). À cause du climat sec présent dans la vallée de Kananaskis, la sublimation des particules de glace est sûrement un facteur dominant dans cette région.

À cause de l'importance d'améliorer notre compréhension de la ligne pluie-neige dans les régions sèches, cette étude a focalisé sur l'importance relative de la sublimation et de la fonte durant une étude de cas. Plus particulièrement, cette étude a utilisé des simulations numériques basées sur un événement de précipitation particulier observé durant une campagne de mesure en Alberta (*Alberta Field Project*) pour tenter de répondre à quelques-unes des questions suivantes :

Est-ce que la ligne pluie-neige se retrouve à une température de 0 °C sur la pente est des Rocheuses canadiennes ?
Quelle est l'importance relative de la sublimation et de la fonte ?
Quel est l'impact de la sublimation de la neige sur l'écoulement dans la vallée ?
Quelle est l'importance de l'accrétion en altitude ?

Des études de sensibilité ont été effectuées pour examiner la relation entre le refroidissement diabatique dû à la sublimation et l'écoulement dans la vallée. Aussi, l'importance de l'accrétion au-dessus de la ligne pluie-neige sur le processus de sublimation et sur la direction du vent est discutée.

Le présent mémoire est organisé comme suit. Le chapitre 1 contient un article scientifique rédigé en anglais. La section 1.1 présente une introduction au contexte scientifique. La section 1.2 présente une vue d'ensemble de la campagne de terrain effectuée au printemps 2015 et décrit l'étude de cas utilisée dans cette étude. La méthodologie utilisée dans la configuration du modèle numérique et les études de sensibilité sont expliquées dans la section 1.3. Les résultats de la simulation contrôle, l'importance relative de la fonte et de la sublimation ainsi que les effets de la suppression du refroidissement dû à la fonte et à la sublimation de la neige et de la neige roulée sont résumés dans la section 1.4. Les effets de l'accrétion et de la présence de neige roulée sur le refroidissement dû à la sublimation et à la fonte ainsi que sur la direction du vent sont examinés dans la section 1.5. Une discussion est fournie à la section 1.6 et les conclusions sont résumées à la section 1.7.

CHAPITRE I

ARTICLE

A study of rain-snow transitions in the Kananaskis valley, Alberta

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À soumettre à la revue *Atmospheric Research*

ABSTRACT

The phase of precipitation affects water resources and this can lead to major disasters such as the Calgary 2013 flooding event where an elevated rain-snow boundary was one of the main factors that led to the catastrophic flooding. The phase changes of precipitation falling through a rain-snow boundary can have a significant impact on the environmental conditions. Cooling by melting of snow generates cold dense air that moves down, which can affect the wind direction. It has been shown that the rain-snow transition is lowering on the windward slope of a mountain during a storm. A reversal in the wind direction from up-valley to down-valley was also observed as a result of cooling by melting of snow. The present study focused on the relative importance of sublimation and melting using numerical simulations based on a particular precipitation event observed during the Alberta Field Project in the Kananaskis valley. Sensitivity experiments were used to examine the relationship between the temperature feedbacks from sublimation and the valley flow field along with the importance of accretion and the impacts of graupel formation. The results showed that rain-snow boundaries can be observed at surface air temperatures above 0°C in the Kananaskis valley, Alberta due to the dry climate. The cooling effect of sublimation and melting affected the precipitation rate and influenced the change in flow direction from upslope to downslope on the west facing slope of the mountain. The numerical simulations also showed that the formation of graupel during the weather event influenced the wind direction reversal. The heating due to accretion aloft during the formation of graupel affected the cooling due to sublimation at lower altitudes and also influenced the precipitation rate at the surface. Overall, this study showed that cooling due to sublimation and also the presence of graupel can affect the intensity and also the duration of precipitation, along with changing the flow direction in a valley.

Keywords: Rain-snow boundary, Sublimation, Accretion, Complex terrain, Valley flow, Precipitation, Microphysics

1.1 Introduction

Precipitation has a major impact on the population of southern Alberta, especially during the spring season when snow melts at higher elevation, causing runoff into rivers. The phase of precipitation affects water resources and this can lead to major disasters such as the Calgary 2013 flooding event. In late June 2013, heavy rainfall and rapidly melting alpine snow caused flooding to the southern part of Alberta. In this particular event, the heavy rain generated rainfall runoff at low and middle elevations, but it was supplemented by rain-on-snow runoff at high elevations due to a late lying snowpack (Pomeroy et al., 2016). The synoptic scale atmospheric conditions also played an important role in the flooding event (Milrad et al., 2015; Liu et al., 2016). The circulation around the surface low-pressure system produced upslope conditions over the southern Alberta foothills that persisted for more than 36 hours. This air was moist and warm which led to a high freezing level in the atmosphere (Pomeroy et al., 2016). Early summer storms usually bring alpine snowfall at high elevations, but in this case rainfall occurred. The rain-on-snow caused by a higher than usual rain-snow boundary was one of the many factors that lead to this catastrophic flooding. This event illustrates the importance of understanding the characteristics and behavior of the rain-snow boundary in mountainous regions.

The rain-snow boundary or transition region is the region of a storm characterized by mixed precipitation. In mountainous regions, it is the boundary between rainfall at low elevations and snowfall at higher elevations. The width of the transition region and the precipitation type distribution within that boundary depend strongly on the weather conditions such as temperature, wind field and relative humidity because these could be altered by the latent heat associated with phase changes (Stewart, 1992). The phase changes of particles falling through the rain-snow boundary can have a significant impact on the environmental conditions. Understanding the effects of the microphysical processes taking place in the vicinity of a rain-snow boundary is a key element to a better forecasting of extreme meteorological events in mountainous regions.

One of the key physical processes within the rain-snow boundary is the melting of falling snow. Melting of snow extracts energy from the environment, which leads to atmospheric cooling. Wexler et al. (1954) first noted that cooling by melting can affect the location of the rain-snow boundary and an area initially receiving rain will eventually receive snow. The precipitation associated with a rain-snow boundary is typically organized into banded features aligned along the transition region and particles falling through this region would be subjected to enhanced aggregation (Stewart, 1992). These particles are associated with a region of enhanced radar reflectivity known as the radar bright band.

Some of the earliest studies on rain-snow transitions in mountainous areas were presented by Marwitz (1983, 1987) using observations over the Sierra Nevada. Doppler radar data showed that the bright band began to increase in depth to 400-600 m while approaching the mountain barrier (Marwitz, 1983). Also, in situ aircraft data showed that the 0°C isotherm descended several hundred meters near the barrier as a result of the diabatic process of melting and the dynamics of orographic airflow (Marwitz, 1987). Another study noted that the bright band fell to a lower height over the windward slopes in the Alps and the Oregon Cascade Mountains during three different alpine storms (Medina et al., 2005). More recently, numerical simulations were also used to study the lowering of the rain-snow boundary on a mountain windward slope (Minder et al., 2011). The cooling of the air by melting of frozen hydrometeors, the adiabatic cooling of rising air and the melting distance of hydrometeors were identified as three physical mechanisms influencing the location of the rain-snow boundary along the mountainside.

Beside the thermodynamic effects of melting, dynamical feedbacks have also been observed. Steiner et al. (2003) showed with Doppler radar measurements that the air motion within the Toce river valley in the Italian Alps during the Mesoscale Alpine Program (MAP) reversed from up-valley to down-valley flow after a few hours of rain. This study suggested that cooling by melting of snow played an important role in generating the down-valley flow and that evaporation of precipitation was of less importance. Medina et al. (2005) also observed this dynamic effect with airborne radar data showing a weak and reversed downslope flow in the deep valleys of the Alps during the MAP program. The results of the model experiments of Asencio and Stein (2006) were in agreement with the analysis of Steiner et al. (2003) that the diabatic cooling associated with the melting of falling frozen hydrometeors was fundamental in generating the down-valley flow during the MAP program. However, Zängl (2007) suggested with numerical simulations that the cooling by melting of snow was of less importance in creating the down-valley flow for the same event. More recently, Thériault et al. (2015) performed numerical simulations and observed a change in the valley flow field direction near a mountain in the Whistler, BC area when the diabatic effect of melting snow was considered. The time to produce the flow reversal was much longer for lower precipitation rates. The speed of the rain-snow boundary traveling down the mountain also increased with higher precipitation rates. This study showed that the rain-snow boundary generally reaches the base of the mountain before the flow reversal has completely filled up the initial depth of the melting layer.

In a saturated environment, the melting of snow starts when the environment surrounding the snowflakes reach a temperature of 0°C. Within subsaturated conditions, the sublimation of snow becomes of greater importance and thus slows down the melting process. In a subsaturated environment, the melting of snow does not start when the falling snowflakes reach the 0°C isotherm, but rather when

$T_w = 0°C$. This is known as the wet-bulb temperature and as long as it is equal to 0°C or below, snow will fall. In regions of low relative humidity, snowfalls can be seen even at high surface air temperature of 4-6°C (Matsuo and Sasyo, 1981; Harder and Pomeroy, 2013).

Unlike cooling by melting, there are few studies addressing the effects of cooling by sublimation in winter storms, especially in mountainous regions. Clough and Franks (1991) examined the evaporative processes in frontal and stratiform precipitation. They showed that sublimation of ice particles was an efficient thermodynamic process and in low subsaturation of 5-10%, appreciable sublimation can still take place. Parker and Thorpe (1995) studied the role of snow sublimation on frontogenesis and showed that the cross-frontal flows in the vicinity of the sublimation were strongly modified. A mesoscale downdraft was produced below the synoptic frontal surface. Sublimation of snow and graupel played an important role in the evolution of the narrow cold-frontal rainband that was modeled in the study of Barth and Parsons (1996), as it increased the intensity and depth of the cold air mass. They also found that the sublimation process contributed more to the latent cooling than the melting process.

The type of precipitation falling has an impact on the sublimation process. When a supercooled liquid droplet hits a frozen hydrometeor, it freezes upon impact and thus produces heating of the environment. The solid particle will then change shape due to riming. These graupel particles have a higher density than snow and thus they sublimate slower (Clough and Franks, 1991; Burford and Stewart, 1998). Also, a higher terminal velocity will give the graupel particles less time to sublimate during descent, as in the case of snow. Graupel consequently has a greater chance to reach the surface while traveling through the rain-snow boundary than dendrites, which fall slower, for example.

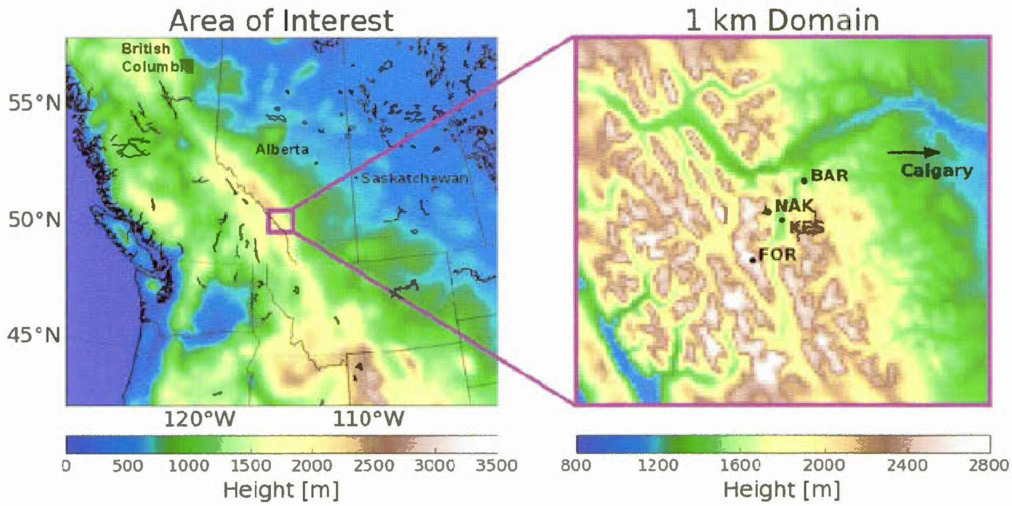

Figure 1.1 Area of interest and 1 km mesh domain in WRF model.

The area of interest for this study is the Kananaskis valley, which is located on the eastern side of the Canadian Rocky Mountains, 60 km west of Calgary, Alberta (Fig. 1.1). Like most mountain regions in continental areas, the Kananaskis valley climate is highly variable. In the winter, the climate alternates between cold, dry periods and periods of comparatively warm, dry, windy Chinook air, which gives to the general area of southwestern Alberta a large winter temperature range (Whitfield, 2014). The Chinook is characterized by a strong westerly flow over the mountains and dry air descending the leeward side which brings both high temperatures and low humidity to the area (Whitfield, 2014). The winter 2015 received lower than average precipitation in this area, which led to a shallower than usual snowpack in the Canadian Rockies (Derworiz, 2015). Due to the dry climate in the Kananaskis valley, the sublimation of ice particles is likely to be a dominant factor in this area.

Given the importance to improve our understanding of the rain-snow transition within relatively dry areas, this study focused on the relative importance of sublimation and melting during a case study. In particular, this study used numerical simulations based on a particular precipitation event observed during the Alberta Field Project, to help answer some of the following scientific questions:

Do rain-snow transitions occur at 0°C on the lee side of the Canadian Rockies?
What is the relative importance of sublimation and melting of snow and graupel?
What is the impact of snow sublimation on the valley flow field?
What is the importance of accretion aloft?

Sensitivity experiments were performed to examine the relation between the temperature feedbacks from sublimation and the valley flow field. Also, the importance of accretion above the rain-snow boundary on the sublimation process and on the wind direction is discussed.

This paper is structured as follows. Section 1.2 provides an overview of the field campaign performed in spring 2015 and describes the case study used in this paper. The methodology used in the model configuration and the sensitivity experiments is explained in Section 1.3. Results from the control simulation, the relative importance between melting and sublimation along with the effects of suppressing diabatic cooling due to melting and sublimation of snow and graupel are summarized in Section 1.4. The effects of accretion and graupel on the cooling associated with sublimation and melting along with the effects on wind direction is examined in Section 1.5. A discussion is provided in Section 1.6 and finally some concluding remarks are given in Section 1.7.

1.2 Case overview

1.2.1 Alberta Field Project

The field campaign Alberta Field Project was a collaboration of researchers interested in the study of spring precipitation on the eastern side of the Canadian Rockies. The campaign took place during March-April 2015 in the Kananaskis valley, Alberta. The goals of the field campaign were to study the rain-snow boundary, the snow crystal types, along with the characteristics of precipitation and the associated weather conditions in the valley.

Most of the observations were collected at the Kananaskis Emergency Services (KES) site just a few kilometers south-east of the Nakiska ski area (NAK) and about 15 km south of the Barrier Lake research station (BAR) (Fig. 1.1). Detailed meteorological and photographic measurements were collected at the main site. Instruments present included a Geonor precipitation gauge, a sounding system, an OTT Parsivel optical disdrometer and a Micro Rain Radar (MRR). Basic meteorological measurements were also available (pressure, winds, temperature, dew point temperature). Visual observations and high-resolution snowflakes macro-photography were taken as well to characterize the type of snow as in Gibson and Stewart (2007). Vertical profiles of basic meteorological features were also obtained using a Kestrel attached to a ski pole and a GPS at two other sites in the presence of rain-snow transitions. These vertical profiles were performed at NAK and also at Fortress Mountain (FOR). The method followed the one used in Thériault et al. (2014).

The forecasting of the different storms during the field campaign was especially challenging because most of the precipitation falling was isolated events that were not necessarily produced by typical low-pressure system located on the lee-side of the Rockies. Precipitation falling at the surface was very irregular through the valley. Sometimes, an appreciable accumulation was observed during a storm near Barrier Lake, while only a trace was reported at the KES observation site. It depended highly on the direction of the mesoscale flow leading to upslope flow (Vaquer, 2017). Rain-snow boundaries were also often observed along the mountainside.

