Large eddy simulation of atypical wildland fire spread on leeward slopes

Colin C. Simpson, Jason J. Sharples, Jason P. Evans and Matthew F. McCabe

School of Physical, Environmental and Mathematical Sciences, University of New South Wales at Canberra, Canberra, ACT 2600, Australia.
B Climate Change Research Centre, Faculty of Science, University of New South Wales, Sydney, NSW 2052, Australia.
CSchool of Civil and Environmental Engineering, University of New South Wales, Sydney, NSW 2052, Australia.
D Corresponding author. Email: colin.c.simpson@gmail.com

Abstract. The WRF-Fire coupled atmosphere–fire modelling system was used to investigate atypical wildland fire spread on steep leeward slopes through a series of idealised numerical simulations. The simulations are used to investigate both the leeward flow characteristics, such as flow separation, and the fire spread from an ignition region at the base of the leeward slope. The fire spread was considered under varying fuel type and with atmosphere-fire coupling both enabled and disabled. When atmosphere–fire coupling is enabled and there is a high fuel mass density, the fire spread closely resembles that expected during fire channelling. Specifically, the fire spread is initially dominated by upslope spread to the mountain ridge line at an average rate of 2.0 km h\(^{-1}\), followed by predominantly lateral spread close to the ridge line at a maximum rate of 3.6 km h\(^{-1}\). The intermittent rapid lateral spread occurs when updraft–downdraft interfaces, which are associated with strongly circulating horizontal winds at the mid-flame height, move across the fire perimeter close to the ridge line. The updraft–downdraft interfaces are formed due to an interaction between the strong pyro-convection and the terrain-modified winds. Through these results, a new physical explanation of fire channelling is proposed.

Received 10 May 2012, accepted 12 December 2012, published online 25 March 2013

Introduction

A wildland fire is capable of exhibiting highly complex behaviour in response to multi-scale interactions between the fire and the local fire environment, namely the fuel, weather and topography. It has previously been identified that terrain-modified atmospheric conditions, particularly in complex terrain, can significantly affect fire spread and behaviour (Sharples 2009; Sharples et al. 2010a). This paper investigates a fire spread phenomenon referred to as ‘fire channelling’, in which atmosphere–terrain–fire interactions are believed to play an important role. McRae (2004) first noted this phenomenon, which he referred to as ‘lee-slope channelling’, through the presence of atypical fire spread patterns in multispectral line-scan data from the Canberra 2003 bushfires. A study by Sharples et al. (2010b) investigated fire channelling at the laboratory scale, through a series of combustion tunnel experiments, and confirmed the incidence of atypical lateral fire spread across a leeward slope, apparently driven by an interaction between the wind, the terrain and the fire.

Previous studies have identified several important distinguishing features of fire channelling (Sharples and McRae 2011; Sharples et al. 2011, 2012). These features include rapid lateral fire spread across the top of a steep leeward slope in a direction approximately perpendicular to the synoptic wind conditions. The upwind edge of the lateral spread is constrained by a major break in topographic slope, such as a mountain ridge line. There is an extension of the active flaming zone downwind of the synoptic flow, possibly through a process such as spotting. Additional features include darker smoke and vigorous convection associated with the laterally advancing flanks of the fire. The rapid lateral fire spread indicates that fire channelling may pose a significant danger to fire fighter and civilian safety.

Sharples et al. (2012) have previously determined that several environmental conditions are necessary for fire channelling. The leeward slope angle of the mountain should be greater than ~25° and the topographic aspect and synoptic wind direction should be within ~40° of each other. The synoptic wind speed should also be greater than ~25–30 km h\(^{-1}\), which allows for flow separation in the lee of the mountain. This study postulated that fire channelling occurs due to an interaction between the fire and a lee rotor, which is formed through leeward flow separation. It was further conjectured that the lateral fire spread is driven by thermal expansion of the air within the lee rotor as heat is added to it from the fire, with the resulting lateral atmospheric flow effectively following a path of least resistance.
An important step in this study is to use a numerical weather prediction model to investigate the nature of the atmospheric flow in the lee of a mountain. Several studies have previously investigated turbulent flow for environmental conditions and atmospheric scales similar to that considered in this paper (Schar and Durran 1997; Allen and Brown 2002; Doyle and Durran 2002, 2007; Ding et al. 2003; Pathirana et al. 2003; Hertenstein and Kuettnern 2005; Sheridan and Vesper 2005; Ayotte 2008; Katurji et al. 2011). Many of these studies used a large eddy simulation model to investigate the nature of the atmospheric flow. Based on the results of these studies, it is evident that the nature of the leeward atmospheric flow is closely associated with many environmental conditions, including the atmospheric stability, surface roughness, upstream wind conditions and the geometric properties of the terrain.

