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'Fire and forest meteorology'

Edited by

SA Ferguson

TJ Brown

M Flannigan

For research on the
physical, ecological,
economic and social
aspects of wildland fire

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International Journal of Wildland Fire

Scientific Journal of the International Association of Wildland Fire

International Journal of Wildland Fire is dedicated to the advancement of basic and applied research concerning wildland fire. The journal has an international perspective, since wildland fire plays a major social, economic and ecological role around the globe. We welcome original scientific papers on all aspects of wildland fire, especially those aimed at understanding the principles of wildland fire as a process, understanding its ecological impact at the stand and landscape levels, modelling fire and its effects, and presenting information on how to effectively manage fire. The Journal wishes to attract papers on a broad range of wildland fire issues that may include subjects beyond the range of papers published in recent issues.

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Preface

The American Meteorological Society symposia series on Fire and Forest Meteorology provides biennial forums for atmospheric and fire scientists to introduce and discuss the latest and most relevant research on weather, climate and fire. This special issue highlights significant work that was presented at the Fifth Symposium in Orlando, Florida during 16–20 November 2003, and held jointly with the Second International Wildland Fire Ecology and Fire Management Congress. This is the second special issue in the *International Journal of Wildland Fire* from these symposia—the first appeared in Volume 11, Numbers 3 & 4, 2002, based on papers presented at the Fourth Symposium.

While the symposia were born out of a historic need for accurate fire weather predictions to help suppress wildfires that threatened timber products, the context has broadened dramatically over the years. Now, in addition to wildfire suppression, the world of fire includes assessments and predictions for prescribed fire, smoke management and seasonal fire potential. Research examines land–atmosphere processes that control fire rehabilitation, coupled fire–atmosphere physics, and the relation of fire to the global carbon budget and climate change. Grass, shrublands and the wildland–urban interface now share the stage with timbered forests. Weather and climate at all spatial and temporal scales—past, current and future—are relevant to fire.

The Fifth Symposium program was developed by Timothy J. Brown and Sue A. Ferguson, co-chairpersons, and by committee members David Billingsly, Barbara Bonefeld, Larry Bradshaw, Mark Finney, Francis M. Fujioka, Scott Goodrick, Beth L. Hall, Tom McClellan, Richard Ochoa, Brian Potter and Carol Rice. Session topics included mesoscale meteorology, predictive services and operational forecasting, and atmospheric stability and fire behavior. There were also a number of joint sessions with the Second Congress meeting. These included GIS/remote sensing, fire and atmosphere interactions, combustion modeling, risk assessment and decision support, wildfire burn severity mapping, smoke management and air quality, assessing and predicting climate impacts on landscapes, Landfire (a fire, ecosystem, and fuel assessment-mapping project), fire and drought indices,

and Fire Consortia for Advanced Modeling of Meteorology and Smoke. There were also opening and closing plenary sessions, and a panel discussion on Fair Weather: Effective Partnerships in Weather and Climate Services.

There were 169 Symposium participants, and 628 individuals attended the combined meetings. Including the joint sessions, 125 papers and 20 posters were presented. Many of these were submitted to the proceedings volume published November 2003 by the American Meteorological Society, 45 Beacon Street, Boston, Massachusetts 02108-3693, USA. The proceedings are dedicated to the memory of David W. Goens, who was instrumental in organizing the first three Symposia on Fire and Forest Meteorology. While all Symposium participants were invited to submit their work to this special issue of the *International Journal of Wildland Fire*, only nine manuscripts contained sufficient detail of background, methods and results to meet Journal standards. These manuscripts cover a broad spectrum of interests including smoke, modeling and dynamics. We hope they will be found beneficial as a scientific contribution, and perhaps some will find immediate utility in the world of applications. They represent fire–atmospheric science research early in the 21st Century, and a hint as to just how much more research work there is to do. In addition to this special issue, a number of papers were submitted for consideration of a special issue in the journal *Agricultural and Forest Meteorology* to appear in early 2005.

We would like to express much gratitude to the authors, reviewers and editors involved in this special issue, and of course to the participants of the Symposium for presenting their ideas and research results that ultimately lead to another special issue. We are indebted to the sponsors of the joint meeting. These include the Joint Fire Sciences Program, NOAA/National Weather Service, NOAA Office of Global Programs, U.S. Fish and Wildlife Service, National Park Service, U.S. Forest Service, Fire Program Solutions, LLC, and the U.S. Bureau of Land Management.

Timothy J. Brown and Sue A. Ferguson
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Seasonal fire danger forecasts for the USA

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Abstract. The Scripps Experimental Climate Prediction Center has been making experimental, near-real-time, weekly to seasonal fire danger forecasts for the past 5 years. US fire danger forecasts and validations are based on standard indices from the National Fire Danger Rating System (NFDRS), which include the ignition component (IC), energy release component (ER), burning index (BI), spread component (SC), and the Keetch–Byram drought index (KB). The Fosberg fire weather index, which is a simplified form of the BI, has been previously used not only for the USA but also for other global regions and is thus included for comparison. As will be shown, all of these indices can be predicted well at weekly times scales and there is even skill out to seasonal time scales over many US West locations. The most persistent indices (BI and ER) tend to have the greatest seasonal forecast skill. The NFDRS indices also have a weak relation to observed fire characteristics such as fire counts and acres burned, especially when the validation fire danger indices are used.

Introduction

Predicting the influence of weather on fire ignition and spread is an operational requirement for national and global fire planning by the National Interagency Coordination Center (NICC), which is the USA's support centre for wildland fire-fighting. NICC is home to seven federal agencies including the Bureau of Land Management, National Park Service, Fish and Wildlife Service and Bureau of Indian Affairs, all in the Department of the Interior; and the Forest Service, in the Department of Agriculture. NICC's Predictive Services produces national wildland fire outlook and assessment products at weekly to seasonal time scales. This is currently done by considering standard National Weather Service (NWS) seasonal forecast products of temperature and precipitation (see Brown *et al.* 2003) along with other indicators, and carefully exercised human judgment.

By contrast, nowcasts of fire danger potential at individual locations have been carried out for decades at individual station locations using the US Forest Service (USFS) National Fire Danger Rating System (NFDRS) (Deeming *et al.* 1977). This process has been automated and implemented nationwide, resulting in web-based displays of the NFDRS indices. The NFDRS explicitly describes the effects of local topography, fuels and weather on fire potential. Fuel moisture models

relate moisture content to cumulative precipitation, precipitation extent and variation, temperature and relative humidity. These fire danger nowcasts are updated almost daily, but they allow fire managers to react only to the current weather and climate conditions, rather than plan for the upcoming fire season.

The goal of this paper is to assess whether the NFDRS indices could also be forecast with a state of the art dynamical seasonal prediction model in the hopes that automatic seasonal forecasts could eventually be developed for NICC predictive services. Again, official NWS forecasts are issued only for temperature and precipitation. Forecasts for several more fire-relevant variables, such as relative humidity and wind speed, are still experimental and in many cases the fire community has had to empirically adapt to the official NWS forecasts of temperature and precipitation.