1.2.2 Case Study: 31 March 2015

On the last day of March, a weather event associated with a rain-snow boundary along the mountainside occurred in the Kananaskis valley. A low pressure system moving to the east was located north-east of the Kananaskis valley at 0000 UTC 1 April 2015 (Fig. 1.2). The data logger and the Alberta Environment weather station located 100 m from each other at KES indicated a decrease in surface temperature starting at 2100 UTC 31 March 2015 from 12°C to 3°C (Fig. 1.3 a). An increase in the dew point temperature was also seen starting at 2000 UTC. The wind speed was between 2-3 m/s from 1900 to 2100 UTC and started to decrease until 0000 UTC (Fig. 1.3 b). The Alberta Environment weather station indicated an increase in the wind speed at 2200 UTC to 4.5 m/s while the data logger showed lower wind speed. The Alberta Environment weather station provided hourly data while the data logger provided 1-min data, which could explain the difference at this time. Manual observations at the KES site showed light rain starting at 2030 UTC 31 March 2015, changing to a mix of rain, snow and graupel, then to a brief period of snow only (Fig. 1.3 c). A vertical sounding launched at 2100 UTC showed clearly subsaturated condition near the surface at the KES site (Fig. 1.4).

During this mixed precipitation event, the rain-snow transition was located at the base of the mountain. A car sounding performed at FOR indicated that the top of the rain-snow boundary was at an altitude of 1830 m above sea level (ASL) at

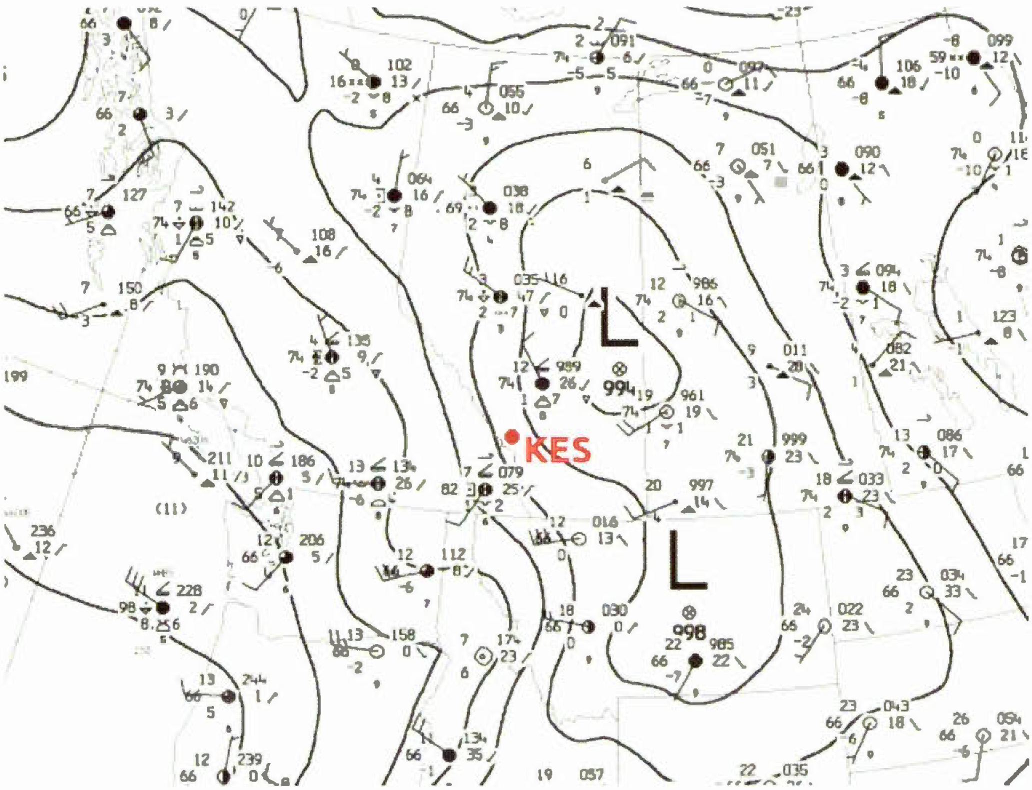

Figure 1.2 Surface analysis at 0000 UTC 1 April 2015. source: Environment and Climate Change Canada.

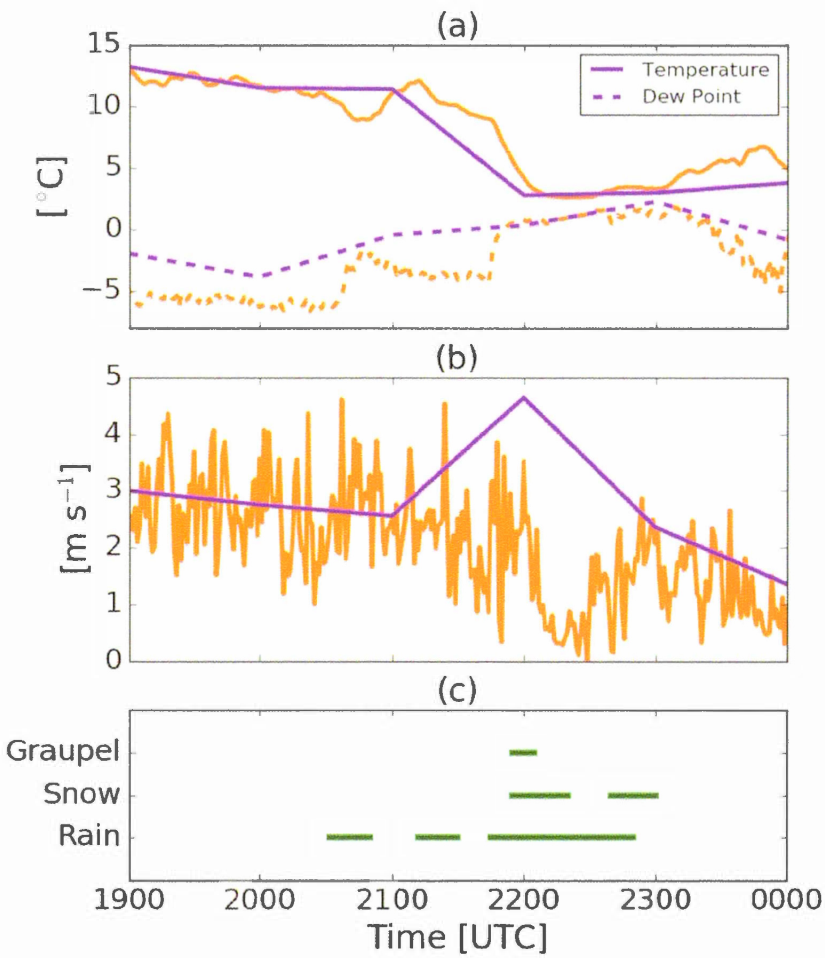

Figure 1.3 (a) Surface temperature (solid line) and dew point temperature (dashed line), (b) wind speed and (c) precipitation types observed at KES. Data are from the data logger (orange), the Alberta Environment weather station (purple) and visual observations (green).

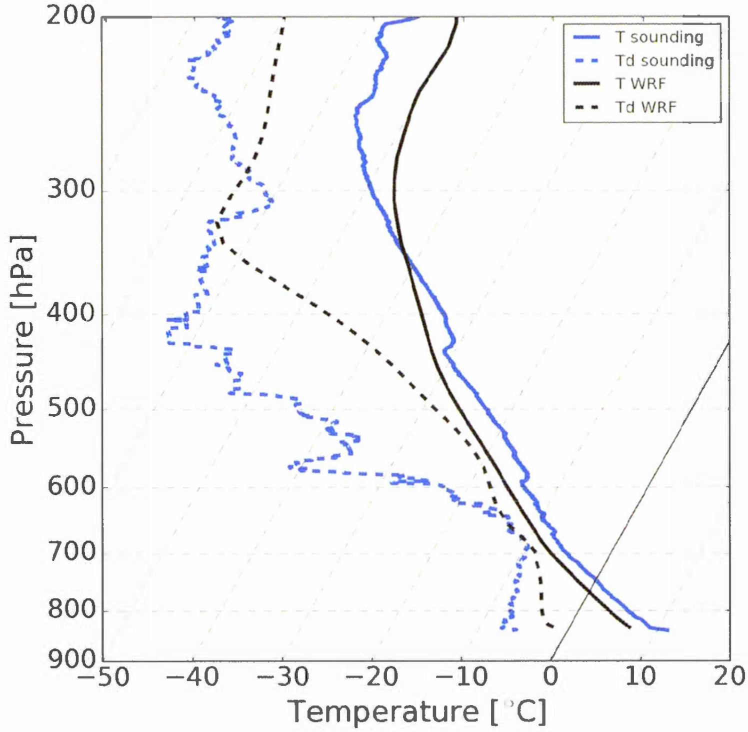

Figure 1.4 Temperature (solid line) and dew point temperature (dashed line) vertical profiles at 2100 UTC 31 March 2015 above the KES site. Both the sounding launched (blue) and the control run (black) described in section 1.3.1 are shown.

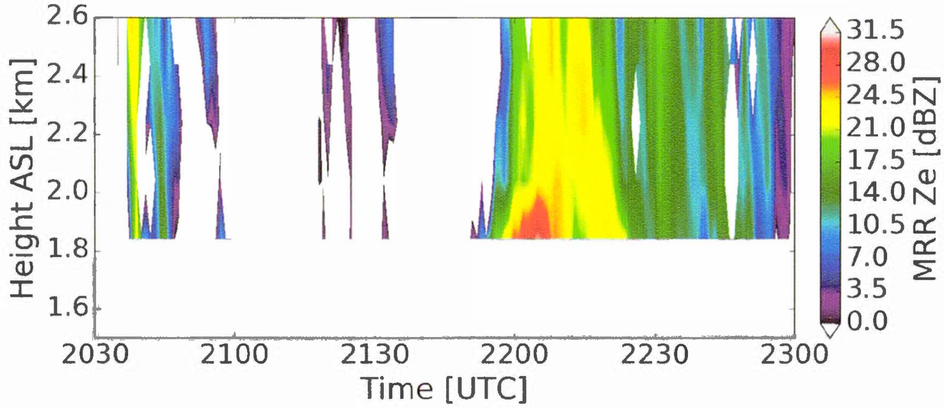

Figure 1.5 Reflectivity measured by the Micro Rain Radar at the KES site on 31 March 2015.

2115 UTC and lowered to 1750 m ASL at 2200 UTC. The KES site being located at an altitude of 1445 m ASL, the thickness of the rain-snow boundary varied between 300-400 m according to manual observations. The reflectivity from the MRR located at KES showed the presence of a bright band (maximum reflectivity) at around 2200 UTC (Fig. 1.5). The top of the bright band was at a height of 2 km ASL, which is, however, at a slightly higher altitude than in the manual observations and simulations.

In the following sections, the results of the numerical simulations for this study will mainly be presented for the KES site at the bottom of the Kananaskis valley.

1.3 Methodology

1.3.1 Model configuration

The simulations in this study were performed using the Weather Research and Forecasting (WRF) model, version 3.7.1 (Skamarock et al., 2008). Three-dimensional (3D) simulations were used with initial and boundary conditions provided by the North American Regional Reanalysis (NARR) data from the National Centers for Environmental Prediction (NCEP) (Mesinger et al., 2006). Two-way nesting with four nested grids (27 km, 9 km, 3 km and 1 km) was used to perform high-resolution simulations over the Kananaskis valley. The high-resolution domain is shown on Figure 1.1. The control run and the sensitivity tests were done using the two-moment version of the Milbrandt and Yau bulk microphysics scheme (Milbrandt and Yau, 2005). This microphysics scheme predicts the mass mixing ratio and total number concentration of six hydrometeor categories: cloud droplets, rain, ice crystals, snow, graupel and hail. Some modifications were made to the Milbrandt and Yau microphysics scheme to get snow sublimation at temperatures above and below 0°C. The differences are described in Appendix A.

Other microphysical parameterizations used in the simulations included the Rapid Radiative Transfer Model (RRTMG) with the Monte Carlo Independent Column Approximation (MCICA) method of random cloud overlap scheme (Iacono et al., 2008) for longwave and shortwave radiation. Also, the Noah Land Surface Model (Tewari et al., 2004) with soil temperature and moisture in four layers, fractional snow cover and frozen soil physics was used. The planetary boundary layer was parameterized in the simulations with the Yonsei University scheme which uses the non-local K approach with an explicit entrainment layer and a parabolic K profile in the unstable mixed layer, where K is the vertical diffusion coefficient (Hong et al., 2006). Cumulus parameterization was used on the coarser grid only

(27 km) with the Kain-Fritsch scheme (Kain, 2004).

To obtain a maximum number of vertical levels within the melting layer, 56 vertical levels have been used where the grid spacing varied from 50 to 350 m in the first 2 km and was about 350 m at higher levels. The simulation on the coarser grids (27, 9 and 3 km) started at 1500 UTC 31 March 2015, 3 hours prior to the higher resolution grid (1 km), which started at 1800 UTC 31 March 2015. The simulations were integrated for a total of, respectively, 12 h and 9 h. The time step used was 90 s on the coarser grid (27 km) decreasing with a ratio of 3 between each nested grid to 3.33 s on the higher resolution grid (1 km).

1.3.2 Sensitivity experiments

To estimate the temperature feedbacks associated with melting and sublimation, the control simulation (CTL) was first run while neglecting the latent heat due to the melting of snow and graupel (MLT). Then, it was neglected for the sublimation of snow and graupel (SBL). The temperature tendency equation from Milbrandt and Yau (2005) is:

$$\frac{dT}{dt} = \frac{1}{\Delta t}\left\{\begin{array}{l} \dfrac{L_f}{c_{pd}}\left(\begin{array}{c} \Delta QCLcs + \Delta QCLcg + \Delta QCLch + \Delta QCLri + \Delta QCLrs \\ + \Delta QCLrg + \Delta QCLrh + \Delta QFZci + \Delta QFZrh \\ - \Delta QMLir - \Delta QMLsr - \Delta QMLgr - \Delta QMLhr \end{array}\right) \\ + \dfrac{L_s}{c_{pd}}\left(\begin{array}{c} \Delta QNUvi + \Delta QVDvi + \Delta QVDvs + \Delta QVDvg \\ + \Delta QVDvh \end{array}\right) \end{array}\right. \tag{1.1}$$

where L_f is the latent heat of fusion, L_s is the latent heat of sublimation, c_{pd} is the specific heat of dry air and Q is for mixing ratio. The types of mixing ratios are noted by CL for collection, FZ for freezing, ML for melting, NU for nucleation, VD for diffusional growth (positive) or sublimation (negative) and the subscripts (c, r, i, s, g, h, v) represent cloud droplets, rain, ice, snow, graupel, hail and water vapor.

In the MLT experiment, the sink terms $QMLsr$ and $QMLgr$ were neglected in Eq. 1.1. Accordingly, the negative portion of the terms $QVDvs$ and $QVDvg$ were also neglected for the SBL experiment. These experiments were used to show the impact of the temperature feedbacks from melting or sublimation of snow and graupel since these two microphysical processes can induce a cooling of the environment.

To assess the impact of accretion aloft and also the impact of the presence of riming (graupel) on the sublimation occurring at lower altitudes, the CTL simulation was also run with no graupel (NOGRP). In the last experiment, a simulation was done to determine the temperature feedbacks associated with the heating due to accretion above the rain-snow boundary (ACR). In this last simulation, the latent heat due to the following processes was neglected in Eq. 1.1: collection of cloud droplets with snow or graupel particles ($QCLcs$, $QCLcg$) and collection of raindrops with ice, snow and graupel particles ($QCLri$, $QCLrs$, $QCLrg$). These different processes are used in the scheme to produce graupel. In the case of accretion, a heating of the environment is produced at high altitude and this experiment will assess the impact of the temperature feedbacks. The sink terms neglected from the temperature tendency equation in the Milbrandt and Yau scheme for each experiment are listed in Table 1.1.