Another important step in this study is to simulate the fire spread across the leeward slope of a mountain using a coupled atmosphere–fire model. Over the past two decades, several studies have used coupled atmosphere–fire modelling to investigate fire spread, fire behaviour and atmosphere–fire interactions. Previous studies focussing on fire spread across flat terrain have been able to reproduce several physically realistic fire spread features (Heilman and Fast 1992; Clark et al. 1996a, 1996b, 2004; Cunningham et al. 2005; Cunningham and Linn 2007; Mell et al. 2007; Sun et al. 2009). These modelled features include the parabolic fire shape that typically develops under the influence of light uniform winds. Previous studies focussing on fire spread across non-flat terrain have similarly yielded useful results (Heilman 1992; Linn et al. 2002, 2007; Clark et al. 2004; Coen 2005).

The aim of this paper is to perform a series of idealised numerical simulations that allow for an evaluation of the fire channelling hypothesis forwarded by Sharples et al. (2012) and facilitate a detailed analysis of the physical mechanisms responsible for driving the lateral fire spread associated with fire channelling. The numerical modelling system used to perform these numerical simulations is described in the next section. The results of the two- (2-D) and three-dimensional (3-D) atmospheric simulations of flow over a mountain are presented in the following two sections. The results of the simulations of fire spread across the leeward slope of a mountain are presented in the subsequent section. A summary of the study is presented in conjunction with several conclusions in the final section.

**Numerical models**

**Atmospheric model**

The atmospheric model used in this study is version 3.3 of the Weather Research and Forecasting model (WRF) (Skamarock et al. 2008). It is used in a highly idealised large eddy simulation (LES) configuration that is well suited to studying turbulent atmospheric flow on length scales of tens to hundreds of metres. The model explicitly resolves the large-scale atmospheric eddies, whereas the effects of subgrid-scale motions on the resolved turbulence are modelled using a subfilter-scale stress model. The model uses fully compressible nonhydrostatic equations with a mass-based terrain-following coordinate system. The model is used in both a 2-D and 3-D configuration.

The model domain is configured to capture the turbulent flow in the lee of a mountain at high resolution. In both the 2-D and 3-D simulations, the west–east dimension (x-axis) has an extent of 30 km. In the 2-D simulations, the south–north dimension (y-axis) is three grid points wide with cyclic boundary conditions, whereas in the 3-D simulations it has an extent of 5 km. The horizontal grid spacing in both the 2-D and 3-D simulations is 50 m. The model top in each simulation is initially set to 10 km, with an initial vertical grid spacing of 50 m. However, due to the use of terrain-following sigma coordinates, the model top, and therefore also the vertical grid spacing, varies in time with the atmospheric pressure. However, the model top descends no more than a few hundred metres in any simulation and the vertical grid spacing is therefore always in the range of 47–50 m. The model grid cells are therefore approximately isotropic throughout the duration of each simulation. The atmospheric model has a computational domain of 600 × 3 × 200 (x, y, z) and 600 × 100 × 200 (x, y, z) grid points in the respective 2-D and 3-D simulations. This model domain setup is similar to that considered in previous WRF LES studies (Mirocha et al. 2010; Kirkil et al. 2012).

The lateral boundary conditions are specified using a one-dimensional input sounding. The surface pressure is 1000 hPa and the surface moisture mixing ratio is zero. The vertical profiles of the vapour mixing ratio, horizontal wind conditions and potential temperature are also specified. The vapour mixing ratio and y-axis wind velocity are zero at all heights. The horizontal wind conditions at the western lateral boundary are set as a temporally uniform incoming westerly wind field, which varies with height according to:

\[
U(z) = \begin{cases} 
20 \times \left(\frac{z}{200}\right)^2 & z \leq 200 \\
20 & z > 200
\end{cases}
\]  

where \(U(z)\) is the wind speed (m s\(^{-1}\)) and \(z\) is the height (m). Moreover, this vertical wind profile is used to define the initial wind conditions over the entire domain.

A total of three different vertical profiles of the potential temperature, which represent a stable, neutral and unstable atmosphere, are used in this study. In the stable atmosphere, the potential temperature is initially equal to 280 K at the surface and increases linearly to 320 K at a height of 10 km. In the neutral atmosphere, the potential temperature is initially equal to 290 K at all heights, including the surface. In the unstable atmosphere, the potential temperature is initially 290 K at the surface and decreases linearly to 280 K at a height of 5 km, with a constant potential temperature of 280 K above a height of 5 km. The constant potential temperature lid on the unstable atmosphere acts to prevent the model top from descending to a very low height above the surface.

In each simulation there is a mountain with a triangular profile, with a height of \(\sim 1\) km, located within the model domain. The mountain starts 10 km to the east of the western lateral boundary and the windward slope has an inclination angle of 20\(^\circ\), giving a windward slope width of 2.75 km. The mountain ridge line is therefore located 12.75 km to the east of the western lateral boundary, which limits any physical
Numerical modelling of fire channelling

int. j. wildland fire

601

connection between the terrain-induced turbulence and the western lateral boundary. The leeward slope has an inclination angle of either 25 or 35°, giving a leeward slope width of either 2.14 or 1.43 km. The mountain is flattened slightly at the ridge line, with the highest two model grid points set to the same height. This ensures that there is no sudden change from positive to negative topographic gradient between adjacent model grid points.