As described previously by Roads *et al.* (2001a, 2001b), Roads and Brenner (2002) and Chen *et al.* (2001), as part of this effort, the Scripps Experimental Climate Prediction Center (ECPC) has been routinely making experimental, near-real-time, long-range dynamical forecasts since 27 September 1997 of several additional variables relevant to fire danger forecasts. Images from these forecasts are regularly shown on the worldwide web site <http://ecpc.ucsd.edu/>

(Roads *et al.* 2003c). The global model is a version of the National Centers for Environmental Prediction's (NCEP's) global spectral model (GSM; Kalnay *et al.* 1996; Roads *et al.* 1999) used for the NCEP/NCAR (National Center for Atmospheric Research) reanalysis. With the GSM forecasts as boundary conditions a higher resolution regional spectral model (RSM; Juang *et al.* 1997) is also run for various US and international regions (see e.g. Roads *et al.* 2003a, 2003b) to provide increased geophysical detail. The initial conditions and sea surface temperature (SST) boundary conditions for these experimental global forecasts come from the NCEP Global Data Assimilation (GDAS) 00 UTC operational analysis, which is available nearly every day in near real time on NCEP rotating disk archives, to interested researchers. Transforming NCEP's higher-resolution operational global analyses to lower (vertical and horizontal) resolution initial conditions for the global model, 7-day global and regional forecasts are made every day and every weekend these global and regional forecasts are extended to 16 weeks.

The Experimental Climate Prediction Center's experimental forecasts are certainly not superior to official forecasts from NCEP, which use not only similar dynamical models (ECPC models are actually older fixed versions of NCEP's constantly improving models), but also take into account other climatic features that are not yet adequately represented in any dynamic model (i.e. various climatic trends, tropical teleconnections, innate human forecast experience, etc.). However, the documented skill of our dynamical system (e.g. Roads *et al.* 2001a) does seem to at least be comparable to official forecasts, which indicates that these experimental forecasts may at least be a useful research tool for developing various forecast applications.

We have therefore attempted to develop experimental forecasts of the National Fire Danger Rating System (NFDRS; Deeming *et al.* 1972) in order to augment current USFS nowcasts from station observations and current seasonal forecast output of only temperature and precipitation. Basically, because our dynamical models have demonstrated some skill for forecasting various meteorological variables like temperature, relative humidity and mean wind speed at seasonal time scales, we wish to determine whether the perceived meteorological forecast skill can carry over to forecasts of fire danger and whether the federal fire agencies should develop a more comprehensive seasonal fire danger forecasting capability. Encouragingly, Roads *et al.* (2001a) did show that a simplified measure of fire danger, namely the Fosberg (1978) fire weather index (FWI) was capable of being predicted at seasonal time scales, mainly because of the inherent predictability of relative humidity, which is a significant component of the FWI, and as we shall see, other NFDRS indices.

In particular, we examine here US regional spectral model (RSM) weekly to seasonal forecasts of fire danger. The next section describes the methodology (models, validating

analysis, error measures) used for this study. Then we compare the geographic temporal and forecast lead variations, and finally present a summary.

Methodology

Global spectral model

The global spectral model (GSM), which was used to provide the boundary conditions for the RSM, has been described previously in many papers (e.g. Roads *et al.* 2003a). Again, the GSM is based on the medium range forecast (MRF) model used for the NCEP reanalyses (Kalnay *et al.* 1996; Kanamitsu *et al.* 2002). The GSM has a triangular truncation of T62 (192×94 global Gaussian grid) and 18 irregularly spaced vertical levels. Seven-day GSM forecasts, initialized every day from the 00 UTC NCEP GDAS operational analysis, and 16-week GSM forecasts made every weekend, provide the basic large-scale driving data for the RSM. During the weekly 16-week (and also daily 7-day) forecasts, it is assumed that the initial SST anomaly is constant (persistent). The GDAS initial conditions have been accessed almost every day from 00 UTC 27 September 1997 to present (28LT126 NCEP global analyses were available from 00 UTC 27 September 1997 to 18 UTC 14 March 2000; 42LT170 were available from 00 UTC 15 March 2000 to 18 UTC 14 July 2002; 62LT256 became available 00 UTC 15 July 2002). These higher resolution analyses are transformed to lower resolution initial conditions (18LT62) by linearly interpolating between vertical sigma levels, spectrally truncating the spectral components, and bilinearly interpolating the higher resolution surface grids to our lower resolution grids (and land mask).

Regional spectral model

The RSM was initially developed by Juang and Kanamitsu (1994; see also Juang *et al.* 1997) to provide a regional extension to the GSM, and thus in principle provides an almost seamless transition between the RSM and the GSM or the associated NCEP reanalyses (Kalnay *et al.* 1996; Kanamitsu *et al.* 2002) and the higher resolution region of interest. For this experiment, the grid spacing was chosen to be 60 km at the central point. This horizontal resolution (and 18 vertical levels) is sufficient to resolve many features of interest, although eventually even higher resolution simulations and forecasts need to be attempted for the US West, where the topography exerts a strong control on near-surface features.

Both the GSM and RSM use the same primitive hydrostatic system of virtual temperature, humidity, surface pressure and mass continuity prognostic equations on terrain-following sigma (sigma is defined as the ratio of the ambient pressure to surface pressure) coordinates. Therefore, in the absence of any regional forcing (and intrinsic internal dynamics, any significant physical parameterization differences, and significant spatial resolution), the total RSM solution should

be identical to the GSM solution. A minor structural difference is that the GSM utilizes vorticity, divergence equations, whereas the RSM utilizes momentum equations in order to have simpler lateral boundary conditions. The GSM and RSM horizontal basis functions are also different. The GSM uses spherical harmonics with a triangular truncation of 62 (T62) whereas the RSM uses cosine or sine waves to represent regional perturbations about the imposed global scale base fields on the regional grids. The double Fourier spectral representations are carefully chosen so that the normal wind perturbations are anti-symmetric about the lateral boundary. Other model scalar variables (i.e. virtual temperature, specific humidity and surface log pressure) use symmetric perturbations. Finally, except for the scale-dependent horizontal diffusion, the GSM and RSM physically are, in principle, identical. However, there are some minor parameterization differences between the NCEP GSM and this RSM, which has an upgraded physics package comparable to what was used for the NCEP Reanalysis II (see Kanamitsu *et al.* 2002).

The RSM can be initialized directly from the GSM, analysis or reanalysis, or it can use a previous integration to initialize itself. Roads *et al.* (2003a) showed that it was best to initialize the RSM every day from the global analysis, as continuous runs affected the regional solution adversely. The first part of an RSM forecast or simulation involves the integration of the GSM, for a nesting period based on the large-scale output. Here, the RSM predicts regional deviations from the large-scale atmosphere base fields, which are linearly interpolated in time between two output periods (6 h). The non-linear advection is first computed at the model grid points by transforming the global and regional spectral components to the regional grid. The global quantities are transformed to the global grid and then bilinearly interpolated to the regional grid; the regional quantities are exactly transformed. These calculations are almost exact (except for the interpolation error for the global quantities) and thus, like the global model, the regional model is free of aliasing and phase error. The linearly interpolated global-scale tendency is then removed so that, in effect, only the portion affecting the regional perturbation is retained. At the horizontal boundaries, the perturbation amplitude approaches zero by a damping function increasing rapidly towards the lateral boundary, which ensures that the boundary tendencies are similar to the original GSM tendencies and features. A semi-implicit time integration scheme is employed to suppress computational modes and also to allow the use of longer time integration steps.