Table 1.1 List of the sensitivity experiments

Experiments	**Sink Terms**	**Description**
CTL	-	control simulation
SBL	$QVDvs(-)$, $QVDvg(-)$	no cooling effect from sublimation
MLT	$QMLsr$, $QMLgr$	no cooling effect from melting
ACR	$QCLcs$, $QCLcg$, $QCLri$, $QCLrs$, $QCLrg$	no heating effect from accretion
NOGRP	-	no graupel

1.4 Impacts of melting and sublimation

1.4.1 CTL simulation

The weather conditions in the CTL simulation were first studied and compared to some observations at the KES site to ensure that atmospheric conditions were well reproduced by the numerical simulations. Figure 1.6 shows precipitation accumulation from 1800 UTC 31 March 2015 to 0000 UTC 1 April 2015 in the Kananaskis area. The CTL run showed accumulation of rain in the valley and solid precipitation (snow and graupel) at higher elevations. The accumulated precipitation simulated at the KES site was slightly higher than the amount recorded in the observations. The rain gauge located at KES site indicated precipitation accumulations of 1.2 mm while the model showed accumulation between 1.5 and 2 mm. The Kananaskis weather station from Environment and Climate Change Canada located near Barrier Lake recorded, however, rain accumulation of 2.4 mm for the day of 31 March 2015 while the model showed no accumulation in this area. Both the vertical profiles from the CTL run and the sounding launched at KES at 2100 UTC 31 March 2015 indicated that the air was subsaturated near the surface (Fig. 1.4).

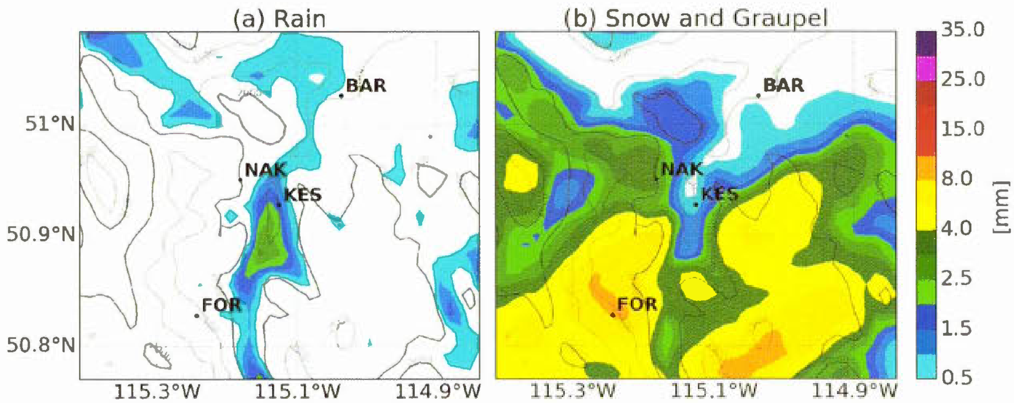

Figure 1.6 Precipitation accumulation from 1800 UTC 31 March 2015 to 0000 UTC 1 April 2015 for (a) rain and (b) snow and graupel.

While the precipitation amount accumulated at KES was relatively low during this event, a study of the different precipitation types for the CTL run showed a high amount of snow at high altitudes. Figure 1.7 illustrates the time evolution of rain, snow and graupel mixing ratio for the CTL run at KES site. Results showed that snow and graupel melted into rain starting at 2100 until 2230 UTC. The presence of a rain-snow transition between 2100 and 2230 UTC in the CTL run is of particular interest since this transition was well-observed in the field. A significant drop in the height of the 0°C isotherm at around 2115 UTC can be seen in the CTL run at the same time the rain-snow boundary was present at the surface. Figure 1.7 also reveals that the rain-snow transition was located more than 200 m below the 0°C isotherm, which confirms the presence of a non-melting layer just above the height where $T_w = 0$°C, as discussed by Matsuo and Sasyo (1981).

Both the manual observations at KES site and the CTL run confirmed that snow was present at a temperature of 3°C at around 2200 UTC. The fact that snow was

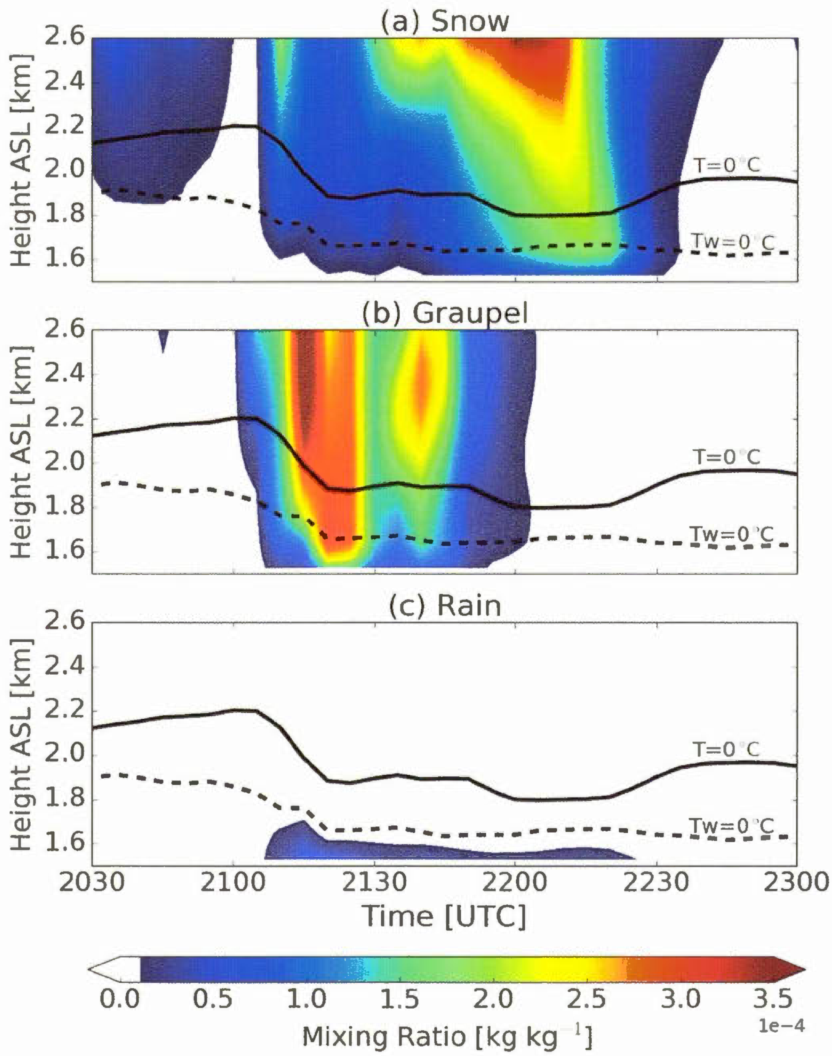

Figure 1.7 Time evolution of (a) snow, (b) graupel and (c) rain mixing ratio at the KES site on 31 March 2015 for the control run. The solid line indicates the height of the 0°C isotherm and the dashed line indicates the height where the wet-bulb temperature is 0°C.

present at the surface at temperatures above 0°C supports the theory of Matsuo and Sasyo (1981) that the precipitation type at the surface depends strongly on relative humidity and air temperature.

The top of the rain-snow boundary simulated by the model was around 1.8 km ASL at 2115 UTC and lowered to an altitude of about 1.7 km ASL at 2200 UTC. The car sounding performed at Fortress Mountain during this period indicated a similar height of the top of the boundary. The measured width of the melting layer dropped from 250 to 150 m from 2115 to 2200 UTC.

The results of the CTL run confirmed that snow can be seen at temperatures above 0°C in the Kananaskis area and that the rain-snow boundary occurred at a warmer temperature than when the environment is saturated. Comparisons between observations and the CTL run indicated that the weather conditions in the Kananaskis Valley on 31 March 2015 were well reproduced by the model, since a rain-snow boundary was present between 2100 and 2230 UTC 31 March 2015, as observed. Now that the weather conditions at the KES site are relatively well represented for the case study, the remaining analysis will focus on the microphysical processes in the vicinity of the transition region described above. Since snow was present at temperatures above 0°C and that the air was subsaturated, sublimation may be important during this time as well as melting, which was limited to a thin layer near the surface.

1.4.2 Relative importance of melting and sublimation

To assess the relative importance of melting and sublimation on the environmental temperature, the impact of the sink terms in Eq. 1.1 were evaluated separately ($QVDvs$ and $QVDvg$ for sublimation of snow and graupel; $QMLsr$ and $QMLgr$ for melting of snow and graupel). The evolution in time of dT/dt for sublimation and melting of snow and graupel is represented in Figure 1.8. The cooling effect

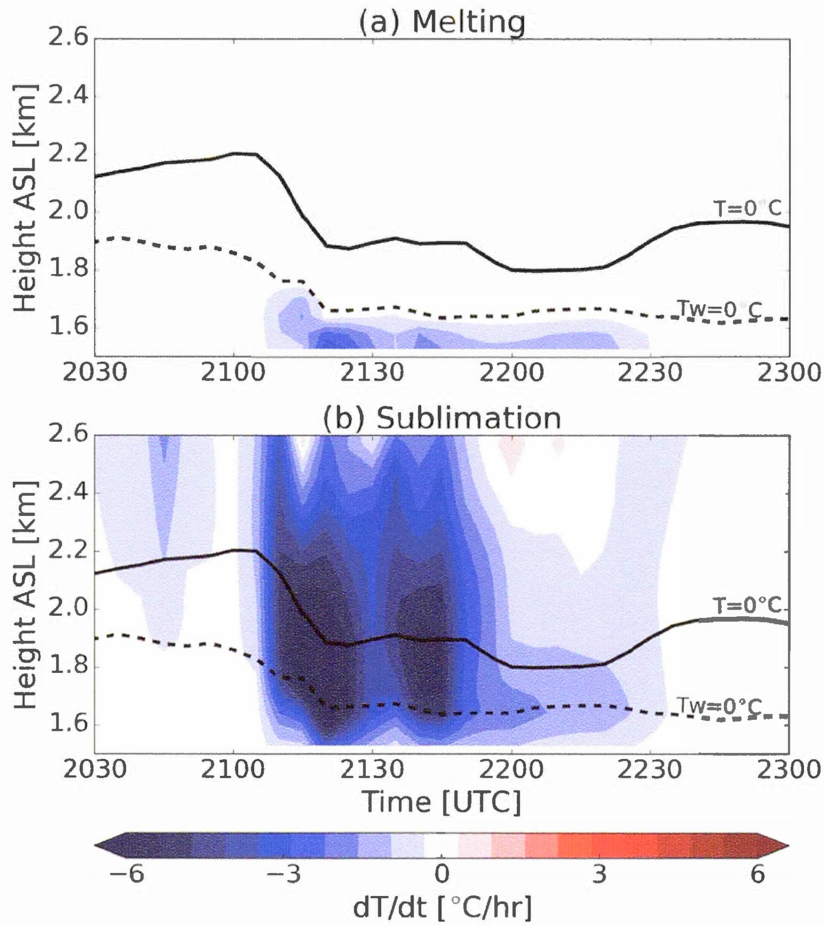

Figure 1.8 Cooling rate (dT/dt) associated with (a) melting and (b) sublimation of snow and graupel above the KES site on 31 March 2015 for the control run. The solid line indicates the height of the 0°C isotherm and the dashed line indicates the height where the wet-bulb temperature is 0°C. The red color indicates an area of heating due to vapor deposition.

of melting is confined within the melting layer near the surface, which is located where the wet-bulb temperature is above 0°C. Furthermore, the cooling effect of the sublimation of graupel and snow can be seen up to a higher altitude since that process can take place at any temperatures. In particular, Figure 1.8 shows that cooling due to sublimation of snow and graupel is greater near the non-melting layer located between $T = 0$°C and $T_w = 0$°C. This is consistent with the fact that the latent heat of sublimation is seven times higher than the latent heat of fusion. Results from this subsaturated case study suggested that sublimation of snow and graupel had a greater impact on the environmental temperature compared to the melting process.

1.4.3 Impacts of sublimation on the valley flow

To study the impacts of phase changes on the valley flow, two sensitivity tests were conducted. First, the MLT experiment was conducted by removing the effect of melting of snow and graupel in the temperature tendency from the CTL run, as shown in Table 1.1. Second, the SBL experiment was conducted in a similar manner by removing the effect of sublimation of snow and graupel in the temperature tendency.

The SBL experiment showed that the precipitation rate at the surface for rain was higher than in the CTL and MLT runs, especially at 2140 UTC (Fig. 1.9 a). The snow and graupel rates were, however, lower in the SBL run, except from 2130 to 2145 UTC, when the graupel rate was higher. Relative humidity was lower in the SBL run before 2140 UTC, but after the peak in rain precipitation, the relative humidity was similar to the CTL and the MLT runs between 70-80% (Fig. 1.9 b). The 0°C isotherm was higher in the SBL experiment for the whole period while the rain-snow boundary was present at the surface (Fig. 1.9 c). The amount of snow and graupel mass aloft was higher in the SBL experiment (Figs.

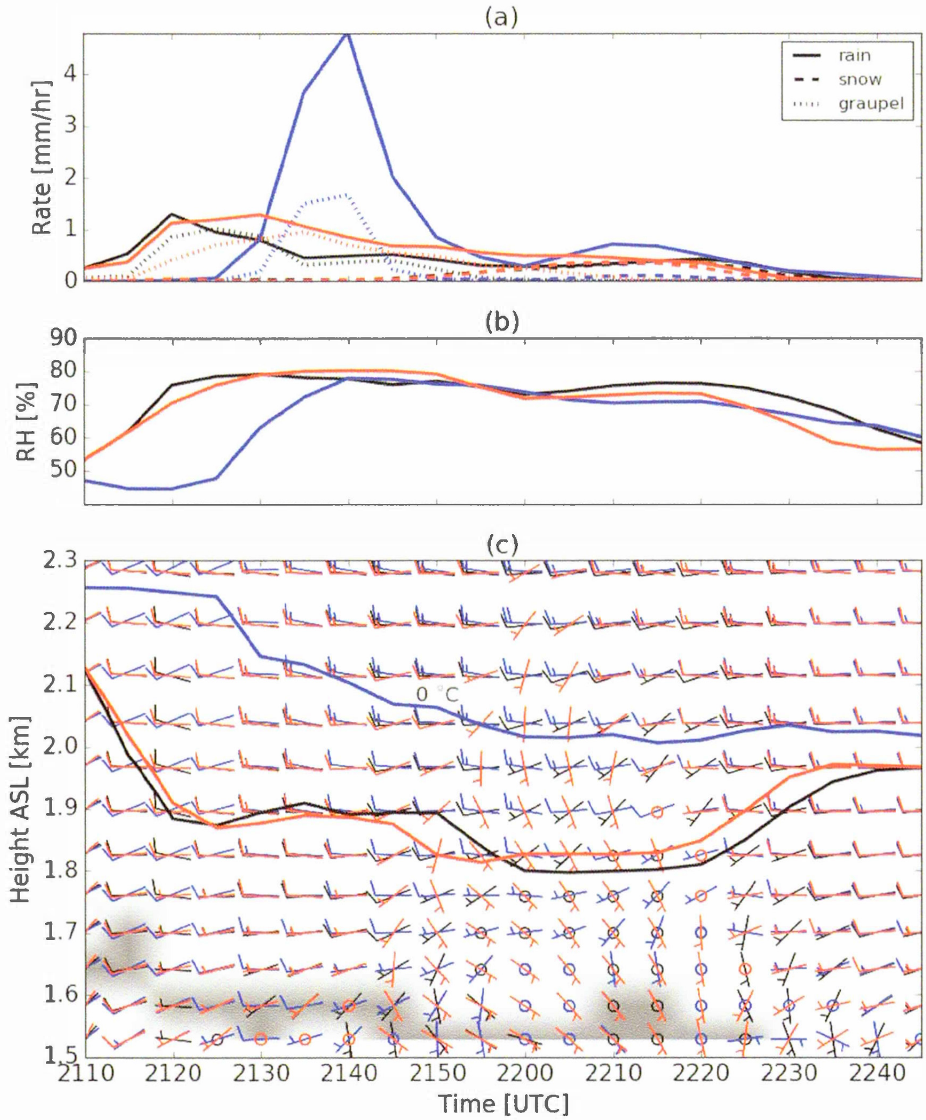

Figure 1.9 Time evolution of (a) the precipitation rate and (b) the relative humidity at the surface, along with (c) the temperature, wind speed and wind direction above the KES site on 31 March 2015. These results are for the control run (black line), the run without the cooling effects of sublimation (blue line) and the run without the cooling effects of melting (orange line). The gray area in (c) is the region where the rain-snow boundary is located.