The model is used in a highly idealised configuration, with many of the model physics schemes disabled, including the microphysics, longwave radiation, shortwave radiation, urban surface physics, planetary boundary layer and cumulus parameterisations. The MM5 surface layer scheme, which is based on Monin–Obukhov similarity, is enabled as surface friction is important for generating the near-surface vertical wind shear. Diffusion in physical space is calculated using the velocity stress tensor and eddy viscosities are calculated using a 3-D prognostic 1.5-order turbulence closure. A Rayleigh damping scheme (Klemp et al., 2008), with a damping timescale of 10 s, is used in the top 3 km of the model to absorb upward propagating gravity wave energy. All lateral boundary conditions other than the cyclic boundary conditions used in the y-axis for the 2-D simulations are open radiative. The main model time integration in WRF is performed using a third-order Runge–Kutta scheme and the primary time step chosen here is 0.1 s. The secondary time step is a time-split small step for acoustic and gravity wave modes, and is equal to one eighth of the primary time step. The initial 30 min of each simulation is considered the spin-up period and is not included in any calculation of the time-averaged variables discussed below.

Coupled atmosphere–fire model

The wildland fire spread model used in this study is SFIRE (Mandel et al. 2011), which is distributed with version 3.3 of WRF. The resulting coupled atmosphere–fire model is referred to as WRF-Fire.

The wildland fire spread is modelled as a temporally evolving fire perimeter that advances through the model domain using a level set function. The spatially and temporally variable fire perimeter that advances through the model domain to as WRF-Fire. The resulting coupled atmosphere–fire model is capable of directly modelling atmosphere–terrain–fire interactions. However, this two-way atmosphere–fire coupling can be switched off by not passing on the lateral and sensible heat fluxes from the SFIRE model to the WRF model.

Two-dimensional atmospheric simulations

In this section the results of six 2-D simulations of atmospheric flow over a triangular mountain are presented. Details of the setup used in each simulation are provided in Table 1. Each 2-D simulation has a unique name derived from three properties: dimensionality (‘2D’), atmospheric stability (‘S’ for stable, ‘N’ for neutral and ‘U’ for unstable) and leeward slope angle (‘25’ or ‘35’). These simulations are used to investigate the leeward atmospheric flow under different atmospheric stability and leeward slope angle conditions.
Fig. 1 shows the temporal evolution of the upstream vertical potential temperature profiles for the 2DS35, 2DN35 and 2DU35 simulations. The neutral profile shows no change with time, whereas the stable and unstable profiles become respectively slightly more stable and unstable with time. The atmospheric stability profiles are therefore relatively constant in time.

Stable atmosphere

The 2DS25 and 2DS35 simulations have a stable atmosphere, which allows for the development of mountain waves in the lee of the mountain. The Brunt–Väisälä frequency in these two simulations is \( \sim 0.01 \text{ s}^{-1} \), which corresponds to a predicted wavelength of \( \sim 10 \text{ km} \). This wavelength is significantly longer than the leeward slope width and implies that the mountain waves will likely have little direct effect on the flow conditions close to the leeward slope.

Fig. 2a shows the time-averaged potential temperature for the 2DS35 simulation. Mountain waves are seen to develop in the lee of the mountain, with amplitude of several hundred metres and a wavelength close to the 10 km predicted using the Brunt–Väisälä frequency. The mountain waves extend up to the bottom of the Rayleigh damping layer, which is located at a height of \( \sim 7 \text{ km} \). Although not shown specifically, the mountain wave characteristics in the 2DS25 simulation are very similar.

Fig. 3 shows the time-averaged horizontal wind velocity for the 2DS25 and 2DS35 simulations. It is evident that there is a high degree of similarity between the wind conditions in these two simulations. The westerly flow is seen to accelerate up the windward slope, reaching a peak velocity of \( \sim 20–24 \text{ m s}^{-1} \) at the mountain peak. There is a region of flow separation at the base of the windward slope, which is not unexpected given the stable atmosphere. Rapid deceleration of the westerly flow is seen as it moves down the leeward slope, which results in flow separation in the lee of the mountain. This region of flow separation starts at a height of \( \sim 600 \text{ m} \) on the leeward slope and extends eastwards out to a distance of \( \sim 2 \text{ km} \) downstream of the mountain. The acceleration and deceleration of the flow as it moves upslope and downslope is most likely due to the pressure gradient encountered by the flow as it lifts and descends across the mountain.

Also of interest in Fig. 3 is another region of flow separation located directly beneath the first mountain wave crest, which is \( \sim 6–12 \text{ km} \) downstream of the mountain. Doyle and Durran (2002) have previously shown that rotors can form under mountain wave crests, resulting in flow separation. The downstream location of this flow separation region confirms the above statement that the wavelength of the mountain waves is too long for the waves to directly influence the flow conditions across the leeward slope.

Fig. 4a shows the instantaneous horizontal vorticity and wind conditions at a time of 60 min for the 2DS35 simulation. The horizontal vorticity \( (\eta_h) \) is a useful quantity for diagnosing flow circulation and is calculated as:

\[
\eta_h = \frac{\partial U}{\partial z} - \frac{\partial W}{\partial x}
\]

where \( U \) and \( W \) are the respective horizontal and vertical wind velocities. Positive and negative vorticity indicate a respective clockwise and counterclockwise flow rotation.