Fire danger indices

The NFDRS indices describe characteristics of fire danger, given the conditions of fuel, topography and weather (Fig. 1a,b). Note that the basic inputs to the NFDRS include precipitation, temperature, relative humidity and wind speed as well as fuels and slope. The standard weather input to the

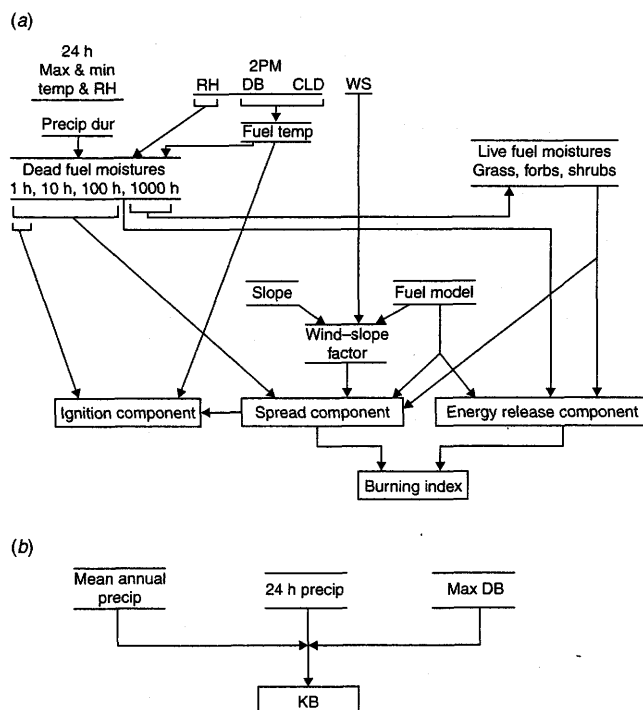


Fig. 1. (a) National Fire Danger Rating System indices and necessary inputs. (b) The Keetch-Byram drought index. DB, dry bulb temperature; CLD, cloudiness; KB, Keetch-Byram drought index; RH, relative humidity; WS, wind speed.

NFDRS comes from weather station data, which is assumed to apply to a large but undefined area surrounding each weather station; vegetation (fuel) types and slope are also defined for each weather station and are assumed to apply to the same surrounding area. The major difference from standard NFDRS calculations is that here gridded fuels, weather forecasts and topography data are used. The fuels and orography (slope) data were defined initially at 1 km spatial resolution and then the nearest 1 km grid point was used for a coarser-scale 100 km grid. The observed precipitation (25 km) and forecast model output (60 km) was similarly interpolated to the 100 km grid. Ultimately, the NFDRS indices calculated on the 100 km grid were interpolated back to the standard RSM 60 km grid.

For later comparison to the gridded fire danger indices, Fig. 2 shows the gridded fuels (see also Table 1) and slopes used for these calculations. Slopes (Fig. 2b) are used to reflect the fact that fire burns faster upslope than on flat ground. The type of vegetation or available fuel (Fig. 2a) is also important for describing the fire danger. Sixteen (Table 1) of the 20 NFDRS fuel models were used to represent the vegetation types across the USA (Burgan 1988), defining fuel characteristics such as depth, load by live and dead classes, heat content, fuel particle size, etc. Because the variability of fuel bed characteristics is almost infinite, each fuel model in the fire danger rating system must necessarily represent a rather

(a)



(b)

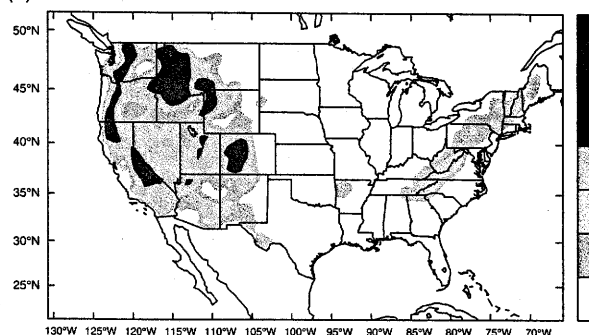


Fig. 2. (a) Fuel models used for these National Fire Danger Rating System (NFDRS) calculations (see Table 1). Fuel pixels are plotted on the 100 km fire danger model grid. (b) Slope used for these NFDRS calculations. Slopes interpolated to the RSM 60 km grid.

broad range of vegetation types. Basic data source was the 1 km resolution land cover map released in 1991 by the EROS Data Center (Loveland *et al.* 1991). The LC map was converted to an NFDRS fuel model map through a combination of 2546 ground sample plots scattered across the USA, and consultation with fire managers from across the country. Again these fuel models and slopes were defined at 1 km resolution but the fire danger calculations were done on a 100 km grid and then interpolated to the RSM 60 km grid for these comparisons.

It should be noted that we do not include all of the standard NFDRS indices and components. Human- and lightning-caused fire occurrence indices, and the fire load index in the NFDRS (Deeming *et al.* 1972, 1977; Burgan 1988) were not used because they have not been particularly useful to most fire managers. The outputs of the NFDRS used here were the spread component (SC), energy release component (ER), burning index (BI), ignition component (IC), Keetch-Byram (KB; Keetch and Byram 1968) drought component and Fosberg fire weather index (FWI), which are further defined below. Basic equations for SC, ER, BI and IC are described by Cohen (1985), and additional NFDRS technical documentation is provided by Bradshaw *et al.* (1983).

1. Spread component is an index of the forward rate of spread at the head of a fire. A spread component of 30 means that the fire should be progressing about 30 feet per minute

Table 1. Fuel models used for the coterminous USA
The slash models (I, J, K) and the leaf-off hardwood model (E) were not mapped

Vegetation type/ fuel model	Description
Grass	
A	Western annual grasslands
C	Open pine stands where grasses and forbs are primary ground fuel
L	Western perennial grasslands
N	Sawgrass prairies of south Florida
Shrub	
B	Mature California mixed chaparral, 6+ feet tall.
D	Palmetto-gallberry understory-pine overstory of SE coastal plains
F	Mature closed chamise and oakbrush, young chaparral, pinyon-juniper
O	Dense brush fields of SE United States, typically 6+ feet tall
S	Alpine tundra with low shrubs, moss, lichens
T	Sagebrush-grass types of the Great Basin and Intermountain West
Conifer	
G	Dense conifer with heavy accumulation of litter and downed woody material
H	Healthy stands of short needle conifers
P	Closed, thrifty stands of long-needled southern pines
Q	Upland Alaska black spruce or jack pine stands in the Lake states
U	Closed stands of western long-needled pines
Hardwood	
R	Hardwood areas after the canopies leaf out

at the head. The SC is quite sensitive to wind speed, and moderately sensitive to slope, especially if the fuel is dry. The SC is expressed on an open-ended scale; it has no upper limit.

2. Energy release component is a number related to the available energy per unit area within the flaming front at the head of a fire. That is, if you think of a fire burning through an area of vegetation, the ER is a measure of the amount of heat released from the time the head of the flaming front first enters an area, until the tail of the flaming front leaves it. The ER is a cumulative or 'build-up' type of index. As live fuels cure and dead fuels dry, the ER values get higher, thus providing a reflection of drought conditions. The scale is open-ended and, as with the other components, is relative. Daily variations in the ER are due to changes in moisture content of the live and dead fuels. ER is not affected by wind speed, thus it is reasonably stable from day to day.
3. Burning index is a number related to the contribution of fire behaviour to the effort of containing a fire. The BI is derived from a combination of the SC and the ER, so it is strongly affected by wind speed, and varies considerably from day to day. It has an open-ended scale.

Table 2. Seasonal correlations among the various USFS fire danger indices (IC, BI, ER, KB, SC), FWI, fire statistics (CN and AC), and a few meteorological variables (P, WSP, T, RH)

	FWI	P	WSP	T	RH	CN	AC	IC	BI	ER	KB	SC
FWI	1.00											
P	-0.32	1.00										
WSP	0.38	0.28	1.00									
T	0.60	-0.13	-0.05	1.00								
RH	-0.89	0.40	-0.06	-0.75	1.00							
CN	0.15	-0.10	-0.02	0.22	-0.19	1.00						
AC	0.36	-0.25	0.07	0.31	-0.40	0.35	1.00					
IC	0.59	-0.26	0.17	0.34	-0.55	0.22	0.29	1.00				
BI	0.65	-0.32	0.13	0.46	-0.67	0.26	0.38	0.90	1.00			
ER	0.43	-0.25	0.01	0.43	-0.53	0.29	0.31	0.54	0.71	1.00		
KB	0.18	-0.17	-0.07	0.22	-0.26	0.11	0.21	0.12	0.33	0.37	1.00	
SC	0.54	-0.23	0.18	0.23	-0.45	0.12	0.25	0.81	0.77	0.24	0.15	1.00

AC, acres burned; BI, burning index; CN, fire counts; ER, energy release component; FWI, fire weather index; IC, ignition component; KB, Keetch-Byram drought index; P, precipitation; RH, relative humidity; T, temperature; SC, spread component; WSP, windspeed.