1.10 a and b). The height where $T_w = 0°C$ was higher in the SBL experiment than in the CTL run during the time when precipitation reached the surface and more melting resulted near the surface producing more rain (Fig. 1.10 c).

Neglecting the cooling due to sublimation in the SBL run resulted in a higher temperature at both the surface and aloft (Fig. 1.9 c), because sublimation occurred within a thicker layer compared to melting, which is near the surface (Fig. 1.8). This higher temperature aloft in the SBL run increased the quantity of snow and graupel and the results suggested that vapor deposition was more effective. In mixed phase clouds, when the air temperature is below -12°C, the difference between saturation vapor pressure over water and saturation vapor pressure over ice increases with increasing temperature (Pruppacher and Klett, 2004). This is the basis of the Wegener-Bergeron-Findeisen mechanism. The ice crystals grow by vapor diffusion at the expense of the supercooled water drops. Consequently, more water vapor was available, which produced more ice crystals and eventually more snow. This snow interacted with the cloud droplets to form more graupel which was seen in Fig. 1.10. Beside the higher amount of snow and graupel, more ice particles and fewer cloud droplets were present in the SBL run while there was less water vapor in the atmosphere just before 2130 UTC. Since the $T_w = 0°C$ isotherm was higher in the SBL experiment, melting occurred on a thicker layer between 2130 and 2230 UTC and thus producing a higher rain rate at the surface. Results from the sensitivity experiments confirmed that sublimation of snow and graupel had a larger impact on meteorological parameters such as temperature, relative humidity and precipitation rate during the presence of the rain-snow boundary compared to melting.

Figure 1.9 c shows the wind speed, wind direction and the air temperature at the KES site for the CTL, the MLT and the SBL runs between 2110 and 2245 UTC. The impact of sublimation and melting on the temperature and horizontal wind

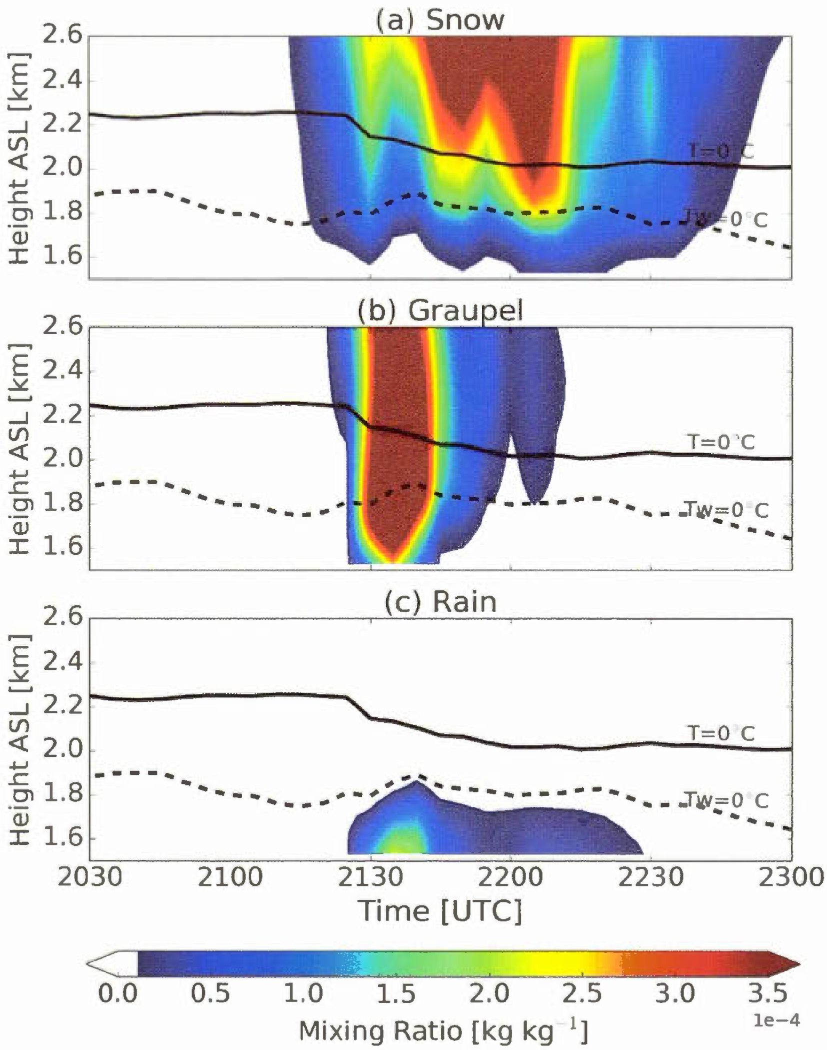

Figure 1.10 Time evolution of (a) snow, (b) graupel and (c) rain mixing ratio at the KES site on 31 March 2015 for the run without the cooling effects of sublimation. The solid line indicates the height of the 0°C isotherm and the dashed line indicates the height where the wet-bulb temperature is 0°C.

direction is shown. Snow and graupel sublimation produced a stronger cooling of the air than melting near the surface at KES site. The 0°C isotherm stayed at a higher altitude between 2110 and 2245 UTC in the SBL run compared to the CTL and MLT runs. Starting at 2130 UTC, a shift in the wind direction from south-west to south-east at 2230 UTC was seen in the CTL run below 1.7 km ASL. From 2130 to 2200 UTC, the shift in the wind direction for the SBL run was of less importance compared with the CTL and MLT runs, suggesting that the cooling due to sublimation may be an important factor in the wind reversal during this period. The wind direction was similar in the three simulations at 2245 UTC and this time coincides with the end of precipitation at the surface at the KES site (Figs. 1.9 a and c). The change in the horizontal wind direction was also observed during the field campaign at the KES site. After 2115 UTC, a change in the wind direction from west to east was seen in the CTL run and a change from south-west to north-east was observed at the Alberta Environment weather station (Fig. 1.11). After the precipitation episode, the wind switched back to its original direction.

This impact of cooling from the sublimation of snow and graupel is investigated from another perspective across the valley and is presented in Figure 1.12 for the CTL and the SBL runs. The Kananaskis valley is oriented north-south, so a west-east cross-section was chosen to show the wind direction across the valley. The cross-section of the Kananaskis valley at 2150 UTC 31 March 2015 showed that the west-east component of the horizontal winds in the CTL run (Fig. 1.12 a) changed direction on the west facing slope and the wind reversal was almost filling up the bottom of the Kananaskis valley. The wind direction switched from upslope to downslope in the CTL run, while it just slowed down in the SBL run at that time.

The change in wind direction due to sublimation and melting can be explained

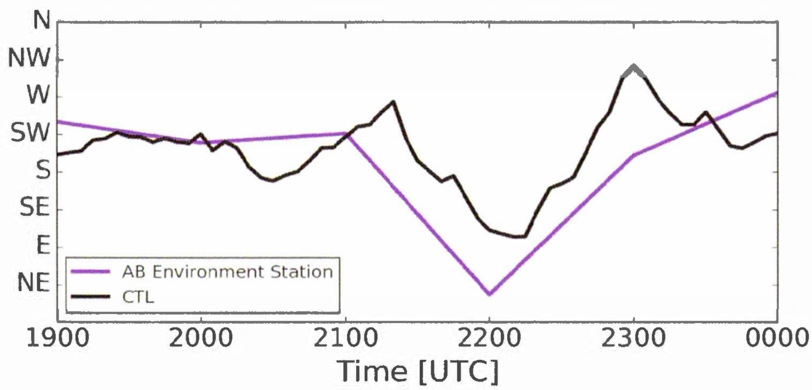

Figure 1.11 Time evolution of the wind direction at KES site on 31 March 2015 for the control run (black line) and for the data from the Alberta Environment weather station located at KES site (purple line).

using the thermodynamic equation. It relates the horizontal advection, vertical motion, diabatic processes and the temperature tendency. Changing one of the parameters will affect another. Sublimation and melting of snow and graupel extract energy from the environment, inducing a cooling. The denser air moves down and, in combination with the presence of the slope on the mountainside, affects the horizontal wind direction. The effect on the winds is less at higher altitudes when we move away from the mountainside. The next section discusses the possible mechanisms responsible for the valley flow reversal.

1.4.4 Discussion on the mechanisms of the valley flow reversal

In a large-scale environment, the flow reversal and downslope flow may be a consequence of either dynamical blocking or locally generated negative buoyancy caused by cooling from diabatic processes (Carbone et al., 1995; Steiner et al., 2003). Dynamical blocking is related to the kinetic energy in the upstream flow

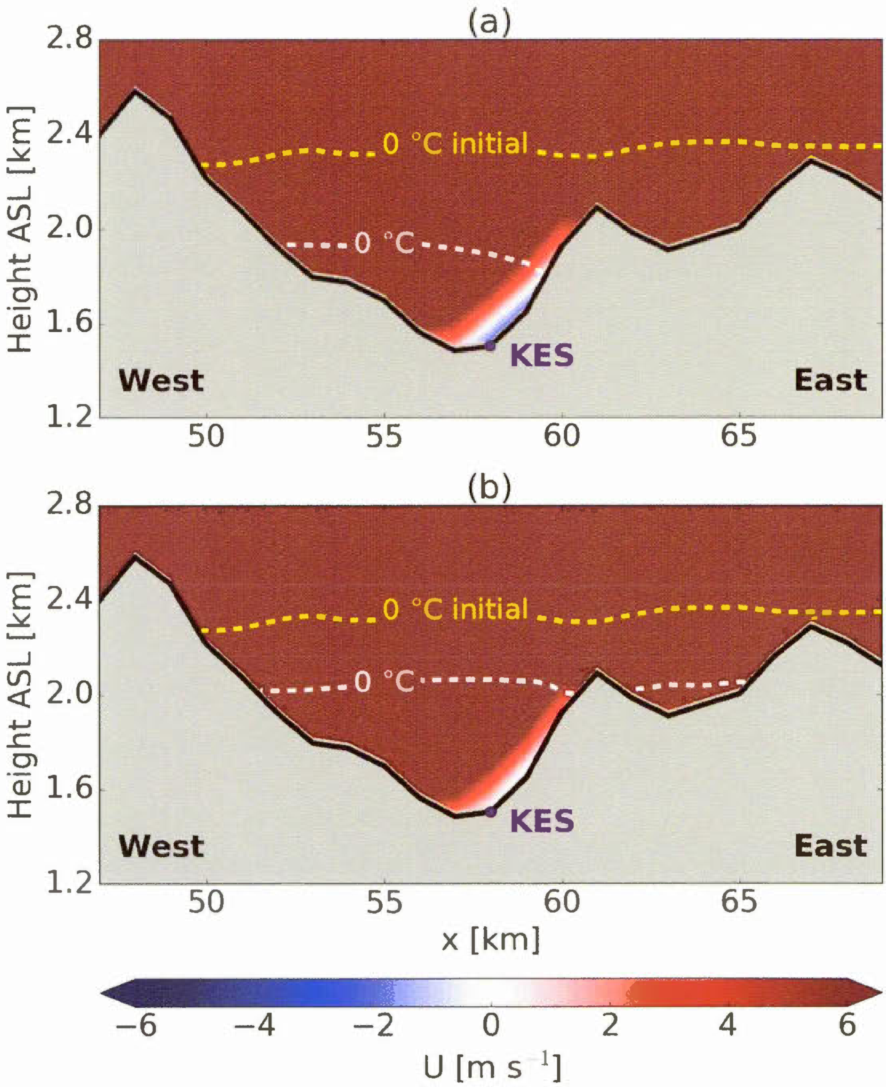

Figure 1.12 Wind speed (west-east wind component) at 2150 UTC 31 March 2015 for (a) the control run and (b) the run without the cooling effects of sublimation. The orange line indicates the position of the 0°C isotherm at 1800 UTC and the white line indicates its position at 2150 UTC.

and the potential energy produced through the lifting of stratified air. The forced lifting creates an accumulation of mass on the windward side of the mountain resulting in high-pressure anomalies that can cause the flow to decelerate, stagnate, or reverse. Alternatively, cooled air from diabatic processes is heavier than its environment and thus starts to subside. This air will then concentrate in the valley.

To assess the evolution of the stratification of the atmosphere in the Kananaskis valley, different atmospheric profiles from the CTL run are represented in Figure 1.13 for the lowest 1 km above ground level. As snow and graupel started falling, sublimation and melting cooled the air (Figs. 1.13 a and b). The profile of the potential temperature θ was mostly neutral at 2100 UTC and became slightly more stable at 2230 UTC (Fig. 1.13 c). However, in the lower 150 m, the profile was mainly unstable. The Brunt-Vaisala frequency also showed this evolution in the atmospheric profile from neutral to stable (Fig. 1.13 d). In the SBL run, the θ profile stayed neutral and the Brunt-Vaisala frequency showed a more stratified atmospheric profile at 2230 UTC (Figs. 1.13 g and h). However, the SBL run showed lower values of N^2 below 2.4 km ASL than in the CTL run at 2230 UTC. These results suggested that sublimation of snow and graupel affected the air stability and may have influenced the reversal of the valley flow.

