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Observations of Anti-winds in a Deep Alpine Valley, Lake Tekapo, New Zealand

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Abstract

This paper presents observations of summertime anti-winds monitored under ideal conditions in the Lake Tekapo hydro-catchment situated in the central Southern Alps, New Zealand. Onset and cessation of anti-winds was observed to coincide with the change in phase of the surface limbs of thermally generated valley and mountain winds under settled anti-cyclonic conditions. Anti-winds were best developed in the early morning before surface heating and associated convective mixing of the valley atmosphere began to mask the boundaries between the surface based limb of the mountain-valley wind and the corresponding anti-wind. By mid-day, the anti-valley wind exceeded the height of the surrounding ridgeline and became embedded in the topographically channeled gradient wind. Observations presented here have both theoretical and applied implications with regard to the development of thermally generated wind systems in deep alpine valleys, and their role in the dispersion of air pollution.

Introduction

Thermally generated winds such as slope, valley-mountain, and plain-mountain circulations have been the focus of numerous studies for more than 100 yr (see reviews by Atkinson, 1980; Sturman, 1987; Whiteman, 1990, 2001). Typically, such studies have been in response to knowledge gaps pertaining to boundary layer processes that are likely to influence the dispersion of air pollution during settled weather conditions. These studies have also had broader application to agricultural operations, including aerial spraying and frost protection, and more recently to wild fire management. As a result, much has been learned about the formation and evolution of thermally generated wind systems in simple valley-basin or valley-plain settings.

However, the conceptual models (or variants thereof) of Urfer-Henneberger (1970) and Whiteman (1982), which are commonly used to schematically illustrate thermally generated airflow in valleys, arguably oversimplify real-world conditions. These models predict daytime up-slope and up-valley airflow, which is replaced by nocturnal down-slope and down-valley airflow, while Whiteman's model also includes an overlying synoptic wind. Neither model shows return flows or anti-winds, even though Ekhardt (1934) first described them almost 70 yr ago. Whiteman's more recent model of the thermally forced winds of the Appalachian Mountains and the Tennessee Valley does include closed circulations with return flows (anti-winds) shown (Whiteman, 2000), thereby acknowledging the existence of these compensatory limbs of thermotopographic circulations. These return flows blow in the opposite direction to the surface limb of thermally generated winds. They develop in response to the law of conservation of mass in which air blowing up or down valley must theoretically be replaced by a return flow or anti-wind above. However, as anti-winds are typically much weaker than the surface limb of thermally generated circulations and are rarely confined by topography, they often become masked by the prevailing synoptic wind. Not surprisingly, observations of anti-winds are reported infrequently (Whiteman, 2000).

Previous Observations of Anti-winds

The study of Buettner and Thyer (1966) of anti-winds in the Carbon River valley, which descends from Mount Rainier, Washington

State, U.S.A., is commonly referred to in the literature when anti-winds are reported. In this study, anti-winds were observed to blow approximately in the opposite direction to the valley and mountain winds, but were found also to deviate considerably due to the influence of the gradient wind, mixing with other anti-winds, or when they took "short-cuts" across valley bends (Buettner and Thyer, 1966). They were found to be of similar thickness as the surface limb of the valley and mountain winds and were occasionally depressed below ridge level, typically at night. However, no mass or volume flux calculations of the valley/mountain winds and their associated anti-winds were performed by Buettner and Thyer (1966).

More recently, Clements (1999) reported the occurrence of anti-winds in the Lee Vining Canyon of the Sierra Nevada, California. Anti-winds at this site were found near or below ridge height, depending on the vertical extent of the drainage flow in the canyon, but Clements cautioned against the significance of the observations because of the limited observations that were made. Banta et al. (1999) observed anti-winds in the upper half of the Grand Canyon, Arizona, using a scanning Doppler lidar positioned on the canyon rim. Anti-winds were found to display a diurnal down-valley (daytime) and up-valley (nocturnal) rhythm, which had an opposite phase relationship with flow in the lower portion of the canyon. However, they also suggested that because of the complex topography, the anti-winds could be influenced by factors such as airflow from tributary valleys and winds from the Kaibab and Marble Plateaus.