Fig. 4a demonstrates that the flow separation at the top of the leeward slope is associated with periodic vorticity shedding in the lee of the mountain. The deceleration of the flow at the top of the leeward slope results in the development of a quasi-permanent region of strong positive vorticity at a height of 800–900 m, which extends out eastwards for 2 km from the leeward slope. Lee rotors are periodically generated beneath this quasi-permanent positive vorticity region. The rotors have an initial horizontal and vertical extent slightly less than the mountain height and are seen to detach from the mountain and propagate downstream with the westerly flow. The downstream movement of these rotors is seen to be closely associated with the mountain waves and the stable atmosphere acts to limit any increase in their vertical extent. This vorticity shedding process is responsible for quasi-periodic changes in the flow direction from the base of the leeward slope up to a height of \( \sim 700 \text{ m} \).
Neutral atmosphere

Fig. 5 shows the time-averaged horizontal wind velocity for the 2DN25 and 2DN35 simulations. As with the stable atmosphere simulations, there is acceleration of the westerly flow as it moves upslope, with a peak velocity at the mountain peak, and rapid deceleration of the flow at the top of the leeward slope, resulting in flow separation downstream of the mountain. There is a high degree of similarity between the two neutral simulations, however there are some important differences in the flow separation between the stable and neutral atmosphere simulations. First, no mountain waves develop in the neutral atmosphere and consequently there is no flow separation associated with mountain waves downstream of the mountain. Second, there is no flow separation at the base of the windward slope in the neutral atmosphere. Third, the leeward flow separation region in the neutral atmosphere seems to be lifted slightly from the surface across the leeward slope at heights under ~400 m. Fourth, the leeward flow separation region extends ~6 km downstream in the neutral atmosphere, which is ~4 km further than in the stable atmosphere.

Fig. 4b shows the horizontal vorticity and wind conditions at a time of 60 min for the 2DN35 simulation. The vorticity shedding process is qualitatively similar to that described above for the 2DS35 simulation. However, in the neutral atmosphere the lee rotors have a greater initial horizontal and vertical extent and they propagate downstream free of any influence from mountain waves.
Unstable atmosphere

Fig. 2b shows the time-averaged potential temperature for the 2DU35 simulation. No mountain waves develop in the lee of the mountain due to the unstable atmosphere. The potential temperature is at a maximum at the top of the leeward slope, where the rapid flow deceleration occurs, and is seen to decrease with both height and distance downstream of the mountain. The roughness of the potential temperature contour lines in the lee of
Numerical modelling of fire channelling

Int. J. Wildland Fire 605

Fig. 4. Instantaneous horizontal vorticity contour and wind barb plots at a time of 60 min for the (a) 2DS35 (b) 2DN35 and (c) 2DU35 simulations. The horizontal vorticity has a contour interval of 0.02 s$^{-1}$. The wind conditions are determined using the west–east and vertical wind components. Standard weather map wind barbs are used to indicate the wind speed (knots, 1 knot = 0.514 m s$^{-1}$) and the direction, with each full feather indicating an additional 10 knots (5.14 m s$^{-1}$) in wind speed. A subset of the full model domain is shown in each plot, with white shading used to represent the mountain.

Fig. 5. As in Fig. 3, but for the (a) 2DN25 and (b) 2DN35 simulations.
the mountain suggests that the flow becomes more turbulent with height further downstream of the mountain. This result implies that the unstable atmosphere acts to vertically lift the lee rotors as they propagate downstream of the mountain.

Fig. 6 shows the time-averaged horizontal wind velocity for the 2DU25 and 2DU35 simulations. There is a high degree of similarity between the two simulations and the acceleration and deceleration of the upslope and downslope flow is qualitatively similar to that discussed for the stable and neutral atmosphere simulations. In the unstable atmosphere there is greater acceleration of the upslope flow than in the stable and neutral atmospheres, with a peak velocity of \( \sim 40 \, \text{m s}^{-1} \) at the mountain peak. The leeward flow separation region in the unstable atmosphere simulations starts at a height of \( \sim 700 \, \text{m} \) on the leeward slope and extends \( \sim 4 \, \text{km} \) downstream of the mountain. As with the neutral atmosphere simulations, there is a slight uplift of the flow separation region above the surface at the base of the leeward slope.

Fig. 4c shows the horizontal vorticity and wind conditions at a time of 60 min for the 2DU35 simulation. The vorticity shedding process is qualitatively similar to that described for the 2DS35 and 2DN35 simulations. The lee rotors in this simulation have a greater initial horizontal and vertical extent than in either the neutral or stable atmosphere simulations. As the lee rotors propagate downstream, they are lifted due to the unstable atmosphere and grow in vertical extent. As these rotors extend vertically, they eventually break down into several smaller-scale sub-rotors.