The Froude number (Fr) is a dimensionless parameter relating the wind, static stability and topography: $Fr = U/Nh$, where U is the wind speed, N is the Brunt-Vaisala frequency and h is a characteristic height of the mountain obstacle (Carbone et al., 1995). The Froude number was calculated from 2110 to 2230 UTC averaging the horizontal winds and the Brunt-Vaisala frequency from the surface to 2.5 km ASL, and using $h = 800$ m. Typically, a flow with $Fr < 0.5$ has insufficient kinetic energy to surmount an obstacle (Carbone et al., 1995). The results showed that the Froude number stayed above 0.5, which suggested that the

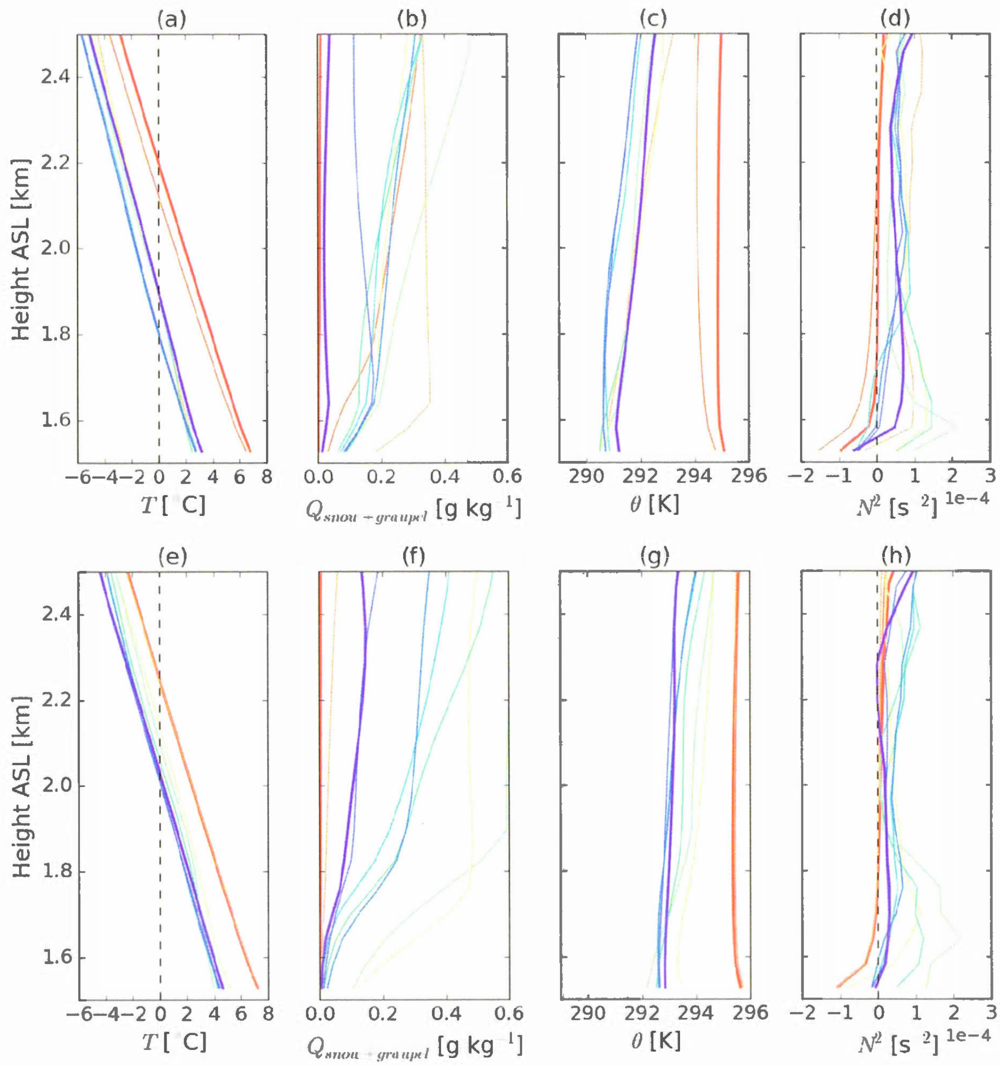

Figure 1.13 Lower atmospheric profiles from the KES site at every 10 min from 2100 to 2230 UTC. Earlier times are in red and shifts towards blue. (a) to (d) are for the control simulation and (e) to (h) are for the run without the cooling effects due to sublimation. In particular, (a) and (e) is the temperature, (b) and (f) is the mixing ratio for snow plus graupel, (c) and (g) is the potential temperature, (d) and (h) is the Brunt-Vaisala frequency.

production of negative buoyancy may have been the main mechanism responsible for the flow reversal in the Kananaskis valley and not dynamical blocking.

1.5 Impacts of accretion aloft

1.5.1 Experiment with no graupel

During the Alberta Field Project, graupel was often observed in the Kananaskis valley. Graupel is formed by accretion of supercooled droplets that freeze on the surface of solid hydrometeors. Accretion leads to an increase of the environmental temperatures. To investigate the effect of the presence of graupel, further simulation was performed assuming no graupel formation (NOGRP).

Results showed that one of the main differences between the NOGRP and the CTL runs was that the snow rate was higher in the NOGRP run, especially between 2115 and 2215 UTC, because no snow had been converted to graupel (Fig. 1.14 a). Relative humidity was lower at almost all times for the NOGRP run compared to the CTL run (Fig. 1.14 b) while the height of the 0°C isotherm was slightly higher in the NOGRP run between 2100 and 2230 UTC (Fig. 1.14 c). Since no graupel were produced in the NOGRP run, less accretion was happening at higher altitudes, and thus, less diabatic heating, which could explain the lower surface temperatures. The rain-snow boundary was present between 2115 and 2145 UTC with a width of 200 m. No prominent change in the horizontal wind direction can be seen in the results from the NOGRP run (Fig. 1.14 c). These results suggested that less diabatic heating aloft in the NOGRP run and similar cooling from sublimation at lower altitudes could explain the absence of change in the wind direction. Diabatic heating aloft combined with diabatic cooling at lower altitudes could have modified the temperature profile to a more stable profile, thus generating vertical motion or subsidence in the CTL run.

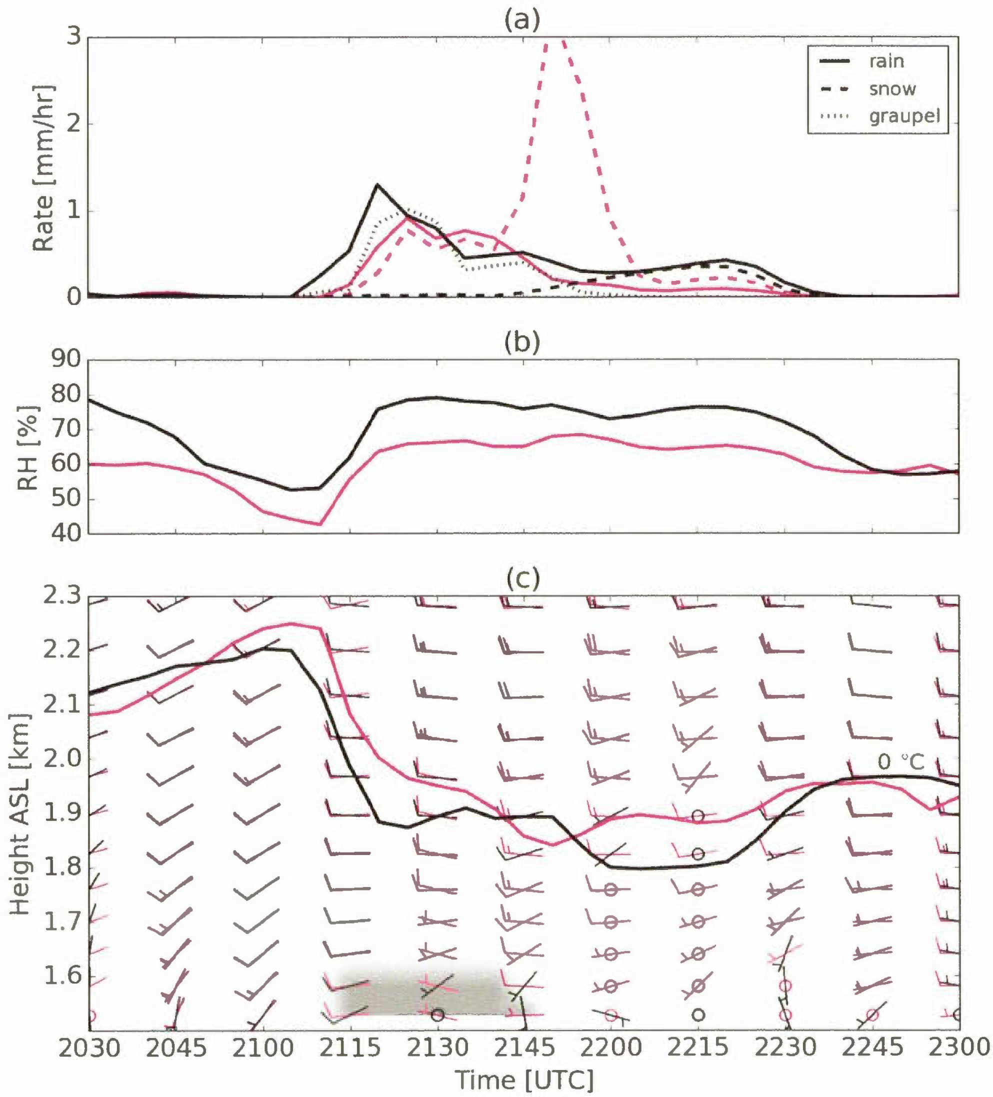

Figure 1.14 Time evolution of (a) the precipitation rate and (b) the relative humidity at the surface, along with (c) the temperature, wind speed and wind direction above the KES site on 31 March 2015. These results are for the control run (black line) and the run assuming no graupel formation (pink line). The gray area in (c) is the region where the rain-snow boundary is located.

The evolution of the cooling associated with melting and sublimation of snow for the NOGRP run (Fig. 1.15) indicated that the width of the melting layer decreased rapidly after 2130 UTC from 300 m to about 60 m while the width stayed constant at 200 m from 2130 to 2230 UTC in the CTL run (Fig. 1.8). The cooling due to sublimation lasted for a similar period in the CTL and NOGRP runs, between 2115 and 2230 UTC. However, there was more cooling between 2130 and 2215 UTC in the NOGRP run when a higher snow rate was seen at the surface (Figs. 1.14 a and 1.15 b). The cooling was slightly more extensive in the NOGRP run at altitudes higher than 2 km ASL. In particular, at 2200 UTC the cooling due to sublimation was weaker in the CTL run and heating due to vapor deposition was present below 2.6 km ASL, which was not shown in the NOGRP run. A hypothesis to explain the peak in precipitation rate and cooling rate at a later time in the NOGRP run is that snowflakes have a much lower terminal velocity than graupel. Having no graupel shifts the accumulation of precipitation at a later time.

The results from this experiment suggested that the presence of graupel on 31 March 2015 may have influenced the change in wind direction in the Kananaskis valley shown in the CTL run. The diabatic heating due to the conversion of graupel aloft combined with the cooling due to sublimation at lower altitudes could have increased the stratification of the air thus producing subsidence that contributed to the wind reversal in the valley. Also, in the absence of graupel, a higher snow rate was seen at the KES site since no graupel was converted to snow.

1.5.2 Role of accretion

To better estimate the temperature feedbacks associated with accretion above the rain-snow boundary, the sink-terms in Eq. 1.1 that are associated with graupel

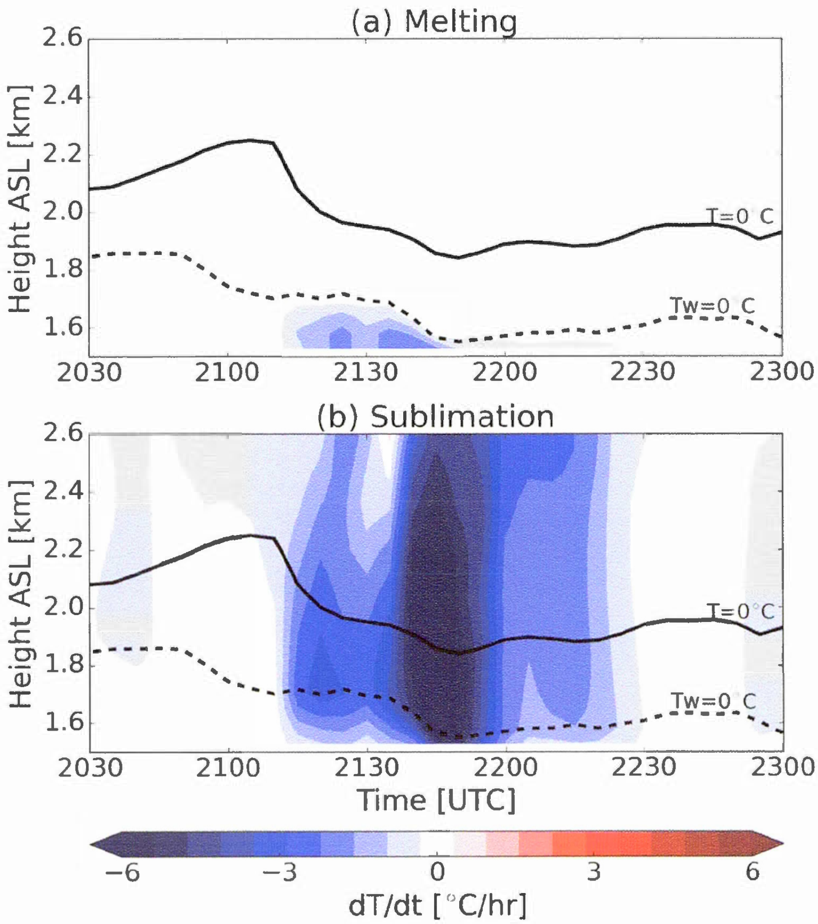

Figure 1.15 Cooling rate (dT/dt) associated with (a) melting and (b) sublimation of snow above the KES site on 31 March 2015 for the run assuming no graupel formation. The solid line indicates the height of the 0°C isotherm and the dashed line indicates the height where the wet-bulb temperature is 0°C.

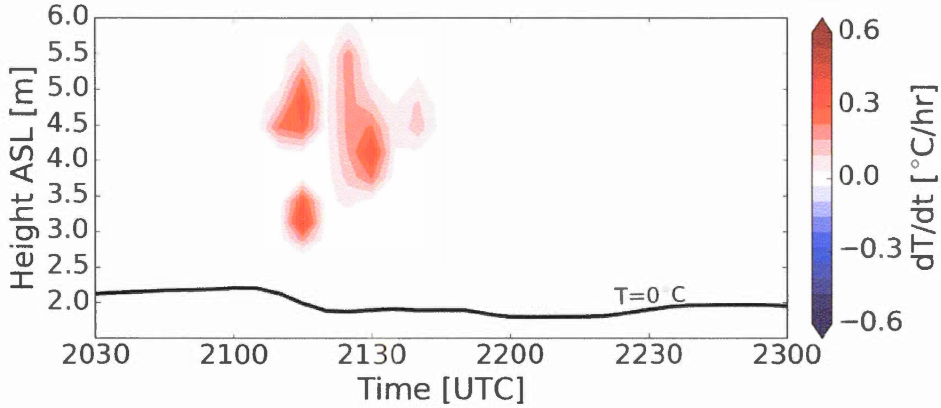

Figure 1.16 Heating rate (dT/dt) associated with accretion above the KES site on 31 March 2015 for the control run. The solid line indicates the height of the 0°C isotherm.

formation have been added together for the CTL run and the value of dT/dt associated with accretion is represented in Figure 1.16. These variables are listed in Table 1.1 for the ACR experiment. The results showed a net heating due to accretion from 2115 to 2145 UTC March 31 2015 in the CTL run. The collection of cloud droplets that freeze upon impact produces a heating of the environment. However, this heating is limited to heights between 2.5 and 6 km ASL since graupel is produced at altitudes above the freezing level where supercooled droplets are present in the atmosphere.