It is important to remember that computed BI values represent the near upper limit to be expected if a fire occurs in the worst fuel, weather and topography conditions for this fuel type.

4. Ignition component is a rating of the probability that a firebrand will cause a fire requiring suppression action. Because it is expressed as a probability, it is scaled from 0 to 100. An IC of 100 means that every firebrand will cause an 'actionable' fire if it contacts a receptive fuel. Likewise, an IC of 0 would mean that no firebrand would cause an actionable fire under those conditions. The IC is more than the probability that a fire will start (given a firebrand); it has to also have the potential to spread. Therefore the SC values are part of the calculation of the IC. Because the IC is a function of both the moisture content of small dead fuel and wind speed, it has significant daily fluctuations.
5. The Keetch-Byram drought index was not part of the original NFDRS (Deeming *et al.* 1972, 1977), but was added in a 1988 revision (Burgan 1988). It is a stand-alone index that can be used to measure the affects of seasonal drought on fire potential. The KB's relationship to fire danger is that, as the index value increases, the vegetation is subjected to increased stress due to moisture deficiency. At higher values desiccation occurs and live plant material is assumed to die, thus adding to the dead fuel loading on the site. Also, an increasing portion of the duff/litter layer becomes available fuel at higher index values.
6. Fire weather index was derived by Fosberg (1978) who assumed constant fuel (vegetation = grass) characteristics (see also Fujioka and Tsou 1985; Roads *et al.* 1991, 1997), based on equilibrium moisture content (a function of temperature and relative humidity) and wind speed. This index is not explicitly a part of the NFDRS and requires only instantaneous values of temperature, relative humidity and wind speed. Hence the FWI is most easily

applied in practice and was included here because it had been used previously to assess the potential for seasonal fire danger forecasting. It also provides a first look at global fire danger conditions.

Using the validation data described below, the anomaly correlations (monthly means removed) of these parameters with each other and a few meteorological parameters over the US West (land region west of 105°W) are shown in Table 2. The standard fire danger indices (IC, BI, ER, KB, SC) all are fairly well correlated, with the IC, BI and ER having the highest correlations. By contrast the ER and SC have little correlation and the KB has little correlation with any of the other indices. The FWI is also well correlated with all indices, except for the KB. All indices (except for the KB) indicate strong correlation (negative) with RH and, except for the KB and SC, significant positive correlations with temperature. Precipitation is weakly correlated with all indices but does show a strong positive correlation with RH, which again is more strongly related to the various indices (except for KB). In short, the ER, KB, SC and perhaps the FWI appear to be somewhat independent measures, whereas the other indices are strongly related to the ER, KB or SC. It should be noted that these correlations change little for weekly and seasonal means and for that matter, even when the entire US is taken into account. The correlations are also similar when forecasts rather than validation data are used. It should be noted that the data appear to follow a normal distribution for seasonal means but less so for weekly means and so it was somewhat surprising that the correlations among weekly means were comparable to correlations among monthly and seasonal means.

Validating analyses

For our main validation effort, we now use 1-day GSM/RSM forecasts, made every day from 00 UTC analysis initial conditions, as the basic meteorological variables for

calculating validating fire danger indices. There are only three missing 00 UTC initial states in more than a 2-year period and for these periods we used a previous 2-day forecast to generate the associated daily flux files. These 1-day forecasts are not exactly the same as the operational analysis, which is based on $4 \times$ daily 6-h forecasts from the latest high-resolution global model, or the reanalysis, which is based on $4 \times$ daily 6-h forecasts with the model we use for our forecasts, but they do form a useful approximation that can at least be used to estimate forecast skill upper bounds. As discussed by Roads *et al.* (2001a), global forecasts validated against the 1-day GSM/RSM forecasts are only slightly more skillful than forecasts validated against NCEP's operational analysis.

Although the 1-day forecast/analysis validation is certainly useful in the absence of other validating datasets, at least there are better approximations to the US precipitation from observations. For example, the National Climatic Data Center and first-order station network along with the precipitation network of the River Forecast Centers were utilized by Higgins *et al.* (2000) to develop gridded daily precipitation (25 km resolution), which we also use here for evaluation of the model GSM and RSM forecasts. In the comparisons below, we mostly use the 1-day RSM forecasts (observed precipitation is used in place of the 1-day RSM precipitation) to validate the RSM weekly to seasonal forecasts.

To summarize, the validation for these fire danger forecasts come from using forecast/observation analysis as input to the fire danger code. Fire danger indices are then initialized from this validation set but subsequently use forecasts after the initial day, including forecast precipitation, to drive the fire danger forecasts. As will be shown, these fire danger forecasts have substantial but regular biases from the validating fire danger analysis. It should be noted that one potential advantage for the NFDRS indices over the FWI indices is that they depend more on previously accumulated features. This means that we can use observed variables (like precipitation) up to the beginning of the forecast and then let the forecast model start from these somewhat realistic initial states. It would be even more useful to eventually include more observations into the validating and initializing fire danger dataset in much the same way current land data assimilation efforts are using available observations to define the surface hydrologic state. It should also be noted that a separate validation effort using actual station observations was carried out by Reinbold *et al.* (2005) and, although the skill was lower, there were many similarities, suggesting that our approximate validation procedure using 1-day forecasts is reasonable.

Fire statistics

While the forecasts need to first be validated with the initial state, which presumably is a better approximation to actual fire danger, we also need to compare the validation and forecast fire danger indices with characteristics of observed fires. Recently Westerling *et al.* (2001, 2002, 2003a, 2003b)

developed an extensive database of gridded (1-degree grid) monthly fire counts and acres burned from fire reports to several federal agencies for most of the western states (the spatial extent of the available fire statistics can be seen in Fig. 3g,h; see also Figs 4, 6). The statistics include only those fires that were reported and managed and hence are not likely to represent all fire counts. The statistics also include anthropogenic as well as natural fires. Despite their limitations, these fire statistics do at least provide a starting point for assessing a fire danger forecast analysis system. For convenience these fire statistics were interpolated to the model grid with the requirement that each model grid point have at least three surrounding grid points with information. The fire data were further interpolated to weekly means although, as shown later, this transformation did not then indicate that weekly forecasts had any additional skill. Following Westerling *et al.* (2002), the log of the acres burned was used, as this provided a more normal distribution. There is some indication that fire counts should also be similarly transformed but, because fire counts are not normally used to assess fire activity, we decided to defer this transformation for this paper. Because of our concern with the interpolation of these statistics to a higher resolution grid, ultimately when correlations were computed, all variables were averaged back to a 1-degree grid. This made the correlations only slightly larger, however.

Skill measures

As discussed by Roads *et al.* (2001a), skill can be measured in several different ways. We first of all compared the means of the forecasts with the validating analyses in order to make sure that the forecast model produced reasonable results. The forecasts tend to have small biases that are related to a tendency for the forecast model to drift away from the initial state towards the model climatology. Still, these biases are substantially larger than model anomalies, especially at seasonal time scale and we need to remove these systematic biases before assessing the skill of the model. The variability of the forecasts (standard deviation) is then compared in order to understand whether the forecast variability is similar to the validating analysis variability. Finally, correlations of the forecasts with the validating analyses as well as with the fire occurrence and size statistics provide us with our standard measure of skill. There are several other measures that can be used to describe skill but correlations do provide a basis for determining whether the skill is at least significant. Further work will be required to determine if this significant forecast skill is actually useful to a fire manager. Details about all of the skill measures used here can be found in Roads *et al.* (2001a).