Results from the MERKUR experiment conducted in the large Inn Valley, Austria, also identified the existence of anti-winds above the surface limbs of both mountain and valley winds. At night the anti-mountain wind (*antibergwind*) prevailed above the ridges and was thought to contribute to subsidence over the foreland, completing a closed mountain wind circulation. The anti-wind was not considered to be channeled by the main valley, but instead was thought to be distributed over the entire catchment. As a result, it could scarcely be detected from the prevailing gradient wind (Freytag, 1987). Reiter et al. (1984) reported on the occurrence of anti-winds in the Inn Valley during daytime almost 50 yr after the observations of Burger and Ekhardt (1937). The anti-valley wind (*antitalwinde*) was identified as a layer of down-valley airflow between 250 and 500 m deep above the prevailing valley wind (*Talwind*), with velocities of 1 to 1.5 m s⁻¹. It was not

confined by topography and most likely extended over the entire valley system beneath the prevailing gradient wind.

Whiteman and Bian (1998) presented observations of daytime (plain-mountain) and night-time (mountain-plain) flows over the Rocky Mountains and Great Plains using 5 yr of data from radar wind profilers. Their analysis identified the existence of a closed diurnally reversing regional mountain-plain circulation with weak return flows. Both the daytime anti-plain and nocturnal anti-mountain wind were found to extend through a much deeper layer of the atmosphere than the surface limbs of these circulations—i.e., up to 6 km asl., well in excess of the mean height of the Rocky Mountains (2.5 km).

Because anti-winds were not the principal focus of these studies, few details were presented of their structure, onset, and cessation, and interaction with other winds. In this paper, detailed observations of anti-winds made during the Lake Tekapo Experiment (LTEX), conducted in the central Southern Alps, New Zealand (Sturman et al., 2003), are presented by way of a case-study approach. Anti-winds in this region were first documented by McGowan et al. (1995) and McGowan and Sturman (1996), when pilot balloon flights made during favorable anti-cyclonic conditions often displayed return flow characteristics above surface-based limbs of valley and mountain winds. Results presented in this paper advance that earlier research and have application to air quality management, agricultural operations, wild fire management, and forecasting in complex terrain settings.

Physical Setting

Research for this paper was conducted in the Godley River valley in the headwaters of the Lake Tekapo hydro-catchment approximately 180 km southwest of Christchurch, New Zealand (Fig. 1). The Godley River valley is oriented almost north-south, extending approximately 30 km from the main divide of the Southern Alps south to the northern shoreline of Lake Tekapo. It is bounded to the east and west by the Sibbald and Hall mountain ranges which both contain many peaks exceeding 2300 m. This large U-shaped valley is approximately 3 km wide for most of its length before widening considerably at the confluence of the Macauley River valley 5 km north of Lake Tekapo (Fig. 1). Airflow within the valley is therefore primarily controlled by topography and channeled either up (southerly wind) or down valley (northerly wind).

Vegetation cover in the Godley River valley consists of tussock grassland on the steep valley walls, with some scrub. In the upper reaches of the valley and above the summer snowline (2000 m), exposed rock, snow, and ice surfaces dominate with approximately 52 km² of glaciers. Vegetation cover on the floor of the Godley River valley is sparse, being mostly tussock grasses, which are confined to relict river channels and river terraces. The remaining valley floor consists typically of braided river channels and extensive areas of exposed alluvium.

The climate of the Godley River valley is typical of the eastern alpine valleys of the Southern Alps where, the absence of a direct maritime moderating influence on the weather, elevation and rain-shadow effects dominate. Annual rainfall averages approximately 5 to 6 m in the headwaters of the valley, falling to only 1200 mm yr⁻¹ near the northern shoreline of Lake Tekapo. Snow can blanket the entire valley for 6 to 8 weeks during winter, when mean daily air temperatures may remain below -10°C. Under settled anti-cyclonic conditions this results in persistent cold-air drainage flowing from the catchment. Thermotopographic circulations dominate the southern half of the valley, while foehn northerlies are more common in its headwaters (McGowan et al., 2002). As anti-cyclones move east of New Zealand, the surrounding mountain ranges cause the strengthening upper-level gradient airflow to enter the valley due primarily to forced channeling (Whiteman and Doran, 1993). At these times, wind speeds in the valley

may regularly exceed 50 m s⁻¹, particularly during prefrontal conditions when the topographically modified gradient flow displays classic foehn characteristics (McGowan et al., 2002).