To better assess the impacts of this heating source at the surface and on the sublimation of snow and graupel, the ACR run was conducted neglecting the temperature tendency associated with accretion as described in Table 1.1. This experiment was performed to see if the release of latent heat in the environment associated with accretion affected the cooling due to sublimation and melting near the surface and also the wind direction in the valley. Vertical profiles at 2120 UTC

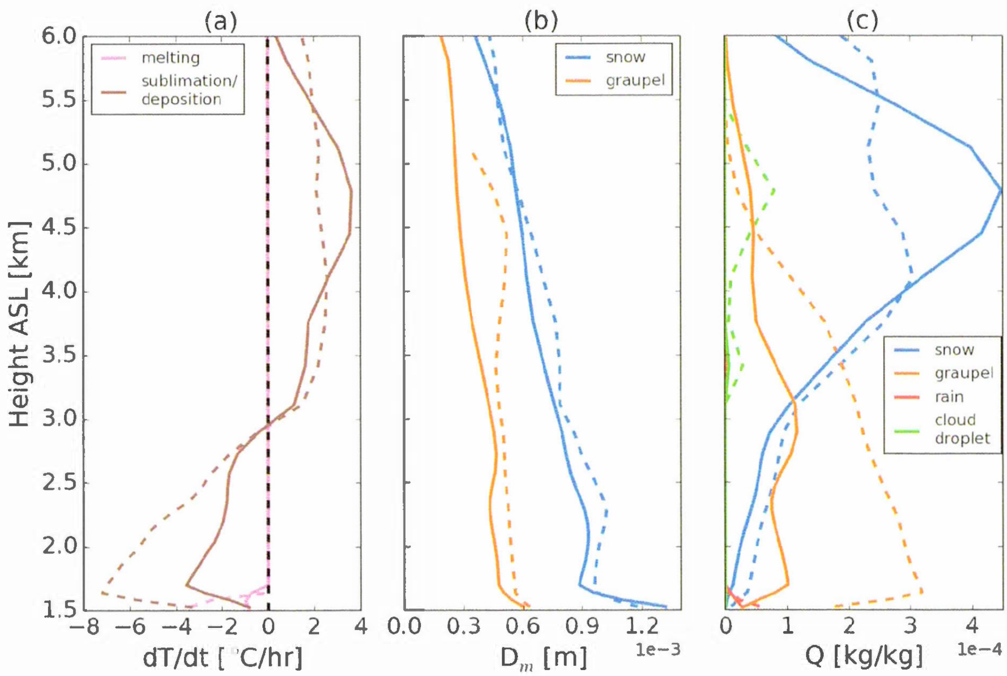

Figure 1.17 Vertical profiles of (a) dT/dt due to melting and sublimation or deposition, (b) mean mass diameter (D_m) of snow and graupel, and (c) mixing ratio (Q) of snow, graupel, rain and cloud droplet at 2120 UTC 31 March 2015 above the KES site for the run without the heating effect of accretion (solid line) and the control run (dashed line).

31 March 2015 for the CTL and the ACR runs at the KES site are shown in Figure 1.17. At this time, the CTL run showed the strongest heating due to accretion at higher altitudes (Fig. 1.16).

In the ACR run, the cooling effect of sublimation at the surface was lower while the heating due to vapor deposition at higher altitude was slightly higher than in the CTL run between 4 and 5.5 km ASL (Fig. 1.17 a). A study of the mean mass diameter for snowflakes and graupel in Figure 1.17 b indicates that it was smaller in the ACR run compared to the CTL run below 4.5 km ASL and similar above 4.5 km ASL. The vertical profile of the mixing ratios for snow, graupel, rain and cloud droplet presented in Figure 1.17 c indicates that the mixing ratio for graupel was lower in the ACR run except above 4.5 km ASL. The mixing ratio for snow was much higher between 4 and 5.5 km ASL in the ACR run compared with the CTL run. Cloud droplets were present between 3 and 5.5 km ASL in the CTL run while there were almost no cloud droplets present in the ACR run. The mixing ratio for rain was lower at the surface in the ACR run than it was seen in the CTL run.

The wind direction at the KES site is shown on Figure 1.18 for both runs. The wind direction showed a very slight difference between the CTL and the ACR runs. The direction changed from west to east from 2115 to 2300 UTC in the CTL run while the wind changed from south-west to south-east from 2145 to 2230 UTC in the ACR run. The results showed that accretion aloft increased the duration of the wind reversal at the KES site in the CTL run by 30 minutes.

The results from the ACR experiment where dT/dt associated with accretion is zero suggested that the presence of cloud droplets at high altitudes in the CTL run allowed the collection of cloud droplets and converted snow to graupel. More vapor deposition in the ACR run between 4 and 5.5 km ASL could explain the higher concentration of snow mass at this same altitude. The results show clearly the relationship between the heating due to accretion at high altitudes and the mass of graupel at lower altitudes. No heating due to accretion in the ACR run showed that the formation of graupel was less efficient. The higher amount of graupel mass below 3.5 km ASL in the CTL run explains the higher cooling rate due to sublimation near the surface in the CTL experiment. These results suggested that heating due to accretion aloft could affect the cooling due to sublimation at lower altitudes as well as the precipitation rate at the surface.

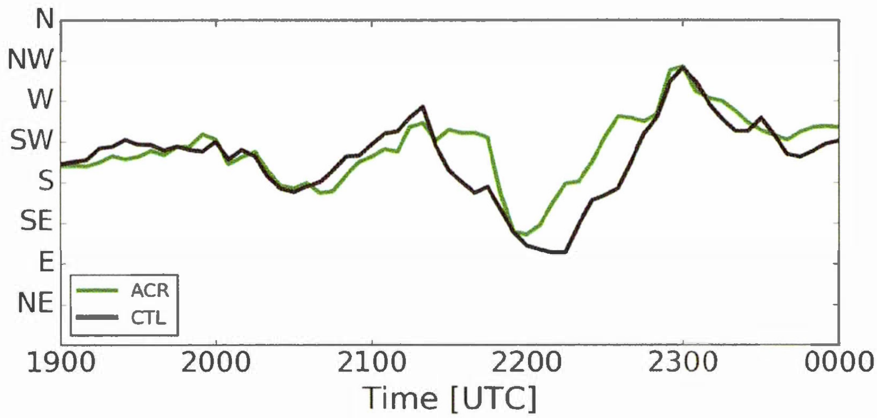

Figure 1.18 Wind direction at the KES site on 31 March 2015 for the control run (black line) and the run without the heating effect of accretion (green line).

The effect of heating due to accretion on the wind direction was increasing the duration of the wind reversal. The NOGRP run showed a strong impact on the wind direction and the ACR run showed a very slight impact. This suggested that accretion and the conversion of snow to graupel were not the main factor affecting the wind reversal on 31 March 2015 in the Kananaskis valley.

1.6 Discussion

1.6.1 Atmospheric conditions during the Alberta Field Project

During the Alberta Field Project, snow was often observed at surface temperatures above 0°C at the KES observation site. Snow was observed at temperatures up to 9°C with relative humidities below 50%. Most of the events with rain at KES or mixed precipitation were associated with a general flow field coming from the west and the events with snow were mainly associated with a flow field from the east (Vaquer, 2017). The results from this study and also from Hung (2016) showed that highly rimed aggregates and graupel were the main solid precipitation types observed at the KES site during the Alberta Field Project. They found that sublimation was present during both type of events and especially when graupel was observed. The conditions at the surface were drier and more unstable in the events with the main flow coming from the west. These events also showed more accretion.

For the 31 March 2015 case study, snow was present at temperatures up to 3°C and winds were coming from the west, which produced mixed precipitation at the KES site and this was consistent with the results from the CTL simulation. The present study demonstrated that considering the dry climate in the area of the Kananaskis valley, phase change and also the presence of graupel both influenced the weather conditions at the KES site. During the field campaign, seven events were similar to the storm of 31 March 2015 and showed a general flow field coming from the west, along with the presence of a rain-snow boundary at the surface or aloft.

On 28 March 2015, winds were mainly from the west and rain was seen at the KES site. To put the results obtained in perspective, the 31 March case is compared in the next section with the 28 March case. In the first case, the conditions were subsaturated both at the surface and aloft, while in the second case, the conditions were subsaturated at the surface and saturated aloft.

1.6.2 Rain-snow boundaries in the Kananaskis valley

The rain-snow boundary on 31 March 2015 was located at the surface and the width varied between 150 to 250 m. Mixed precipitation were seen at the surface and the conditions were subsaturated both at the surface and aloft. Results suggested that cooling due to sublimation was greater and had a larger impact on meteorological parameters such as temperature, relative humidity and precipitation rate compared to melting. Cooling due to sublimation influenced the change in flow direction from upslope to downslope and this reversal was observed during the field campaign in the Kananaskis valley.

Results from the 31 March 2015 case study showed that cooling due to sublimation of snow and graupel affected the wind direction and the surface precipitation rate at the KES site. These results were compared to another case study on 28 March 2015 where a rain-snow boundary was present a few hundred meters above the KES site. This case study was chosen because a rain-snow boundary was present at a higher altitude than for 31 March 2015 and only rain was seen at the surface, which means more melting could have occurred in this case.

The CTL run for 28 March 2015 showed a rain-snow boundary 150 m above the KES site between 1100 to 1400 UTC and the width varied from 500 m to 800 m during this period. The precipitation type seen at the surface was rain only and the environment was also subsaturated near the surface while the conditions were near saturation aloft. Figure 1.19 shows the time evolution of the mixing ratio for snow, graupel and rain above the KES site during this event. The amount of snow was much higher aloft than the amount of graupel for the 28 March event compared to the 31 March event. Considering the lowest amount of graupel for 28 March 2015, the ACR and NOGRP runs were not performed for this case.

The SBL and MLT experiments were performed for the case study of 28 March 2015. Results showed that the surface temperatures were higher in the SBL run compared to the CTL and MLT runs (Fig. 1.20 a). The relative humidity was slightly lower in the SBL run compared to the CTL and MLT runs (Fig. 1.20 b). Between 1130 and 1200 UTC, the results showed that the wind direction changed from south to north in the CTL run at 1100 UTC for a brief period of time, while it stayed in the same direction in the MLT run and changed to north-east in the

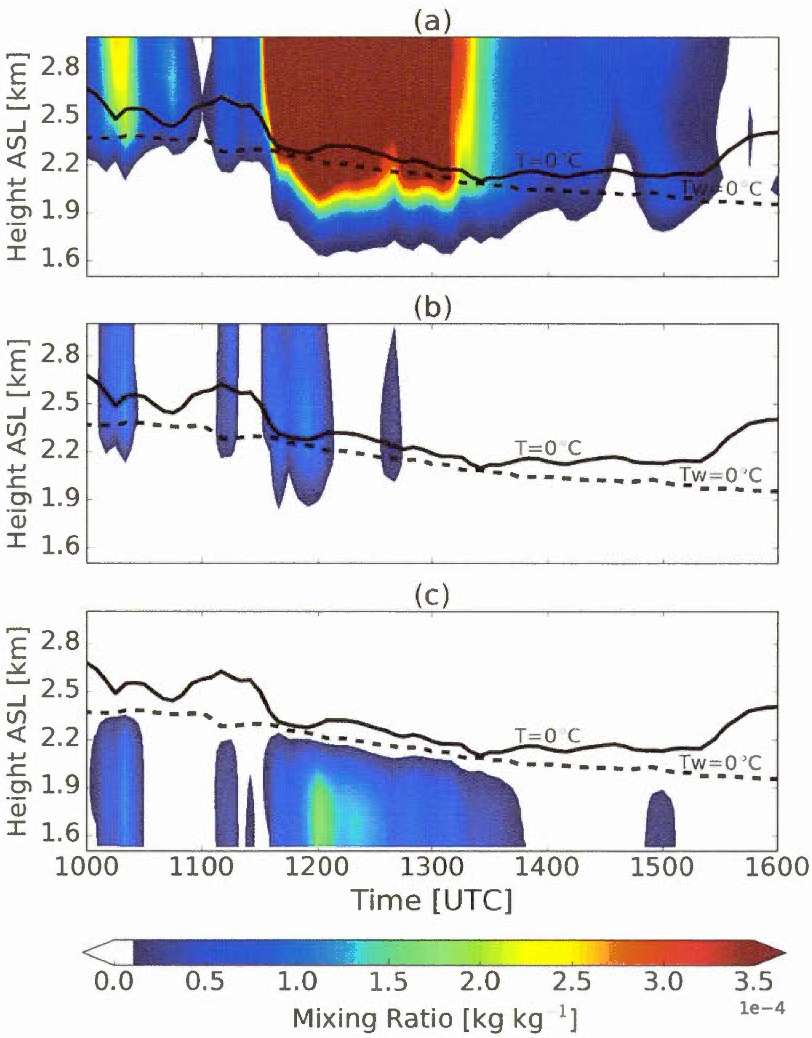

Figure 1.19 Time evolution of (a) snow, (b) graupel and (c) rain mixing ratio at the KES site on 28 March 2015 for the control run. The solid line indicates the height of the 0°C isotherm and the dashed line indicates the height where the wet-bulb temperature is 0°C.

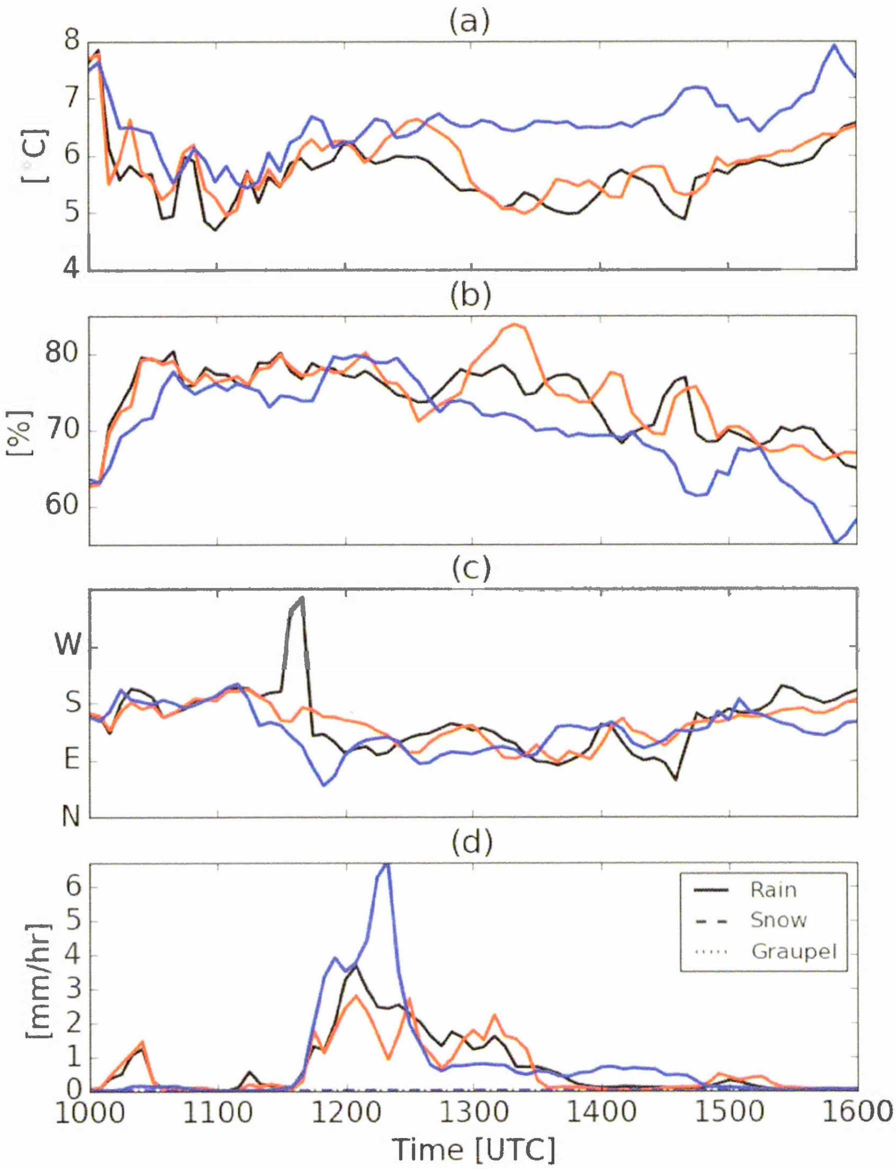

Figure 1.20 Time evolution of (a) the temperature, (b) the relative humidity, (c) the wind direction and (d) the precipitation rate at the KES site on 28 March 2015 for the control run (black line), the run without the cooling effects of sublimation (blue line) and the run without the cooling effects of melting (orange line).