Forecasts

Geographical variations

As shown in Fig. 3, summertime (ensemble mean of seasonal forecasts initialized in May, June, July) fire danger is certainly

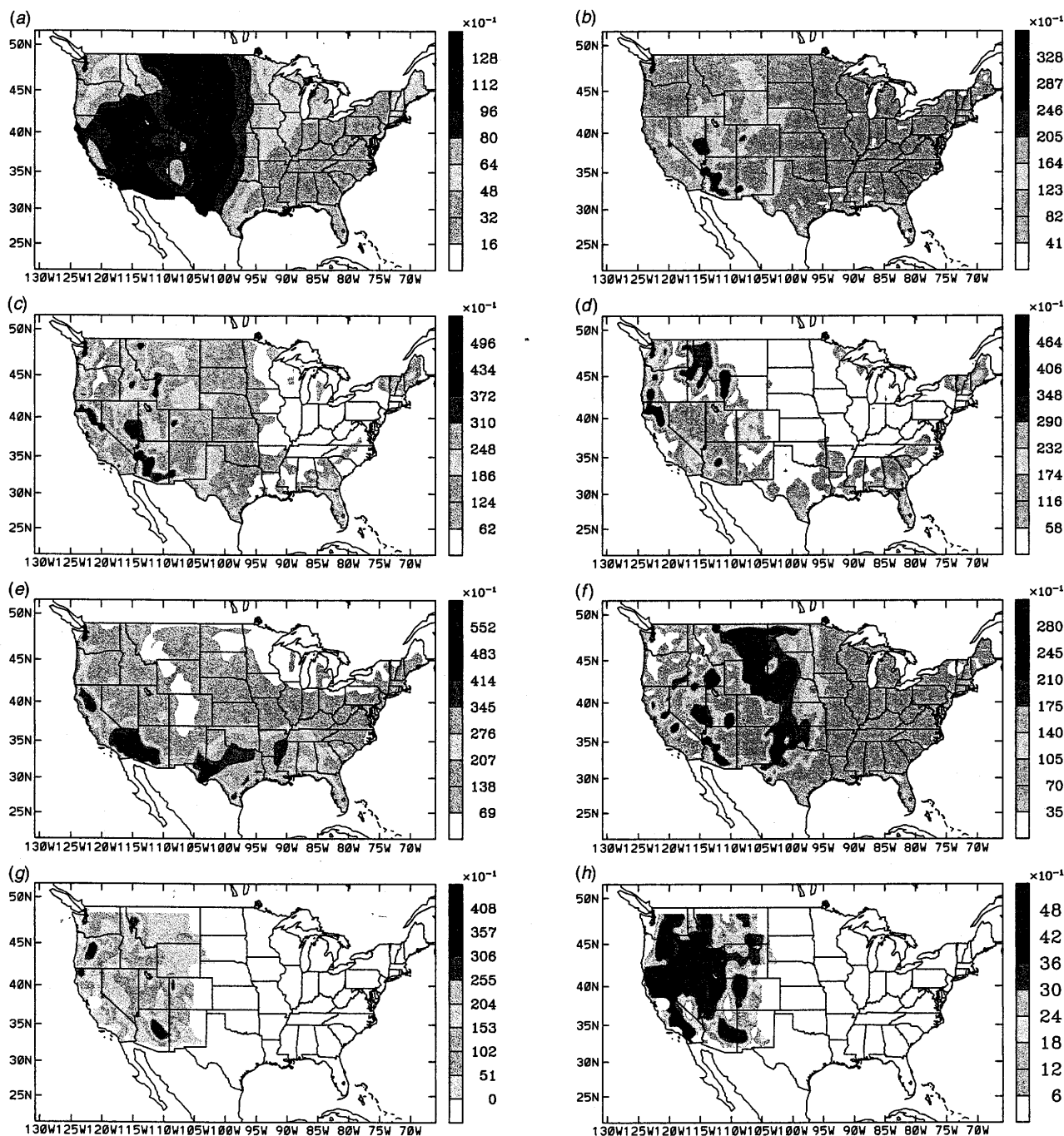


Fig. 3. Summer seasonal mean validation means: (a) fire weather index; (b) ignition component; (c) burning index; (d) energy release component; (e) Keetch-Byram drought index; (f) spread component; (g) fire counts; and (h) acres burned.

greatest in the US West, with the ER emphasizing the Northwest forests, the IC and BI emphasizing the dry Southwest, and the SC emphasizing the Rocky Mountain Front Ranges. The FWI is similar to the SC, in part because of the wind speed influence, which is strong east of the Rocky Mountains, and in part because the FWI assumes grass as the fuel model everywhere, which may result in greater fire spread. There is also some influence on the FWI by the relative dryness of the

Southwest. Again this dryness is also present in the IC, BI and KB although, as shown previously in Table 1, variations in the KB and FWI are not well correlated. Presumably the higher resolution features of the standard NFDRS indices (IC, BI, ER, SC) are related to the higher-resolution fuel characteristics embedded in these indices. In fact, the FWI can have remarkably high variations over the adjacent ocean (not shown), due to the higher wind speeds, which is another