Observational Network

Observations presented in this paper were made during the austral summer field measurement campaign of the LTEX study in February 1999. Near-surface conditions were monitored continuously by a network of automatic weather stations, mobile surface energy balance stations, and two Doppler sodar. In addition, three pilot ballooning stations and an instrumented Piper Archer aircraft provided information on the thermodynamic structure of the valley atmosphere up to heights of 2000 m agl. Data on the regional wind and pressure field was made available from the New Zealand Meteorological Service including rawinsonde soundings from Hokitika (Fig. 1). In this study, observations made by pilot ballooning stations located in the Godley Valley and on the eastern lakeshore are presented, along with supporting aircraft and rawinsonde data. A more detailed review of the LTEX instrumentation set-up is provided in Sturman et al. (2003).

Synoptic Setting and Regional Windfield: 12 February 1999

The mid-day mean sea level synoptic analyses for 11 and 12 February 1999 are presented in Figure 2. They show an anti-cyclone centered east of Christchurch on 11 February, which over the following 24 h became “squeezed” between a southward-tracking subtropical depression and an approaching frontal system from the Southern Ocean. This resulted in a weak synoptic pressure gradient over central New Zealand throughout the lowest 5500 m of the atmosphere as indicated by the 500 hPa analysis (Fig. 2c). As a result, winds throughout this layer were light and veered from the southeast to northeast at heights above 2500 m, as indicated by rawinsonde flights conducted at Hokitika (Fig. 3). Below this level, local thermally generated circulations such as sea-land breezes and valley-mountain and plain-mountain winds dominated as discussed by Kossman et al. (2002) and Zawar-Reza et al. (2004).

Structure of the Valley Windfield

Time-height sections of *u* (westerly) and *v* (southerly) wind components computed from pilot balloon flights conducted at P1 in the Godley River valley and P2 on the eastern lake shoreline (Fig. 1) are presented in Figure 4. Both sites recorded a down-valley northerly component (-*v*) until approximately 0900 NZST at P2 and 1000 NZST at P1. The height of this cold air drainage (mountain wind) varied between 400 and 800 m agl. at P1, which is 1100 to 700 m below the surrounding ridgeline (Fig. 4a). At P2, the airflow was approximately 900 m in height (Fig. 4c), but still below the surrounding ridges. The greater thickness of flow at this site was the likely result of cold air entrainment from the surrounding slopes and the Macauley River valley located to the north of P2 (Fig. 1). Above this airflow both stations recorded a weak southerly up-valley anti-mountain wind, which at P2 was dominated by a southerly jet between 1300 and 1650 m agl. This is above the height of the Two Thumb Range located east of this site (Fig. 1), while in the Godley River valley this airflow was most dominant below the surrounding ridgeline.

By 1000 NZST both sites began to record the onset of up-valley southerly airflow at the surface as indicated by the +*v* component. This was the onset of the daytime valley-wind system (Figs. 4a, 4c), which was approximately 800 m in depth at P2 and 500 m at P1. This reversal in the dominant component flow at the surface from a down-valley