Three-dimensional atmospheric simulations

In this section the results of three 3-D simulations of atmospheric flow over a triangular ridge line are presented. Details of the setup used in each simulation are provided in Table 1. The above analysis of the 2-D simulations shows that, for the range of conditions tested, the atmospheric stability plays a more important role than does the leeward slope angle in the leeward flow separation. Therefore, the 3-D simulations use a fixed leeward slope angle of 35° and the atmospheric stability is varied. Each 3-D simulation has a unique name derived from three properties: dimensionality (‘3D’), atmospheric stability (‘S’, ‘N’ or ‘U’) and leeward slope angle (‘35’). These simulations are used to investigate the important differences between the leeward flow in two and three dimensions.

Fig. 7 shows the time-averaged potential temperature on a vertical cross-section for the 3DS35 and 3DU35 simulations. It is evident that there is a high degree of similarity between the equivalent 2-D and 3-D simulations. The mountain waves formed in the stable atmosphere have a wavelength of \( \sim 10 \, \text{km} \) and amplitude of several hundred metres. There are no mountain waves in the unstable atmosphere, and the potential temperature is at a maximum at the top of the leeward slope and decreases with height and distance downstream of the mountain.

Fig. 8 shows the time-averaged horizontal velocity on a vertical cross-section for the 3DS35, 3DN35 and 3DU35 simulations. As in the equivalent 2-D simulations, there is acceleration of the flow as it moves up the windward slope, reaching a peak velocity at the mountain peak, and rapid deceleration and flow separation across much of the leeward slope and extending some distance downstream of the mountain. However, there are some important differences between the equivalent 2-D and 3-D simulations. First, the flow separation is typically weaker and more spatially confined in the 3-D simulations. Second, the flow separation is not lifted slightly above the surface as it was in the neutral and unstable 2-D simulations. Third, the time-averaged horizontal velocity in the 3-D simulations has a higher degree of spatial variability, as shown by the rougher horizontal velocity contour lines.
Fig. 7. As in Fig. 2, but for the (a) 3DS35 and (b) 3DU35 simulations. The vertical cross-section is taken through the middle of the $y$-axis.

Fig. 8. As in Fig. 3, but for the (a) 3DS35, (b) 3DN35 and (c) 3DU35 simulations. The vertical cross-section is taken through the middle of the $y$-axis.
Fig. 9 shows the horizontal vorticity and wind conditions at a time of 60 min on a vertical cross-section for the 3DS35, 3DN35 and 3DU35 simulations. The horizontal vorticity is calculated identically to how it was calculated for the 2-D simulations. As in the equivalent 2-D simulations, a quasi-permanent region of strong positive vorticity develops at the top of the leeward slope and extends out eastwards \( \sim 1 \text{ km} \) from the leeward slope. However, there are no large-scale lee rotors and the leeward flow is instead dominated by chaotic fine-scale features. The 3-D turbulence is therefore subject to energy cascade down to smaller-scales than was seen in 2-D.

Doyle and Durran (2007) have previously considered some of the important differences between the 2-D and 3-D atmospheric turbulence downstream of a mountain. They found that the 2-D turbulence is steadier and greater in spatial extent, whereas the 3-D turbulence is more chaotic and fine-scale. They proposed that processes such as tilting and stretching of vortical structures between different directional components may be important in 3-D. The differences between the turbulence seen in these 2-D and 3-D simulations are consistent with this previous study, however the tilting and stretching of vortical structures is not specifically considered in this study. The absence of large-scale lee rotors in these 3-D simulations does not support the fire channelling hypothesis proposed by Sharples et al. (2012).

**Coupled atmosphere–fire simulations**

In this section the results of four WRF-Fire simulations of fire spread on a leeward slope are presented. Details of the setup used in each simulation are provided in Table 1. Each simulation has a unique name derived from two properties: atmosphere–fire coupling (‘FC’ for atmosphere–fire coupling or ‘FN’ for no atmosphere–fire coupling), and fuel type (‘F05’ for brush fuel type or ‘F13’ for heavy logging slash fuel type). By comparing simulations with the atmosphere–fire coupling enabled or disabled, a direct evaluation of the importance of atmosphere–fire interactions on the fire spread can be made. The ‘F05’ and ‘F13’ fuel types are based upon the ‘Brush (2 feet)’ and ‘Heavy Logging Slash’ Anderson fuel categories (Anderson 1982) and were chosen as they have very different fuel properties.

**Fire spread on leeward slope**

Fig. 10a shows the fuel fraction remaining at times of 60, 90 and 120 min for the FCF05 simulation. Between 30 and 60 min, the fire spreads predominantly upslope asymmetrically and comes within 20 m of the mountain ridge line. As the fuel conditions and leeward slope angle are constant, this fire spread asymmetry must be due to a combination of the turbulent mid-flame wind conditions and the asymmetric fire ignition pattern. Between 60 and 90 min, the fire spreads in a predominantly lateral direction
Fig. 10. Contour plots of the instantaneous fuel fraction remaining at times of 60, 90 and 120 min for the (a) FCF05, (b) FNF05, (c) FCF13 and (d) FNF13 simulations. White shading is applied to regions where the fuel fraction remaining is over 99% or under 1%. Terrain contour lines are given at 100-m intervals and the solid black lines represent the mountain ridge line and the base of the leeward slope. The fire ignition region is indicated by the dash-filled region. A subset of the full SFIRE model domain is shown in each plot.
and there is no fire spread west of the ridge line due to the strong westerly flow across the ridge line. Between 90 and 120 min, the fire spread is still predominantly lateral, with some fire spread downslope of the ignition region. By 120 min, the fire perimeter has a maximum south–north and west–east extent of \( \sim 1.7 \) and \( \sim 1.3 \) km respectively.