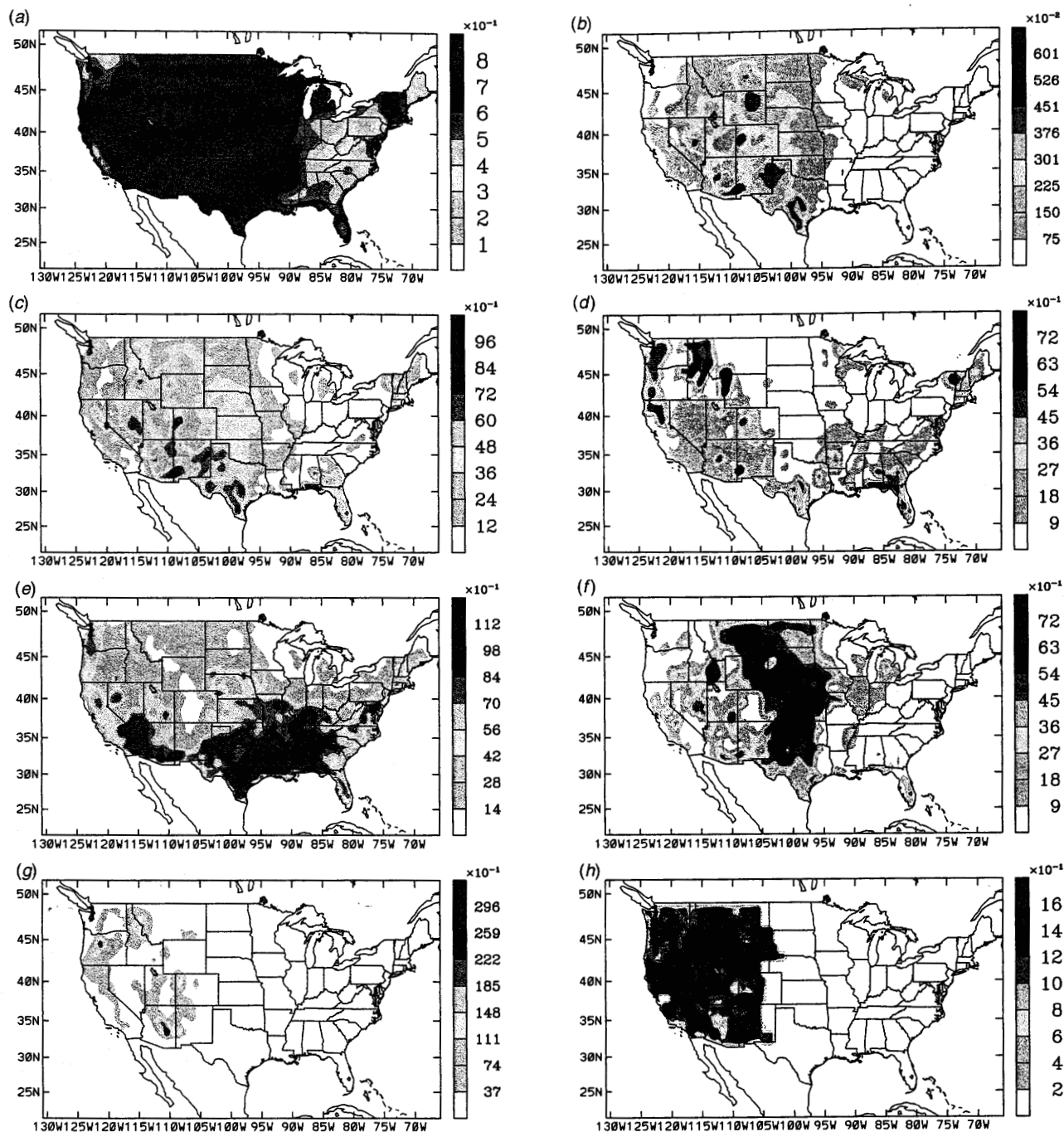


Fig. 4. Summer seasonal mean validation standard deviations: (a) fire weather index; (b) ignition component; (c) burning index; (d) energy release component; (e) Keetch-Byram drought index; (f) spread component; (g) fire counts; and (h) acres burned.

limitation associated with its use (a basic assumption of the FWI is an implicit grassland fuel model everywhere). Finally, as shown in Fig. 3g,h, the NFDRI indices do have an overall relationship with fire counts (CN) and acres burned (AC) although it is clearly not a one-to-one relationship. Fire counts tend to be especially large over Oregon, northern Idaho and Arizona, which are also related to the acres (log) burned.

The seasonal forecast has several seasonal biases (not shown). The SC forecasts tend to be too extreme in the region of Great Plains high wind speed, and this is also reflected in the FWI, which also has biases along the West coast. All indices have bias over Texas, which is presumably related to a bias in the forecast relative humidity (RH). For the most part the biases are positive but some negative biases do show up in

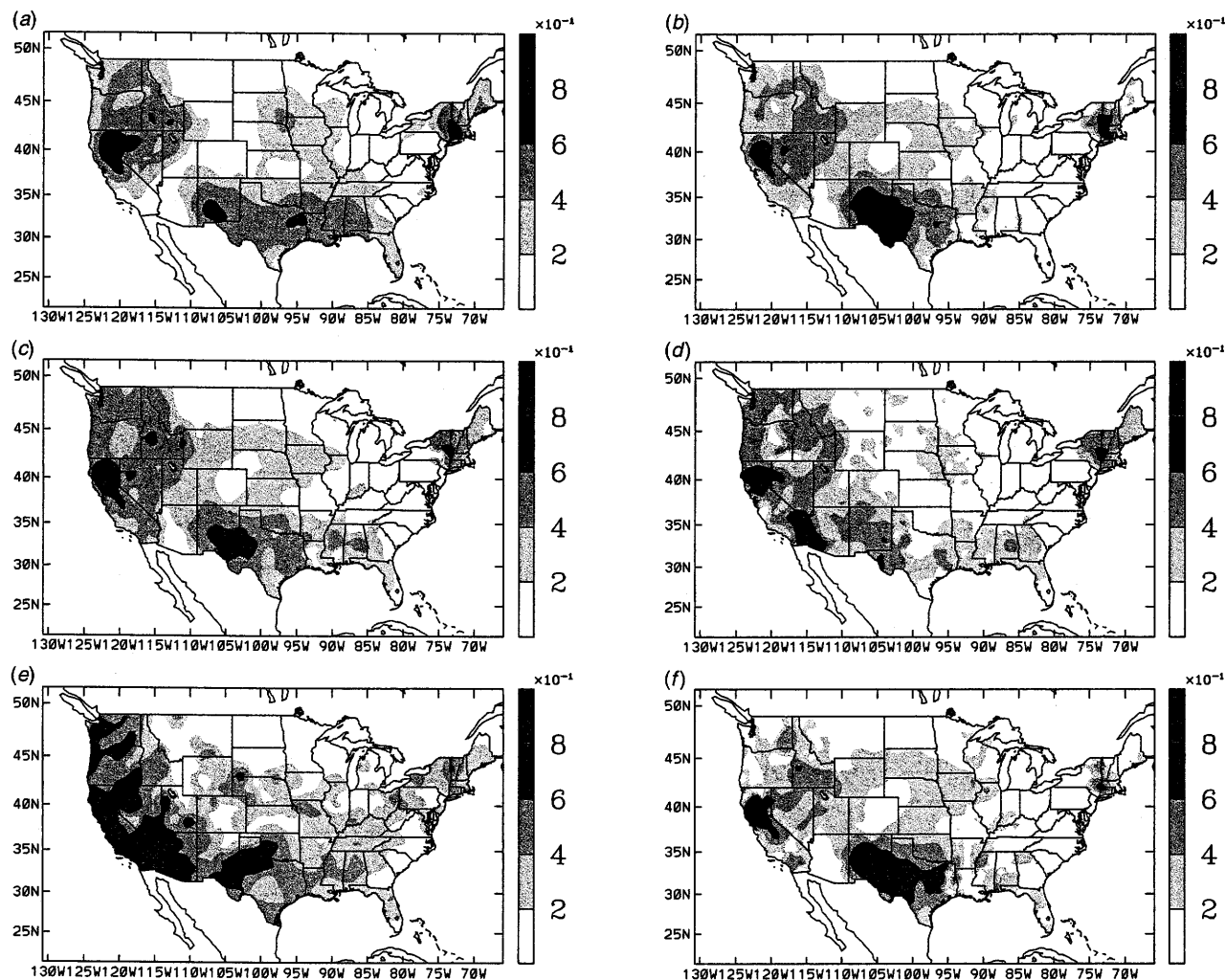


Fig. 5. Summer seasonal mean correlations between forecast and validation: (a) fire weather index; (b) ignition component; (c) burning index; (d) energy release component; (e) Keetch-Byram drought index; and (f) spread component.

the KB over the US West and East. The SC and FWI also have low biases over Montana and Wyoming. However, the biases are somewhat smaller than the climatological values and thus seasonal forecast climatologies have a high resemblance to the validating seasonal climatologies.

The variability as measured by the seasonal standard deviations (Fig. 4) reflects, in part, the summertime climatic means. For example, the SC and FWI s.d. values are strongest over the Central US; the ER s.d. is largest over the Northwest; the KB s.d. is strongest in the south; the BI and IC s.d. values are strongest in Arizona, New Mexico and Texas; the CN variability is strong in specific regions of Oregon, northern Idaho and Arizona. Interestingly, the regions of largest seasonal forecast discrepancies (not shown) occur in regions where weak s.d. occurs. For example, the forecast IC, ER, SC, BI and to some extent the FWI is much stronger over the US East. The forecast KB s.d. is especially strong over the

North Central. Only in a few places is the forecast s.d. weaker than the validating s.d.

This erroneous s.d. over the US East may help to explain why, at seasonal time scales, the US West is relatively better predicted (Fig. 5). Unfortunately, because the seasonal biases are greater than the standard deviations of seasonal averages, it is mandatory that some bias correction be made for the forecasts. Here, this bias correction was implicit as the correlations use only the anomalies from the forecasts or validations. Again, a separate climatology was developed for each forecast lag. Interestingly, the forecast correlations are not greatest where the means or s.d. values are greatest and may be more related to where the skill in making the forecasts of basic meteorological variables, like relative humidity, is greatest.