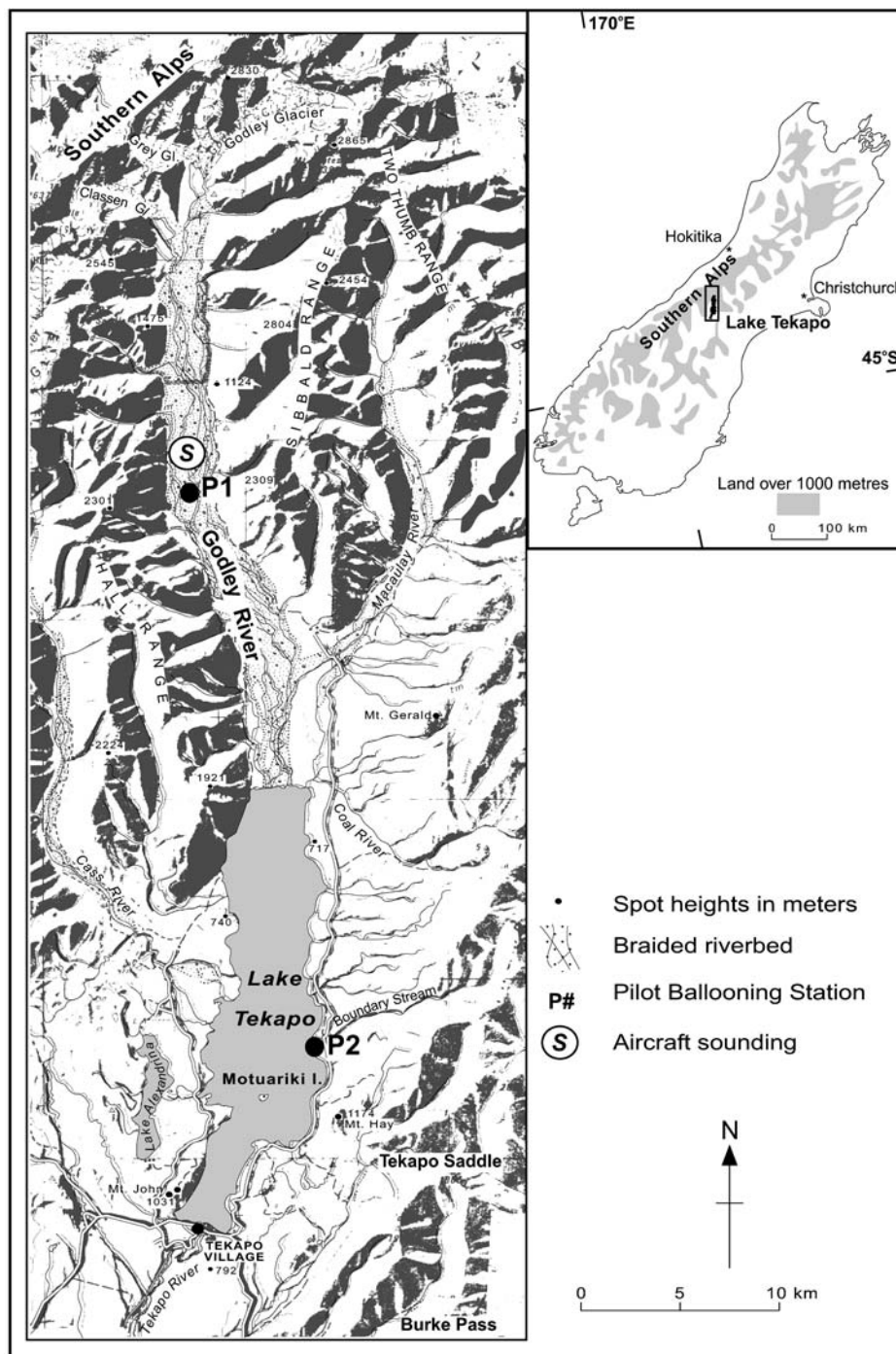


Figure 1. Map of the Lake Tekapo study area and the location of measurement sites.

northerly mountain wind to an up-valley southerly valley wind was associated with a change in phase from southerly to northerly component flow aloft, indicating a reversal in the compensatory anti-winds. However, as shown in Figure 4, the daytime northerly component flow extended well above ridge height, indicating that the weak synoptic airflow was coupling with the valley atmosphere. Above ridge height, the gradient wind veered from the northeast to the southeast throughout the day with velocities of $<5 \text{ m s}^{-1}$.

The intrusion of the east coast meso-scale plains to mountain wind into the Lake Tekapo Basin occurred between 1730 and 1800 NZST. This resulted in an abrupt increase in the depth of the up-valley southerly to heights greater than 2000 m agl. as the local wind system became embedded in this much larger scale wind (Fig. 4). It seems

reasonable to suggest that a plains to mountain anti-wind may have subsequently developed as discussed by Whiteman and Bian (1998), but this could not be identified as it would have exceeded the height of pilot balloon flights ($>2000 \text{ m agl.}$).

Analysis of the u-component (Figs. 4b, 4d) (westerly wind) identified a light easterly airflow ($-u$) above ridge top at both sites, with an easterly jet recorded at P2 with velocities $>2 \text{ m s}^{-1}$ between 1500 and 1600 m agl. from 0730 to 1000 NZST. At P1 this easterly component flow remained light all day, lowering to about 1100 m agl. at 1500 NZST before cessation at approximately 1900 NZST. This is consistent with rawinsonde data from Hokitika presented in Figure 3. Below the easterly component flow, cross-valley airflow at P1 in the Godley River valley was negligible until early afternoon when