SBL run (Fig. 1.20 c). The precipitation rate at the surface for rain in the MLT experiment was similar to the CTL run, while the rate in the SBL experiment was higher between 1145 and 1230 UTC (Fig. 1.20 d).

While the precipitation rate was much higher at the surface during this event compared to the event on 31 March 2015, the results suggested that the cooling due to sublimation had less impact than melting on the wind direction and the precipitation at the surface. This could be explained by the rain-snow boundary being much thicker above the KES site and more melting being produced near the surface, along with the conditions aloft near saturation.

1.7 Conclusion

This study examined the weather conditions associated with a rain-snow boundary in the Kananaskis valley, Alberta. The weather events in this area often occur in dry conditions and this is why this study looked at sublimation. In particular, the experiments looked at the relative importance of snow and graupel sublimation and melting on the evolution of weather events. The temperature feedbacks associated with sublimation and melting on the valley flow were also examined. The impacts of graupel and also the temperature feedbacks associated with accretion were studied.

To address these issues, numerical simulation using the WRF model were used to simulate a case study from the Alberta Field Project. This field campaign, which took place in spring 2015 in the Kananaskis valley, Alberta, provided various weather observations that were compared to the numerical simulations to ensure a good representation of the weather event by the numerical model.

This study led to the following conclusions:

— Mixture of rain and solid precipitation can be present at surface air temperatures above 0°C in the Kananaskis valley, Alberta. This is mainly because the environment is often subsaturated in this area. For example, the 31 March 2015 case study indicated the presence of mixed precipitation at the surface and snow was noted at temperatures up to 3°C with subsaturated conditions both at the surface and aloft. Furthermore, rain was present at the surface on 28 March 2015 at temperatures just above 4°C with subsaturated conditions near the surface and saturated conditions aloft. The rain-snow transition was located at the surface on 31 March but was about 150 m above the surface on 28 March 2015.

— The cooling effect due to sublimation of snow and graupel was more important than melting for the subsaturated case study of 31 March 2015. This was not the case when the air was near saturation on 28 March 2015.

— The cooling effect of sublimation and melting affected the precipitation rate at the KES site and influenced the change in flow direction from upslope to downslope on the west facing slope of the mountain. This result is consistent with the findings of Steiner et al. (2003) and Thériault et al. (2015) who showed that cooling by melting of snow influenced the reversal of the flow on a mountainside under saturated conditions.

— The presence of graupel in the numerical simulation during the weather event of 31 March 2015 influenced the wind direction reversal near the KES site. Further investigation will be necessary to determine the processes associated with the presence of graupel that could impact the flow direction in the valley.

— The heating due to accretion aloft during the formation of graupel could affect the cooling due to sublimation at lower altitudes and also influence the precipitation rate at the surface. The heating at high altitudes influenced the solid precipitation size and mass mixing ratios. It also impacted the cooling rates due to phase change near the surface. Accretion aloft only contributed to increase the duration of the wind reversal in the valley and was not the main driving factor for the reversal in the present case study.

This study had some limitations. First, due to some instrumentation issues, the measurements of wind speed and wind direction during the Alberta Field Project were sometimes inconsistent. Second, due to the wind effect on the precipitation gauge, it was difficult to get an accurate snowfall measurement. Third, this study also had some numerical limitations due to the choice of microphysics parameterization in the WRF model. The use of another microphysics scheme like the Thompson scheme (Thompson et al., 2008) produced a higher snow rate and a smaller graupel rate for the 31 March 2015 case study. Different microphysics scheme would produce different precipitation rate and thus affect the cooling rate associated with sublimation and melting.

Future studies could investigate the impacts of microphysics parameterization on graupel formation and its evolution, as well as accretion. Since sublimation of particles is sometimes critical, further study should be conducted on the shape of the particles falling through the subsaturated layer. Finally, more field measurements of precipitation types, microphysics characterization aloft and at the surface, along with atmospheric conditions should be conducted in this area as it is often associated with extreme precipitation and drought events.

Overall, this study showed that cooling due to sublimation and also the presence of graupel can affect the intensity and also the duration of precipitation, along with influencing the change in the flow direction in a valley. Microphysical processes in the vicinity of rain-snow boundaries are very important to understand since they can lead to extreme meteorological events in mountainous regions.

Acknowledgments

We would like to thank the Changing Cold Region Network (CCRN) funded through the Climate Change and Atmospheric Research (CCAR) Initiative of the Natural Sciences and Engineering Research Council of Canada (NSERC) for providing financial support to accomplish this work. One of the co-author (EP) would like to acknowledge NSERC, the Fonds de recherche Nature et technologies du Québec (FRQNT) and Environment and Climate Change Canada for a graduate fellowship. We would also like to acknowledge all the participants to the Alberta Field Project and a special thanks to Paul Vaquer and Ida Hung for sharing data from the field campaign.

1.8 Appendix A: Parameterization of snow sublimation

In the original Milbrandt and Yau two-moment scheme, snow sublimation is allowed only when air temperature is below 0°C, while graupel sublimation is allowed at all temperatures. Since the area of interest for this study is located in an environment often subsaturated, the scheme was modified to allow snow to sublimate at temperatures above 0°C. Note that graupel is allowed to sublimate at all temperatures, so the same steps were followed for snow.

The equation to calculate the rate of sublimation of snow from Milbrandt and Yau (2005) is:

$$QVD_{vs} = \frac{1}{AB_i}\left[\, 2\pi(S_i-1)N_{0s}VENT_s - \frac{L_sL_f}{K_aR_vT^2}QCL_{cs}\,\right] \tag{1.2}$$

where

$$AB_i = \frac{L_s^2}{K_aR_vT^2} + \frac{1}{\rho q_{is}\psi} \tag{1.3}$$

is the thermodynamic function. Also, S_i is the saturation ratio with respect to ice, N_{0s} is the intercept parameter for snow, $VENT_s$ is the mass-weighted ventilation factor (Ferrier, 1994), K_a is the thermal conductivity of air, R_v is the gas constant for water vapor, T is the temperature of air, ρ is the density of air, q_{is} is the saturation vapor mixing ratio with respect to ice and ψ is the diffusivity of water vapor in air.

The sublimation rate equation was moved in the microphysics scheme so that snow and graupel sublimation are computed in the same conditions, which is at all air temperatures. The function *polysvp* was also corrected in the microphysics scheme in order to calculate the saturation vapor pressure properly at all temperatures.

The CTL run using the modified version of the Milbrandt and Yau scheme for snow sublimation and saturation vapor pressure calculation was compared to another run using the original version of the scheme. Figure 1.21 shows the cooling rate associated with sublimation of snow for the original Milbrandt and Yau scheme (Fig. 1.21 a) and for the modified version of the scheme (Fig. 1.21 b). The results show that cooling due to sublimation is now present below the 0°C isotherm. The use of the modified version of the scheme produced more sublimation and thus more cooling. Following these results, the modified Milbrandt and Yau scheme to allow snow sublimation above 0°C and with the corrected calculation for saturation vapor pressure was used in all the simulations presented in this study since the area of interest is located in an area often subsaturated.

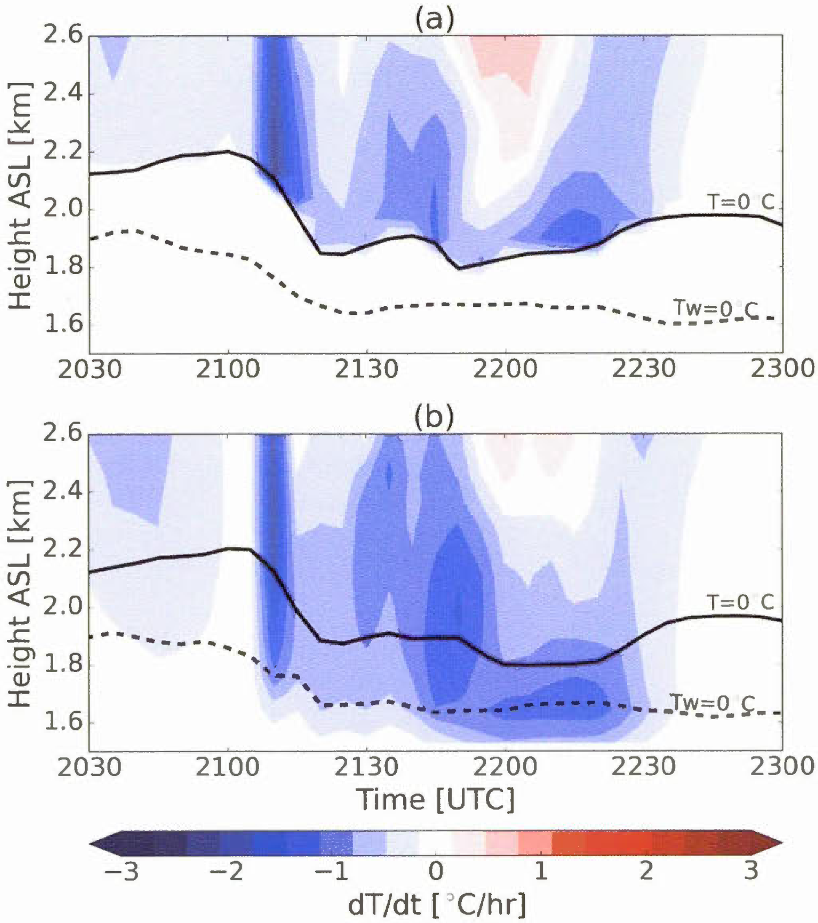

Figure 1.21 Cooling rate (dT/dt) associated with the sublimation of snow for (a) the original Milbrandt and Yau scheme and (b) the modified scheme.

CONCLUSION

Cette étude a examiné les conditions météorologiques associées à une ligne pluie-neige dans la vallée de Kananaskis en Alberta. Les événements météorologiques dans cette région surviennent souvent dans des conditions sèches et c'est pourquoi la sublimation a été étudiée ici. Des tests de sensibilité ont regardé l'importance relative de la sublimation et de la fonte de la neige ainsi que de la neige roulée sur l'évolution des événements météorologiques. Les effets de la variation de température associés à la sublimation et à la fonte sur l'écoulement dans la vallée ont aussi été examinés. L'impact de la présence de neige roulée et de la variation de température associée avec l'accrétion a aussi été étudié.

Afin d'examiner ces questions, des simulations numériques utilisant le modèle numérique *Weather Research and Forecasting* (WRF) ont été utilisées pour simuler une étude de cas provenant d'une campagne de terrain en Alberta. Cette campagne de terrain qui a eu lieu au printemps 2015 dans la vallée de Kananaskis en Alberta a permis d'amasser des données d'observations météorologiques qui ont ensuite été comparées à des simulations numériques pour s'assurer d'obtenir une bonne représentation de l'événement météorologique par le modèle numérique.

Cette étude a mené aux conclusions suivantes :

— Des lignes pluie-neige peuvent être observées à des températures de l'air à la surface au-dessus de 0 °C dans la vallée de Kananaskis en Alberta. Ceci est principalement dû au fait que l'environnement est souvent sous-saturé dans cette région. Par exemple, dans le cas du 31 mars 2015, des précipitations mixtes ont été observées à la surface et de la neige était présente à des températures allant jusqu'à 3 °C avec des conditions sous-saturées à la surface et en altitude. Par contre, de la pluie a été observée à la surface le 28 mars 2015 avec des températures juste au-dessus de 4 °C, ainsi que des conditions sous-saturées près de la surface et saturées en altitude. La ligne pluie-neige était située à la surface pour le cas du 31 mars et à environ 150 m au-dessus de la surface le 28 mars.

— L'effet de refroidissement associé à la sublimation de la neige et de la neige roulée était plus important que l'effet du refroidissement dû à la fonte pour l'étude de cas sous-saturé du 31 mars 2015. Cela n'a pas été le cas le 28 mars 2015 où l'air était tout près de la saturation.

— L'effet de refroidissement associé à la sublimation et à la fonte a affecté le taux de précipitation au site de KES et a aussi influencé le changement de direction de l'écoulement de montant à descendant sur la pente de la montagne face à l'ouest. Ceci est en accord avec les résultats de Steiner et al. (2003) et de Thériault et al. (2015) qui ont démontré que la fonte de la neige influençait l'inversion de l'écoulement sur la pente d'une montagne dans un environnement saturé.

— La présence de neige roulée dans la simulation numérique de l'événement du 31 mars 2015 a influencé l'inversion de la direction du vent près du site de KES. Des recherches plus approfondies seront nécessaires afin de déterminer le processus associé à la présence de neige roulée qui pourrait influencer la direction de l'écoulement dans la vallée.

— Le réchauffement associé à l'accrétion à haute altitude lors de la formation de la neige roulée pourrait affecter le refroidissement associé à la sublimation à plus basse altitude et aussi influencer le taux de précipitation à la surface. Le réchauffement à haute altitude a influencé la taille et le rapport de mélange de précipitation solide. Il a aussi influencé le taux de refroidissement causé par les changements de phase près de la surface. L'accrétion en altitude a seulement contribué à augmenter la durée de l'inversion du vent dans la vallée et n'était pas le facteur principal responsable de l'inversion pour cette étude de cas.

Cette étude comportait certaines limites. Premièrement, les mesures de la vitesse et de la direction du vent durant la campagne de terrain en Alberta étaient parfois incohérentes à cause de certains problèmes liés aux instruments. Deuxièmement, il était difficile d'obtenir une mesure précise de l'accumulation de neige au sol à cause de l'effet du vent sur la jauge de mesure de la précipitation. Troisièmement, cette étude comportait aussi des limites numériques dues au choix de paramétrisation physique dans le modèle WRF. L'utilisation d'un autre schéma microphysique comme celui de Thompson et al. (2008) a produit un taux de précipitation en neige plus important et un taux de neige roulée plus faible pour l'étude de cas du 31 mars 2015. L'utilisation d'un schéma microphysique différent pourrait produire un taux de précipitation différent et donc affecter le taux de refroidissement associé à la sublimation et à la fonte.

Des études futures pourraient examiner l'impact de la paramétrisation microphysique sur la formation de la neige roulée et son évolution, en plus de l'accrétion associée. Puisque la sublimation des particules est parfois critique, de plus amples études pourraient examiner la forme des particules traversant la couche sous-saturée. Finalement, un plus grand nombre de campagnes de mesures devraient être effectuées dans cette région afin d'observer les types de précipitation, la microphysique en altitude et à la surface, ainsi que les conditions atmosphériques, puisque cette région est souvent associé avec des événements de précipitation extrême et des périodes de sécheresse.

Dans l'ensemble, cette étude a démontré que le refroidissement associé à la sublimation et aussi la présence de neige roulée peuvent affecter l'intensité et aussi la durée de la précipitation, en plus d'influencer le changement de direction de l'écoulement dans la vallée. L'étude des processus microphysiques dans la région d'une ligne pluie-neige est très importante, car ces derniers peuvent mener à des événements météorologiques extrêmes en terrain montagneux.

APPENDICE A

CONFIGURATION DÉTAILLÉE DU MODÈLE

Les simulations présentées dans cette étude ont été effectuées en utilisant le modèle WRF, version 3.7.1 (Skamarock et al., 2008). Ce modèle numérique de prévision est conçu pour la recherche atmosphérique et la prévision opérationnelle. Il est maintenu par le *National Center for Atmospheric Research* (NCAR) au Colorado. Il donne la possibilité d'effectuer des simulations avec des conditions idéalisées et aussi avec des données réelles.

Des simulations réelles en trois dimensions (3D) ont été pilotées par des conditions initiales et des conditions aux frontières provenant des réanalyses *North American Regional Reanalysis* (NARR) du *National Centers for Environmental Prediction* (NCEP) (Mesinger et al., 2006). Une cascade à double sens a été utilisée avec quatre domaines de résolution horizontale différente (27 km, 9 km, 3 km et 1 km) afin de réaliser des simulations à haute résolution sur la vallée de Kananaskis. Le double sens signifie que l'échange d'information entre le domaine à plus basse résolution (27 km) et les domaines à plus haute résolution (9, 3 et 1 km) est bidirectionnel. La Figure A.1 montre la configuration du domaine dans le modèle numérique.