As shown in Fig. 6, the correlation of the validation NFDRS indices with AC is somewhat spotty but certainly

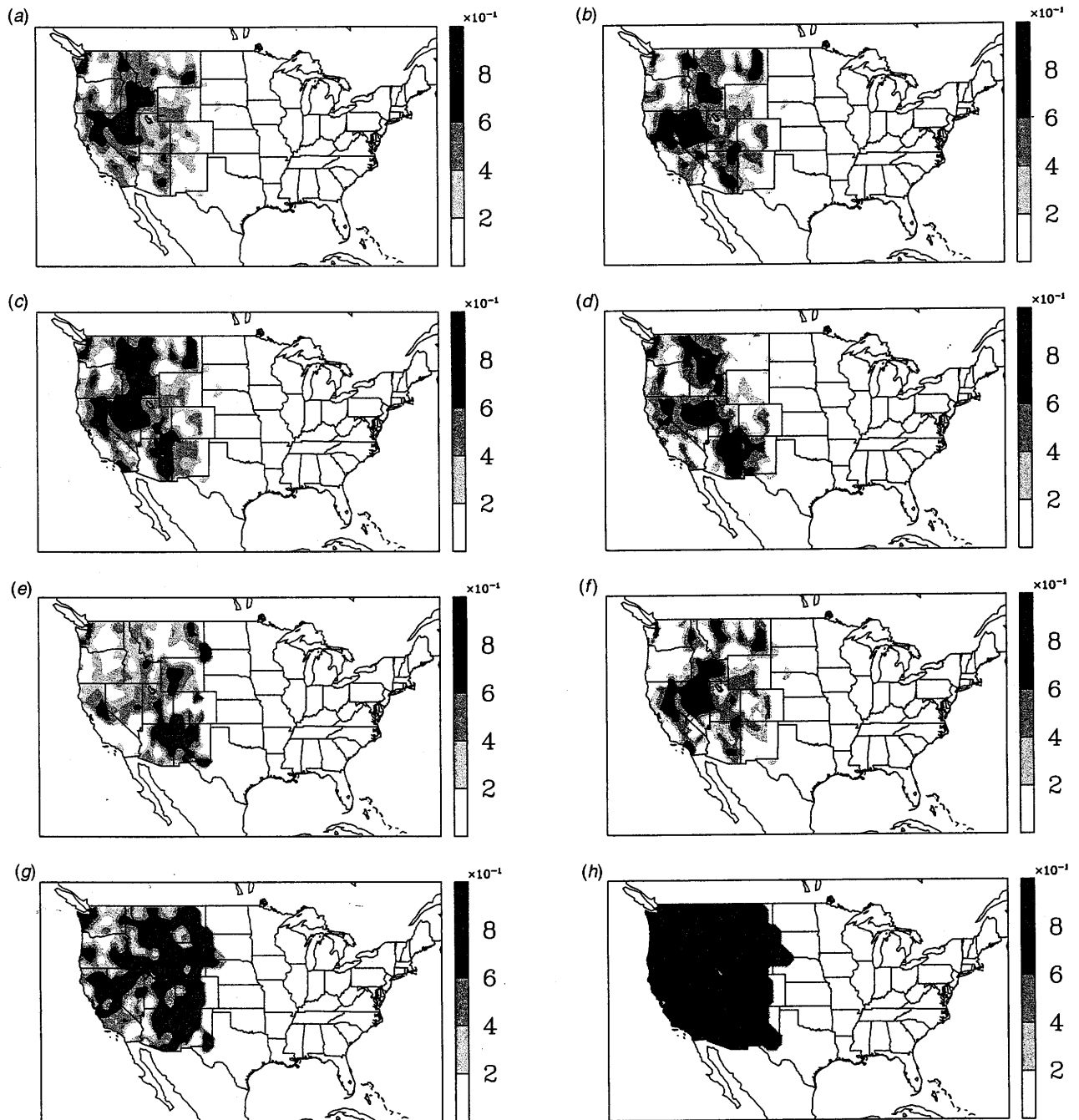


Fig. 6. Summer seasonal mean correlations between validation National Fire Danger Rating System indices, fire counts and acres burned with: (a) fire weather index; (b) ignition component; (c) burning index; (d) energy release component; (e) Keetch–Byram drought index; (f) spread component; (g) fire counts; and (h) acres burned.

significant in specific regions. For example, the FWI tends to be most useful over Nevada and Idaho, whereas the KB is more useful over the four corners area. These correlations are substantially reduced for the forecast correlations (Fig. 7), which have a tongue of high correlation over northern California, Nevada, Oregon and Idaho.

Temporal variations

Figure 8 shows the US West (US land region west of 105°W where fire statistics are available) average time series for the various NFDRS indices. Note that all forecast indices tend to be somewhat larger than the validating indices, especially

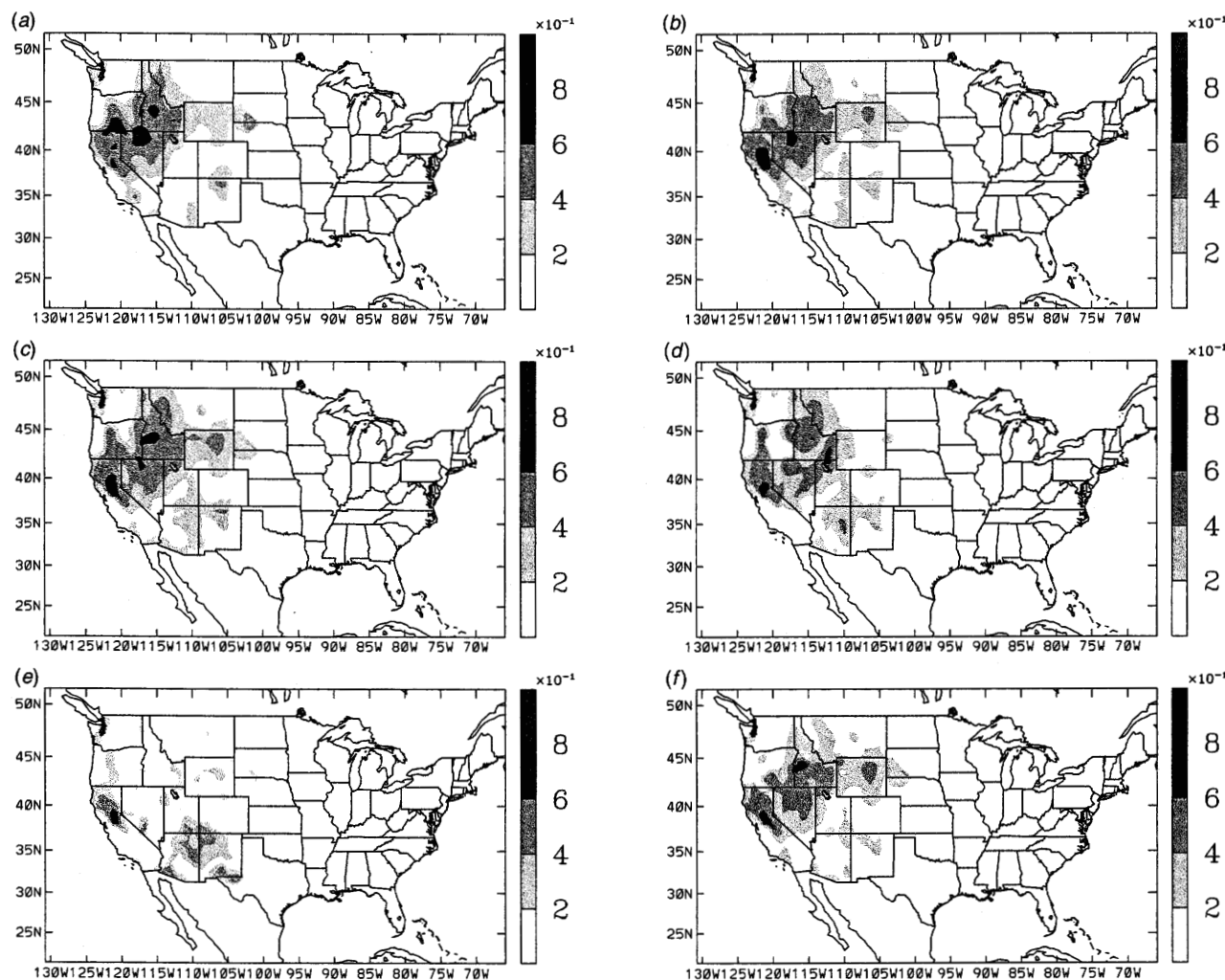


Fig. 7. Summer seasonal correlations with acres burned and forecast National Fire Danger Rating System indices: (a) fire weather index; (b) ignition component; (c) burning index; (d) energy release component; (e) Keetch–Byram drought index; and (f) spread component.