Fig. 10b shows the fuel fraction remaining at times of 60, 90 and 120 min for the FNF05 simulation. Between 30 and 60 min, the fire spreads both laterally and upslope, and the fire perimeter is approximately elliptical. Between 60 and 90 min, the fire spreads upslope to the mountain ridge line, but there is no further westward spread due to the strong westerly winds across the windward slope. The fire spread during this period is therefore predominantly lateral, in particular to the north. Between 90 and 120 min, the fire spread is still predominantly lateral, with some fire spread downslope of the ignition region. By 120 min, the fire perimeter has a maximum south–north and west–east extent of \( \sim 2.0 \) and \( \sim 1.3 \) km respectively.

The similarity between the fire spread in these two simulations suggests that atmosphere–fire interactions do not play a significant role in the FCF05 simulation. Instead, the fire spread is primarily driven by the steep slope and the turbulent wind conditions at the mid-flame height. The only possible exception to this is the initial 30-min period of predominantly upslope fire spread after ignition. The upslope fire spread occurs at a faster rate in the coupled simulation, which indicates that the pyro-convection must influence the wind conditions across the leeward slope at the mid-flame height. The lateral fire spread does not seem to be closely associated with the distance from the mountain ridge line, which implies that the lateral fire spread seen is not consistent with fire channelling. Fig. 10c shows the fuel fraction remaining at times of 60, 90 and 120 min for the FCF13 simulation. Between 30 and 60 min, the fire spreads predominantly upslope at an average rate of \( \sim 1.3 \) km/h. By 60 min, the fire perimeter is approximately elliptical and extends up to 100 m west of the mountain ridge line. Between 60 and 90 min, the fire spread is dominated by lateral spread both northwards and southwards directly to the east of the ridge line. The northwards and southwards lateral spread occurs at a respective average rate of \( \sim 1.9 \) and \( \sim 1.0 \) km/h. Between 90 and 120 min, the fire spread is still dominated by lateral spread close to the ridge line at an average rate of \( \sim 1.0 \) km/h. By 120 min, the fire perimeter has a highly asymmetric shape, with a maximum south–north extent of \( \sim 2.8 \) km directly to the east of the ridge line.

Fig. 10d shows the fuel fraction remaining at times of 60, 90 and 120 min for the FNF13 simulation. Between 30 and 60 min, the fire spreads predominantly upslope up to a height of \( \sim 800 \) m on the leeward slope, which is well below the mountain ridge line. By 60 min the perimeter is approximately circular with a diameter of \( \sim 1.0 \) km. Between 60 and 90 min, the fire spread both upslope and laterally, however the westward spread does not extend beyond the mountain ridge line. Between 90 and 120 min, the fire spreads in a predominantly lateral direction, however there is also limited fire spread to the west of the ridge line and downslope from the ignition region. By 120 min, the fire perimeter is approximately elliptical, with a maximum south–north and west–east extent of \( \sim 1.9 \) and \( 1.4 \) km respectively.

The fire spread seen in the FNF13 and FNF05 simulations is very similar. This suggests that the fire spread is more closely associated with the slope angle and wind conditions than with the fuel conditions. The fire spread seen in the FCF13 simulation is very different to the other three simulations and shares in common several characteristics with the fire spread expected during fire channelling. The rapid lateral fire spread occurs intermittently in both directions across the leeward slope in close proximity to the mountain ridge line. The fire spread seen to the west of the mountain ridge line, however, is not typically seen during fire channelling. The downwind extension of the active flaming zone expected during fire channelling is not seen in this simulation. However, this can be explained by the absence of spotting in the WRF-Fire model, which likely plays an important role in the downwind extension of the active flaming zone. A direct comparison of the FCF13 and FNF13 simulations indicates that the atmosphere–fire interactions play an important role in the fire spread for this fuel type. It is likely that the rapid lateral fire spread seen close to the ridge line results from a modification of the mid-flame wind conditions by pyro-convection.

**Dynamical atmosphere–fire interactions**

The previously discussed differences in the fire spread seen between the heavy logging slash and brush fuel type simulations is likely to be closely associated to the pyro-convection and the resulting atmosphere–terrain–fire interactions. The heavy logging slash fuel type has a significantly higher fuel depth and density than does the brush fuel type, giving it a higher fuel mass per unit area and a higher corresponding sensible and latent heat release rate.