Le résolution verticale utilisée était de 56 niveaux verticaux où l'espacement entre les points de grilles variait de 50 à 350 m dans les premiers 2 km et d'environ

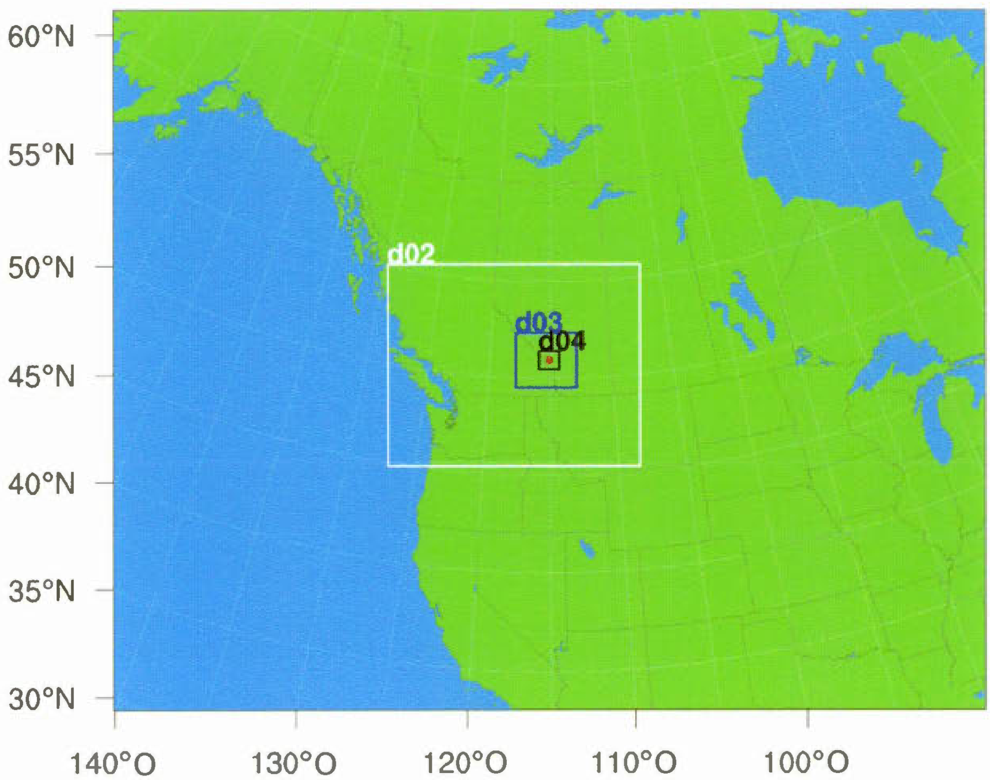

Figure A.1 Configuration du domaine dans le modèle numérique WRF où d01 est le domaine à résolution horizontale de 27 km (188 x 150 points de grille), d02 le domaine à 9 km (163 x 130 points de grille), d03 le domaine à 3 km (118 x 106 points de grille) et d04 le domaine à 1 km (118 x 106 points de grille). Le point rouge indique l'emplacement du site d'observation de KES.

350 m pour les plus hauts niveaux. Les simulations sur les grilles à plus basse résolution (27, 9 et 3 km) ont débuté à 1500 UTC 31 mars 2015, 3 heures plus tôt que pour la grille à haute résolution (1 km) qui a débuté à 1800 UTC 31 mars 2015. Les simulations ont été calculées pour un total de respectivement 12 h et 9 h. Le pas de temps utilisé était de 90 s sur le domaine à basse résolution (27 km) et se réduisait avec un rapport de 3 entre chaque grille. Le pas de temps de la grille à haute résolution (1 km) était de 3.33 s.

APPENDICE B

SCHÉMA MICROPHYSIQUE MILBRANDT ET YAU

Les simulations effectuées dans cette étude utilisaient la paramétrisation physique complète disponible dans le modèle numérique WRF : microphysique, radiation, schéma de surface, couche limite planétaire et schéma de convection (domaine à 27 km seulement). Plus particulièrement, les nuages et la précipitation sont représentés par le schéma microphysique *bulk* de Milbrandt et Yau (2005). Ce schéma utilise la distribution gamma afin de déterminer la distribution de taille de chaque type de précipitation qui est représentée par l'équation :

$$N(D) = N_0 D^{\alpha} e^{-\lambda D} \tag{B.1}$$

où $N(D)$ est la concentration en nombre de particules de diamètre D par unité de volume, N_0 est le point d'intersection, α est le paramètre de forme de la distribution et λ est le paramètre de pente.

Le schéma est à deux moments, ce qui signifie que deux paramètres sont prédits et résolus dans la distribution gamma. Le contenu en masse et la concentration en nombre des différentes catégories d'hydrométéores sont prédits et résolus par N_0 et λ, tandis que le paramètre α est lui égal à zéro dans la version du schéma utilisée. Le schéma de Milbrandt et Yau permet de prédire la concentration en nombre ainsi que le rapport de mélange pour six catégories d'hydrométéores : les gouttelettes de nuage, la pluie, la glace, la neige, la neige roulée et la grêle.

APPENDICE C

DESCRIPTION DÉTAILLÉE DU CAS DU 28 MARS 2015

Le 28 mars 2015, un système de basse pression se situait au nord-est de la vallée de Kananaskis amenant des vents généralement de l'ouest dans la région (Fig. C.1). Les données provenant de la station météorologique d'Environnement et Changement climatique Canada située près du Lac Barrier dans la vallée de Kananaskis indiquaient une accumulation de pluie de 9 mm pour la journée du 28 mars 2015.

Un radiosondage lancé à 1900 UTC 28 mars 2015 au-dessus du site de KES indiquait une couche sous-saturée près de la surface et des conditions saturées en altitude (Fig. C.2). Les données des instruments situés au site de KES indiquaient une diminution de température de 9 °C à 5 °C (Fig. C.3 a) de 1000 à 1400 UTC et une augmentation de l'humidité relative de 50 % à près de 95 % (Fig. C.3 b) durant la même période. Les vents ont variés de 3 à 9 m/s et provenaient principalement du sud-sud-ouest (Figs. C.3 c et d). Les données provenant du MRR ont démontré la présence d'une ligne pluie-neige vers 1130 UTC, puis vers 1300 UTC (Fig. C.4).

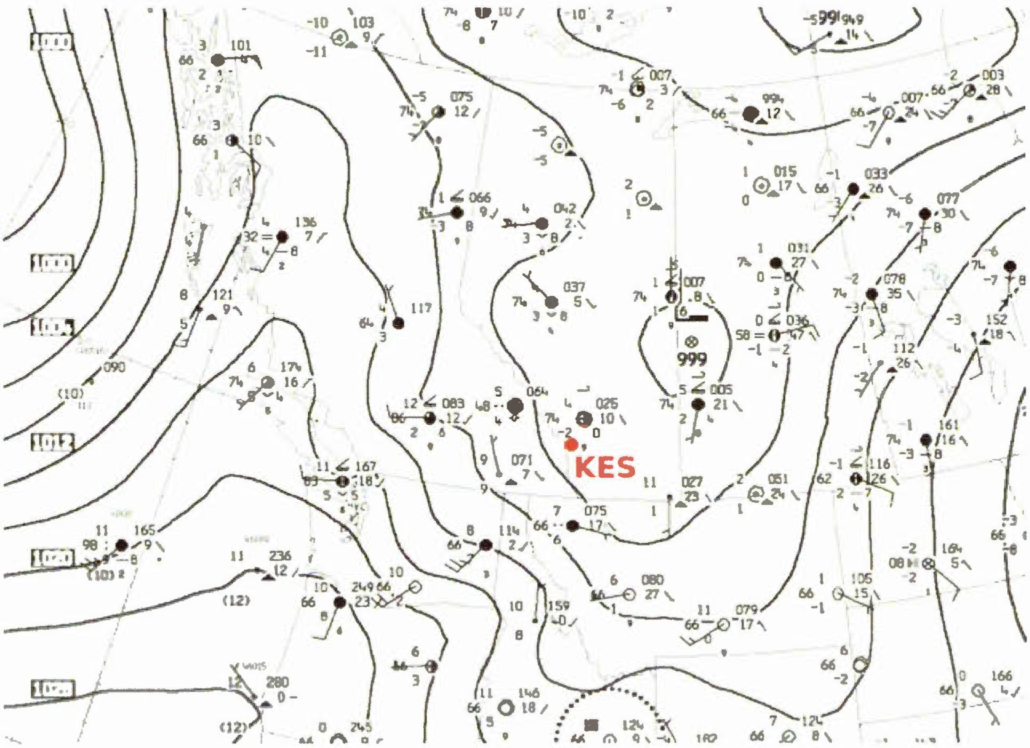

Figure C.1 Analyse de surface à 1200 UTC 28 mars 2015. source : Environnement et Changement climatique Canada.

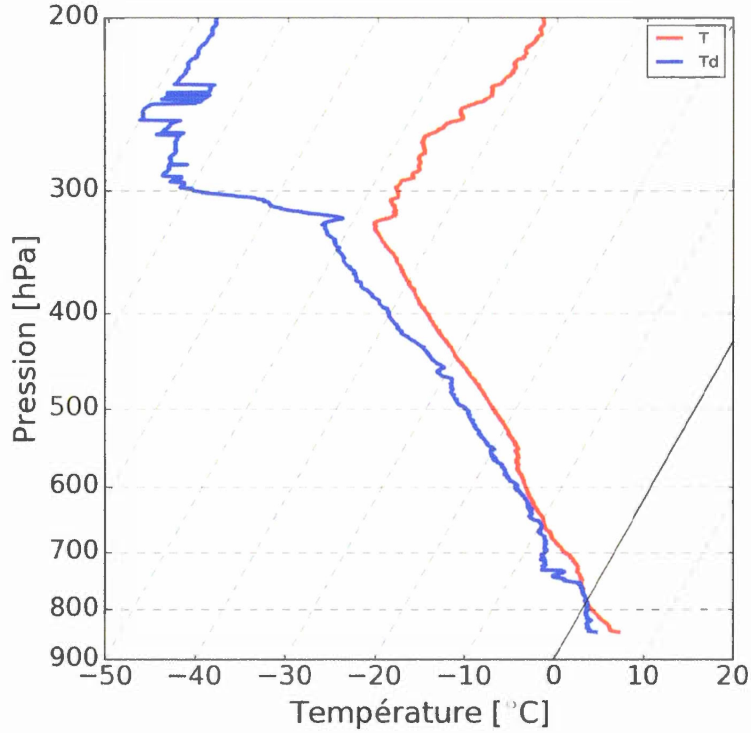

Figure C.2 Profil vertical de la température (ligne bleue) et de la température du point de rosée (ligne rouge) provenant d'un radiosondage lancé au-dessus de KES à 1900 UTC 28 mars 2015.

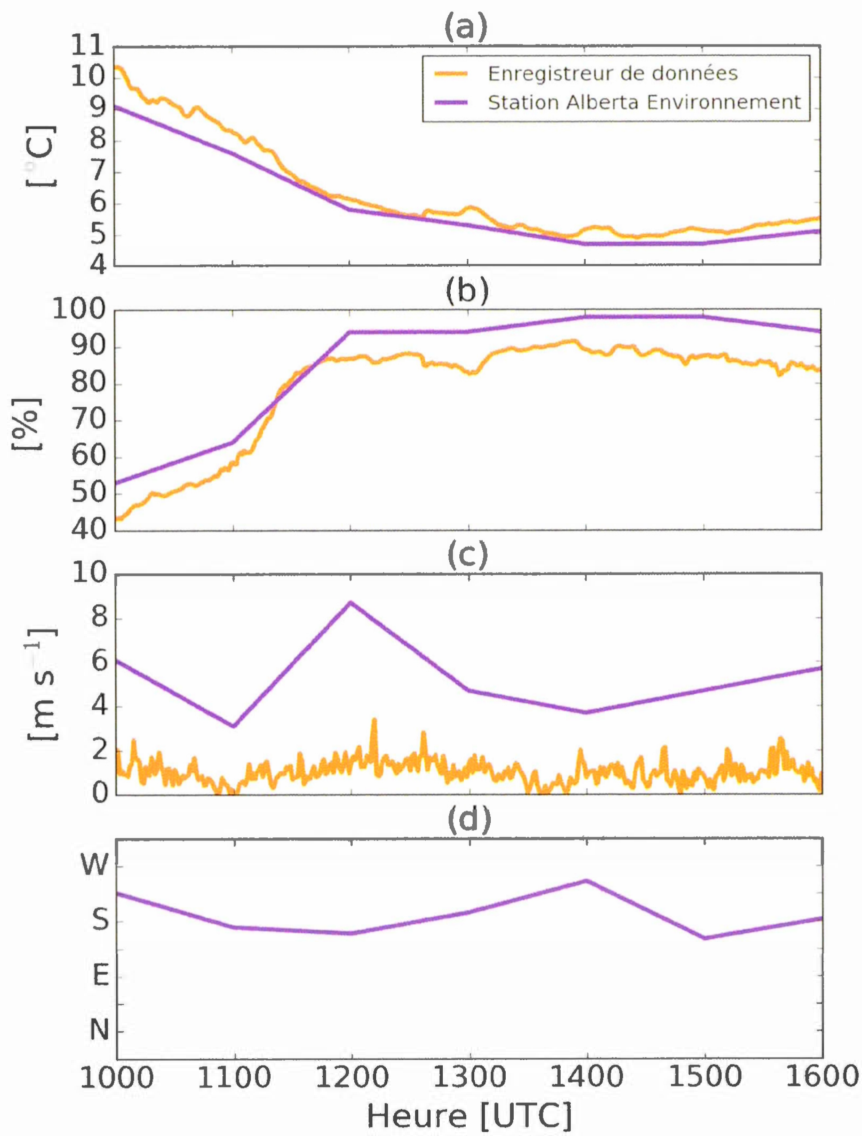

Figure C.3 (a) Température à la surface, (b) humidité relative, (c) vitesse du vent et (d) direction du vent à KES le 28 mars 2015 provenant de l'enregistreur de données (lignes oranges) et de la station Alberta Environnement (lignes mauves).

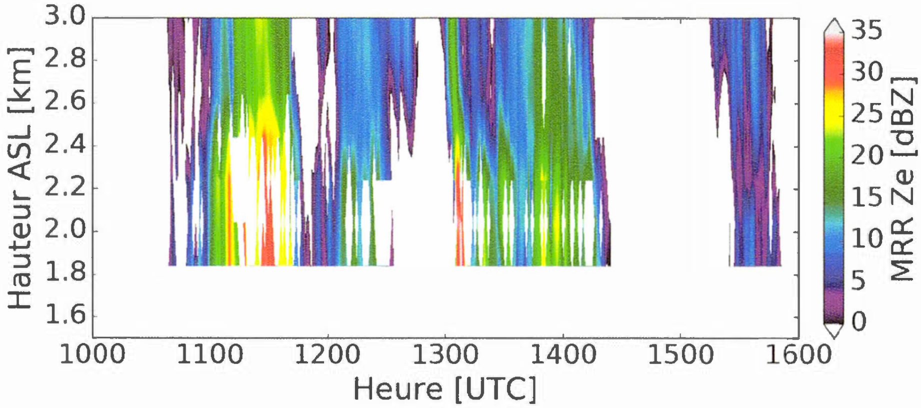

Figure C.4 Réflectivité mesurée à partir du *Micro Rain Radar* situé au site de KES le 28 mars 2015.

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