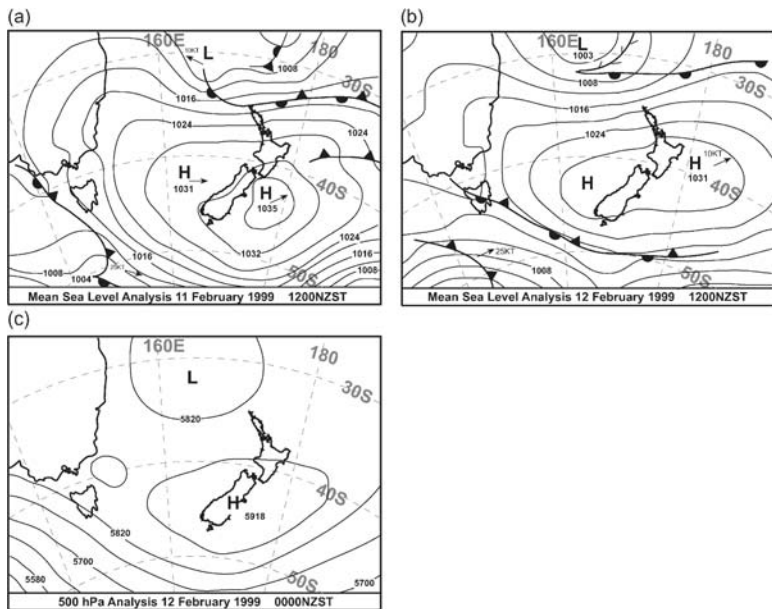


Figure 2. (a, b) Mean sea level analyses for 1200 NZST on 11 and 12 February 1999; (c) the 500 hPa analysis for 0000 NZST on 12 February 1999.

a westerly component became prominent. This developed in response to the dry and step west-facing slopes of the valley receiving maximum insolation at this time of day, while the east-facing slopes became increasingly shaded as the sun began to set (Fig. 4b). Down-valley at P2, the onshore westerly grew in depth throughout the day as (1) local anabatic slope winds and (2) the lake breeze became established at this site. A brief period of easterly airflow occurred at 1600 NZST and was associated with the onset of the plains to mountain wind that was channeled through Tekapo Saddle 6 km southeast of P2 (Fig. 1), before westerly airflow became dominant again (Fig. 4d).

Vertical profiles of wind speed and direction determined from pilot balloon flights conducted at P1 in the Godley River valley and P2 on the eastern lake shore are presented in Figure 5. Onset of the northerly mountain wind was recorded by an automatic weather station located at P1 at approximately 2000 NZST on 11 February 1999. By 0200 NZST on 12 February, the mountain wind was approximately 550 m in height with a southerly anti-mountain wind prevailing above, which extended up to at least ridge height (Fig. 5a). Above the

ridgeline, a light southeasterly gradient wind prevailed that veered to the northeast over the following 11 h (Fig. 5a), which reflects rawinsonde observations presented in Figure 3 from Hokitika.

At 0700 NZST the mountain wind reached a maximum height of approximately 745 m agl. at P1 and 900 m agl. at P2 (Fig. 5). It was also clearly identifiable in the vertical mixing ratio profile obtained by a light aircraft that completed a sounding in the valley atmosphere near P1 at 0700 NZST (Fig. 6a). Above this height, the anti-mountain wind was characterized by a 655 m layer of southerly airflow below a much drier southeasterly gradient wind with a mixing ratio of only 1.5 g kg^{-1} . At P2 the anti-mountain wind extended well above the ridgeline (Fig. 5b), similar to that reported by Reiter et al. (1984) for the large Inn Valley, Austria. Under such conditions the upper limits of the anti-wind could not be defined, and mass flux estimates were not possible.

Cessation of the mountain wind was recorded at P1 at 1035 NZST with an abrupt front-like change to the southerly valley wind (Fig. 5a). This wind change was associated with an equally rapid reversal in the anti-wind to a down-valley northerly anti-valley wind as displayed in Figure 5a. Over the following 8 h the anti-valley wind varied considerably in height, mostly in response to increased mixing of the valley atmosphere as the higher west-facing slopes of the valley began to receive maximum insolation from mid-afternoon onwards. The light gradient wind also appeared to erode the anti-valley wind from above, most notably in the mid-afternoon when it descended into the valley atmosphere as seen in Figures 5a and 6b. The mixing ratio profile presented in Figure 6b indicates that the valley wind was not that much drier than the mountain wind, which it replaced especially near the surface, highlighting the dry nature of the valley atmosphere. However, above 800 m the valley atmosphere became more humid between 0700 NZST and 1500 NZST. This is likely to be the result of advective processes transporting moisture from the valley side walls toward the valley center.