Fig. 11a shows the time-varying total heat release rate from the fire for the FCF05 and FNF05 simulations. The heat release rate is very similar between the two simulations, which further demonstrates that the atmosphere–fire interactions do not significantly influence the fire spread in the FCF05 simulation. Fig. 11b shows the time-varying total heat release rate from the fire for the FCF13 and FNF13 simulations. The total heat release rate is approximately an order of magnitude larger for the heavy logging slash fuel type, which is closely associated with its higher fuel depth and density. Between 50 and 90 min, the heat release rate is higher in the coupled simulation, which implies that the atmosphere–fire interactions make a significant difference to the fire spread before the fire perimeter first reaches the mountain ridge line at a time of \( \sim 56 \) min. The FCF13 simulation has a local and global maximum in the total heat release rate at times of \( \sim 64 \) and \( \sim 75 \) min. These correspond to the times at which there is significant lateral fire spread in close proximity to the mountain ridge line.

Fig. 12 shows the potential temperature anomalies and wind conditions at a time of 56 min on a vertical cross-section for the FCF13 simulation. The potential temperature anomalies are a useful indicator of pyro-convection and are calculated relative to the vertically averaged potential temperature upstream of the mountain. A pyro-convective plume, which is tilted eastwards by the westerly flow, is visible downstream of the mountain up to a height of \( 4 \) km. There are positive anomalies, associated with ongoing combustion, across much of the leeward slope,
extending from the fire ignition region to the mountain ridge line. The pyro-convection is at a maximum directly eastwards of the ridge line and results in strong upslope flow across much of the leeward slope. This upslope flow associated with the pyro-convection contributes to the high average upslope fire spread rate between 30 and 60 min. Strong horizontal convergence between the windward and leeward slope flows is seen at the ridge line, which results in a strong updraft near the ridge line. The downstream wind conditions indicate that there is an extensive turbulent wake downstream of the pyro-convective plume. The pyro-convective plume dissipates quickly with distance away from the mountain, indicating that there is a high level of mixing and entrainment of mean flow with the plume. The pyro-convective plume dynamics may have important implications for downstream transport of firebrands, however this is not specifically considered in this study.

Fig. 11. Timeseries of the total heat release rate for the (a) brush and (b) heavy logging slash fuel type simulations. In each simulation the fire is ignited at a time of 30 min, so the total heat release rate before this time is zero. The total heat release rate, which is determined across the full SFIRE model domain, is measured in gigawatts and the y-axis scales in (a) and (b) are an order of magnitude different.

Fig. 12. Potential temperature anomaly contours and wind barbs at a time of 56 min for the FCF13 simulation. The potential temperature anomaly has a contour interval of 3 K. The wind conditions are determined using the west–east and vertical wind components. Standard weather map wind barbs are used to indicate the wind speed (knots) and the direction, with each full feather indicating an additional 10 knots (5.14 m s⁻¹) in wind speed. A subset of the full model domain is shown in each plot, with white shading used to represent the mountain. The vertical cross-section is taken at the mid-point of the y-axis.

Fig. 13 shows the fire perimeter, vertical wind velocities taken at a height of ~16 m, and the horizontal wind conditions taken at the mid-flame height, at a time 60, 66 and 72 min for the FCF13 simulation. By 60 min, the fire has spread westwards from the ignition region to ~100 m west of the mountain ridge line. By this time the fire spread has been predominantly upslope, driven partly by the strong easterly flow associated with the pyro-convection. A downdraft region is visible ~100 m to the south of the fire perimeter and directly to the east of the ridge line. This downdraft region is in close proximity to the base of the pyro-convective plume, which extends across much of the fire area. The inflow and outflow associated with these updraft and downdraft regions interact to form a region of counterclockwise rotating flow at the mid-flame height in close proximity to the southern flank of the fire. Between 60 and 66 min, this updraft–downdraft interface moves across the
southern flank of the fire perimeter and the associated counter-clockwise rotating flow acts to spread the fire in a lateral direction close to the ridge line. Between 66 and 72 min, this lateral fire spread continues at an average rate of $3.6 \text{ km h}^{-1}$ and ignites an area of $12 \text{ ha}$. This rapid lateral spread rate is therefore significantly higher than the average upslope spread rate of $2.0 \text{ km h}^{-1}$ between 30 and 60 min. By 72 min the updraft–downdraft interface moves inside the fire perimeter and the lateral spread rate rapidly decreases to nearly zero.

**Summary and conclusions**

The atypical wildland fire spread phenomenon known as ‘fire channeling’ has been investigated through a series of idealised 2-D and 3-D numerical simulations. These simulations were performed using the WRF-Fire coupled atmosphere–fire numerical model and provide important new insights into the physical processes responsible for the lateral fire spread seen during fire channelling.

The 2-D atmospheric flow over a triangular mountain was investigated under varying atmospheric stability and leeward slope angle conditions. In each simulation there was acceleration of the flow as it moved up the windward slope and rapid deceleration of the flow as it moved down the leeward slope. This acceleration and deceleration of the flow is most likely due to the pressure gradient encountered by the flow as it moves across the mountain. The rapid leeward flow deceleration resulted in flow separation across much of the leeward slope and several kilometres downstream of the mountain. This flow separation was associated with the periodic formation of large-scale rotors in the lee of the mountain, which were then seen to detach from the mountain and propagate downstream. In a stable atmosphere, mountain waves were seen to develop in the lee of the mountain, with associated flow separation under the first downstream mountain wave crest. The wavelength of these mountain waves was, however, too long for any direct effect of the mountain waves on the flow across the leeward slope. For the range of conditions tested, it was found that the leeward flow behaviour was more closely associated with the atmospheric stability than with the leeward slope angle.