Onset of the daytime wind field was recorded at 0900 NZST at P2 as the wind backed to the southwest. This was associated with shallow anabatic slope winds which became masked by the larger scale valley wind at 1000 NZST (Fig. 5b). The valley wind reached a maximum height of approximately 950 m agl. and veered throughout the day to the southwest in response to the local lake breeze effect (McGowan et al., 1995). Above the valley wind, no clear distinction could be made between the anti-valley wind and the gradient flow at this site. These local winds were replaced after 1530 NZST by the regional-scale plain

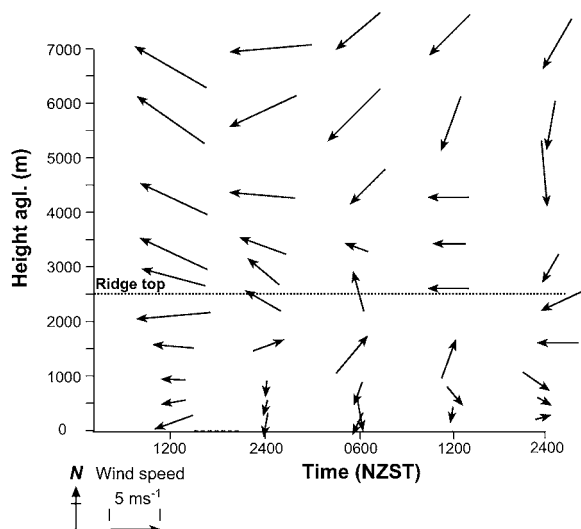


Figure 3. Vertical profiles of wind speed and direction obtained from rawinsonde soundings at Hokitika, 1200 NZST on 11 February 1999 to 2400 NZST on 12 February 1999.

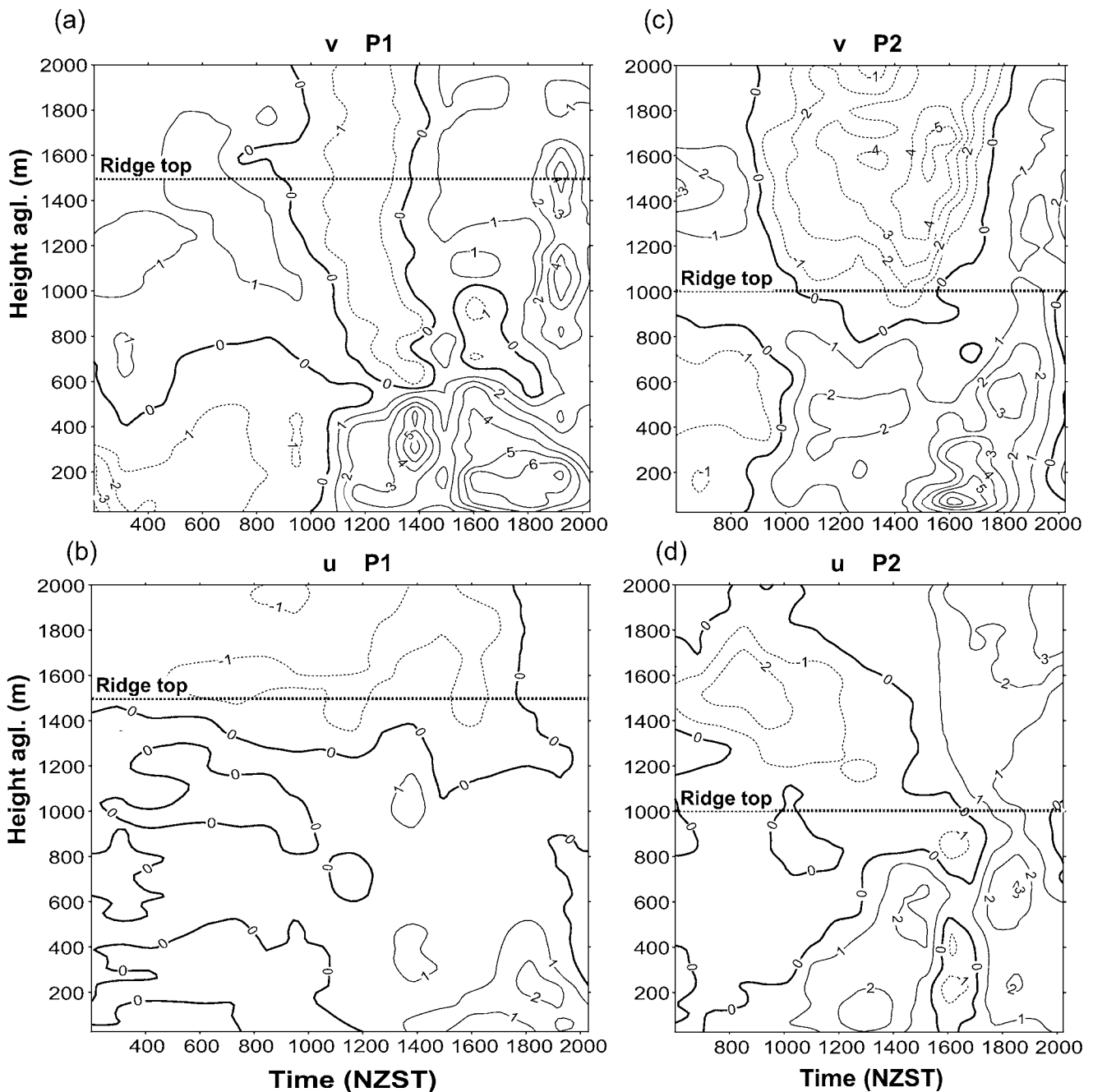


Figure 4. Time height sections of the u -component and v -component of winds measured on 12 February 1999 by pilot balloons at P1 (a, b) in the Godley River valley and P2 (c, d) on the eastern shore of Lake Tekapo.

to mountain wind, which enters the lake basin through mountain passes to the east of this site as discussed by Kossmann et al. (2002) and Zawar-Resa et al. (2004). This large scale flow reached P1 at 1830 NZST and became dominant throughout the valley atmosphere, reaching a height exceeding the surrounding mountain ranges (>2300 m agl.).

Summary

The inability to accurately define the upper boundary of anti-winds is at present a significant obstacle to their research. This has led many scientists to question their existence, even though wind-profile data from pilot balloon flights, as presented in this paper, and acoustic sodar profiles regularly identify the existence of return flows. The frequent use

of the term *circulation* by scientists when referring to thermally generated winds implies the existence of anti-winds and the concept of continuity of mass through some form of return flows. But the inability to accurately define the upper boundaries of anti-winds when they regularly exceed the height of ridge-lines and merge with larger scale winds as observed by this study is a major research challenge.

Observations from the Godley River valley and shore of Lake Tekapo made during ideal anti-cyclonic conditions, clearly identified both anti-valley and anti-mountain winds. Analysis of along-valley (north-south) component flow and pilot balloon profiles revealed that the reversal in the surface limb of the valley-mountain wind system was associated with a change in phase in the corresponding anti-wind. These compensatory winds did not appear to exceed the height of surrounding mountain ridges in the Godley River valley, but were complex in