The 3-D atmospheric flow over a triangular mountain was investigated under varying atmospheric stability conditions. The atmospheric flow features seen in the 3-D simulations were found to be broadly similar to the flow features of the equivalent 2-D simulations, but with some important differences. First, the instantaneous horizontal vorticity indicated that the 3-D flow was dominated by chaotic fine-scale features, with an absence of the large-scale lee rotors seen in 2-D. Second, the flow separation close to the leeward slope was typically weaker and more spatially confined in the 3-D simulations. These results are consistent with the previous study by Doyle and Durran (2007), who identified that 3-D tilting and stretching of vortical structures may explain these differences, although this is not specifically tested in this study. The absence of large-scale lee
Numerical modelling of fire channelling

Doyle JD, Durran DR (2002) The dynamics of mountain-wave-induced rotors in the 3-D simulations allows this study to refute the fire channelling hypothesis proposed by Sharples et al. (2012), which depends upon their existence.

The fire spread across a leeward slope was investigated for two different fuel types and with the atmosphere–fire coupling switched either on or off. The two fuel types considered were based on the brush and heavy logging slash Anderson fuel categories (Anderson 1982). The patterns of fire spread in the coupled and non-coupled brush fuel type simulations were very similar, indicating that the atmosphere–fire interactions did not play an important role in the fire spread for this fuel type. This was supported by examination of the total fire heat release rate, which was very similar for the two simulations and low in both cases, suggesting little pyro-convection. The fire spread was very different between the coupled and non-coupled heavy logging slash fuel type simulations and in the coupled simulation was found to closely resemble that expected during fire channelling. The fire spread was initially predominantly upslope until it reached the mountain ridge line, after which the fire spread was dominated by intermittent rapid lateral spread in close proximity to the ridge line. The atmosphere–fire interactions were found to make an important difference to the fire spread for the heavy logging slash fuel type, with a noticeable increase in the total heat released from the fire for the coupled simulation.

The intermittent rapid lateral fire spread seen in the coupled heavy logging slash fuel type simulation was found to be driven by a process where an updraft–downdraft interface moved across the fire perimeter. The inflow and outflow associated with these updrafts and downdrafts resulted in either clockwise or counterclockwise flow rotation at the mid-flame height near the northern and southern flanks of the fire. When an updraft–downdraft interface moved across the fire perimeter, the associated rotating flow acted to significantly increase the lateral spread rate up to 3.6 km h\(^{-1}\) close to the mountain ridge line for a period of a few minutes. When compared with the average spread rate of \(-2.0 \text{ km h}^{-1}\) during the initial run up the leeward slope with the upslope wind, this lateral spread rate is significantly higher. This result is important given that the upslope spread is traditionally assumed to yield the highest rates of spread. The formation of these updraft–downdraft interfaces results from an interaction of the pyro-convection and the terrain-modified winds though precise mechanisms remain to be investigated. This proposed physical process shares some characteristics in common with the description of the generation of fire whirls on leeward slopes given by Countryman (1971), who proposed that fire whirls are likely to develop wherever there is significant convection and also eddies in the atmospheric flow. The dynamics of the convective plume seen in the coupled simulation suggest that spotting could result in the transport of firebrands downstream of the mountain. Such a spotting process could explain the downwind extension of the active flaming zone typically seen during fire channelling.

This study has provided several important new insights into the fire channelling phenomenon. First, this study has shown that it is possible to simulate the fire channelling phenomenon using the WRF-Fire coupled atmosphere–fire model. Second, the study has provided important new insights into the physical processes that may be driving the atypical fire spread seen during fire channelling. Future work will incorporate additional coupled atmosphere–fire simulations to better understand the environmental conditions, such as fuel type, terrain configuration and atmospheric stability, required for fire channelling to occur. The atypical fire spread seen in these simulations suggests that the fire channelling phenomenon has particular relevance to fire management and fire fighter safety.

Acknowledgements

This project was supported by a UNSW Engineering-ADFA research collaboration scheme grant. This research was undertaken as part of a doctoral thesis funded by the Bushfire Cooperative Research Centre, Melbourne, Australia. We acknowledge beneficial discussions with Keith Ayotte, Todd Lane, John Taylor and Michael Reeder on high resolution atmospheric modelling. Computer simulations were performed using the BlueFern supercomputing system operated by the University of Canterbury, New Zealand.

References

Countryman CM (1971) Fire whirls…why, when, and where. USDA Forest Service, Pacific Southwest Forest and Range Experiment Station, Technical report (Berkeley, CA)


Rothermel RC (1972) A mathematical model for predicting fire spread in wildland fuels. USDA Forest Service, Intermountain Forest and Range Experiment Station, Research Paper INT-115. (Ogden, UT)