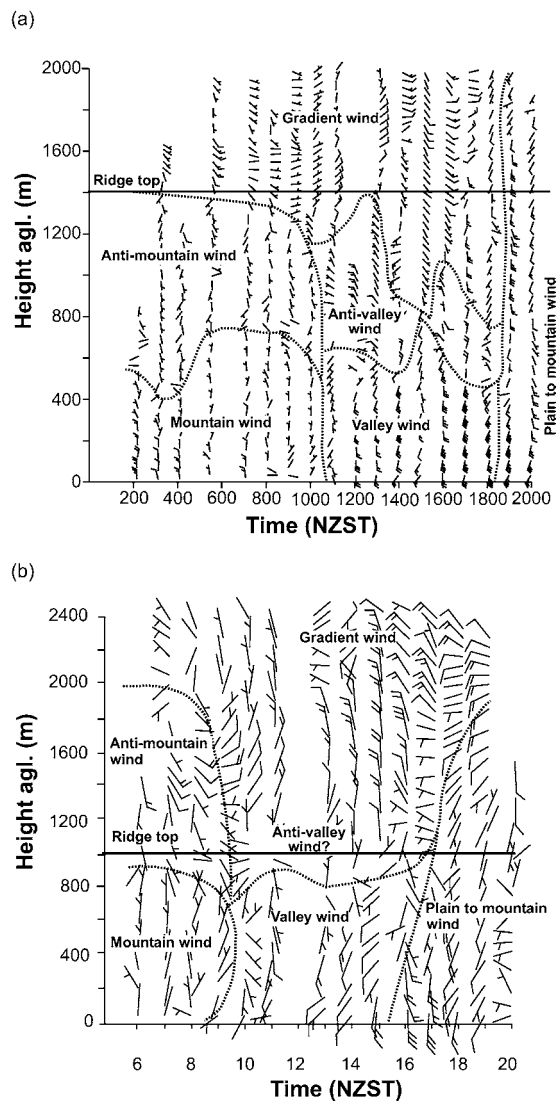


Figure 5. Vertical profiles of wind speed and direction obtained at pilot balloon site (a) P1 in the Godley River valley and site (b) P2 on the eastern shore of Lake Tekapo, 12 February 1999. Short bars = 1.25 m s^{-1} and long bars = 2.5 m s^{-1} .

structure and varied greatly in height as reported by Buettner and Thyer (1966). However, where the valley is much wider at site P2, the anti-winds were more difficult to distinguish from the prevailing gradient wind. At this site the anti-mountain wind was believed to extend well above the surrounding mountain ranges, which is similar to observations made in the Inn Valley, Austria, by Reiter et al. (1984). Furthermore, coupling with the prevailing gradient wind during the afternoon when the anti-valley wind displayed considerable variability in height at P1 can not be discounted due to intense heating and associated mixing of the valley atmosphere as discussed by Kossmann et al. (2002). This may be the reason why on this occasion no clear boundary could be identified between the anti-valley wind and the prevailing gradient flow at P2.

While the value of accurate measurement of anti-winds is unquestionable, confirmation of their spatial domain is a major research challenge. For example, attempts made by this study to achieved closure of mass-balance calculations for the valley and mountain wind systems (not presented) were unsuccessful as the precise boundary between the anti-winds and the prevailing gradient wind could not be determined. Similar results have led many scientists to question whether anti-winds exist, in spite of the mounting observational evidence of

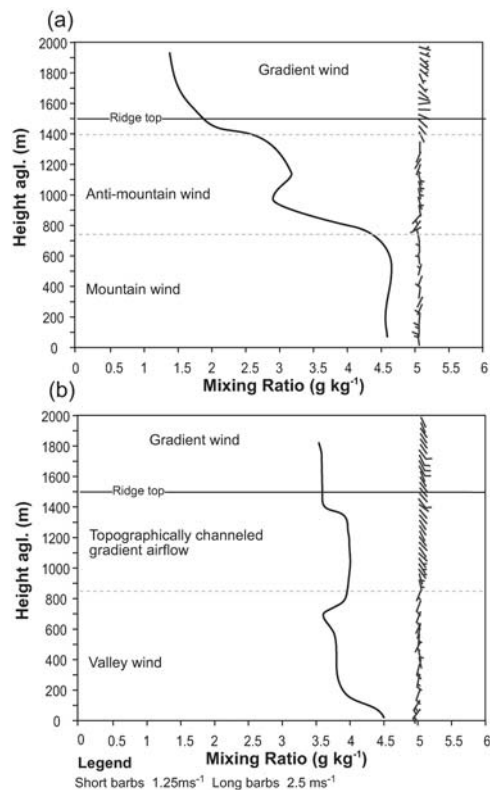


Figure 6. Vertical profiles of wind speed and direction obtained by pilot balloons, and mixing ratio profiles obtained by an instrumented light aircraft at site P1 in the Godley River valley for (a) 0700 NZST and (b) 1500 NZST, 12 February 1999.

their occurrence as presented by this study. The deployment of horizontally and vertically scanning lidar on ridges of simple valley systems, as attempted by Banta et al. (1999), may help to accurately identify the boundary between winds. The use of this technology would allow the surface and anti-wind limbs of thermally generated circulations to be clearly defined within simple valleys. However, such observations would require detailed monitoring of other thermally generated circulations to fully resolve the structure of anti-winds and their role in local and regional scale mass, moisture, and heat budgets. For example, the identification of an anti-plain to mountain wind in the Lake Tekapo area would require vertical profiles of wind speed and direction possibly exceeding 4000 m a.g.l., based on the observations of Whiteman and Bian (1998), and those made by pilot balloon soundings in the present study. Such observations should be conducted over an extended period of time to fully resolve the physical nature and incidence of anti-winds under a range of synoptic conditions. The benefit of such research is clear with regard to its application to air quality management and air pollution dispersion modeling in complex terrain settings, which are increasingly under pressure from development.

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