during the springtime. However, despite the bias, it is clear that both the seasonal forecasts and validating 1-day forecasts/observations have similar time series. The summer and early fall have the greatest potential for fire danger, with most indices peaking in late summer (August); the KB peaks even later in fall and early winter. There appears to be a particularly anomalous KB during the fall and winter (2001/2002), which does not show up in weekly or even monthly time series. It is not clear why this KB anomaly shows up in these forecasts, given that the precipitation forecasts do not appear overly anomalous (Roads 2004a) and the other forecasts (Roads 2004b) do not appear similarly anomalous. However, it should be noted that there does appear to be an increased bias in all components beginning in July 2001, which may be related to when NCEP changed its global analysis system, including an increase in resolution. All of the indices do have the same seasonal variation as the fire statistics (CN and AC), with

the largest values occurring during the summer. The validation NFDRS indices also have the same overall variability with the largest values occurring during the summer of 2000. Unfortunately, the forecast variability does not really capture this as forecast fire danger remained high during the summer of 2001.

As shown in Fig. 9, both the seasonal forecasts and validations have similar time series of variability (s.d.). The summer and early fall have the greatest potential for fire danger, with most indices peaking in late summer (August); the KB peaks even later in fall and early winter. It should be noted, however, that the FWI does not have a clear seasonal variation, presumably because of the counteracting effects of decreased summertime winds and decreased relative humidity and increased wintertime winds and relative humidity.

The correlations of the forecast indices with the validating indices (Fig. 10) show the large differences between

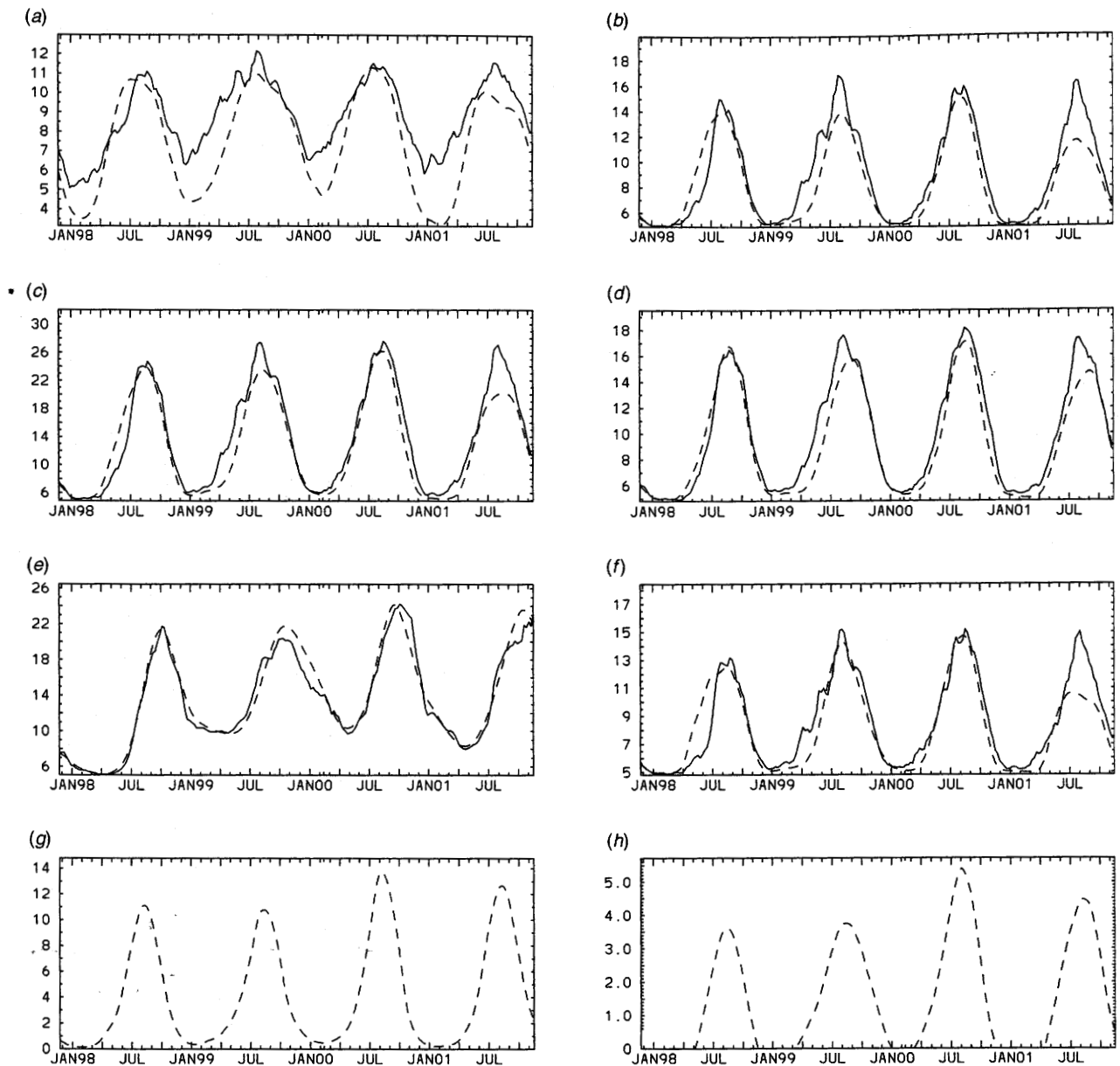


Fig. 8. Seasonal forecast (solid lines) and validation (dashed lines) US West means: (a) fire weather index; (b) ignition component; (c) burning index; (d) energy release component; (e) Keetch-Byram drought index; (f) spread component; (g) fire counts; and (h) acres burned.

the weekly forecasts and the seasonal forecasts. Correlations change from around 0.7 for the weekly forecasts to 0.2 for the seasonal forecasts. The time of minimum skill appears to be in early summer. Late summer and fall are relatively better predicted, except for the KB, which is predicted best during the winter and spring. There is also a weak indication of a forecast skill decline in the fall of 2001, when an especially anomalous KB was noted previously.

The correlations of the seasonal validating and forecast NFDRS indices with AC (Fig. 11) is fairly low although there is at least a tendency for the validating NFDRS indices to have

somewhat higher skill than the forecast NFDRS indices, as indicated previously in Figs 6 and 7. There is also a tendency for the wintertime forecast skill to be somewhat greater, presumably because of the increased difficulty in describing and forecasting summertime precipitation as well as other relevant meteorological variables. The interannual variability is especially large though. Note for example the increased skill in the winter and spring of 1998, followed by a slow decline until the summer of 2000 when increased skill again occurred. This may indicate that it is easier to forecast extreme events.

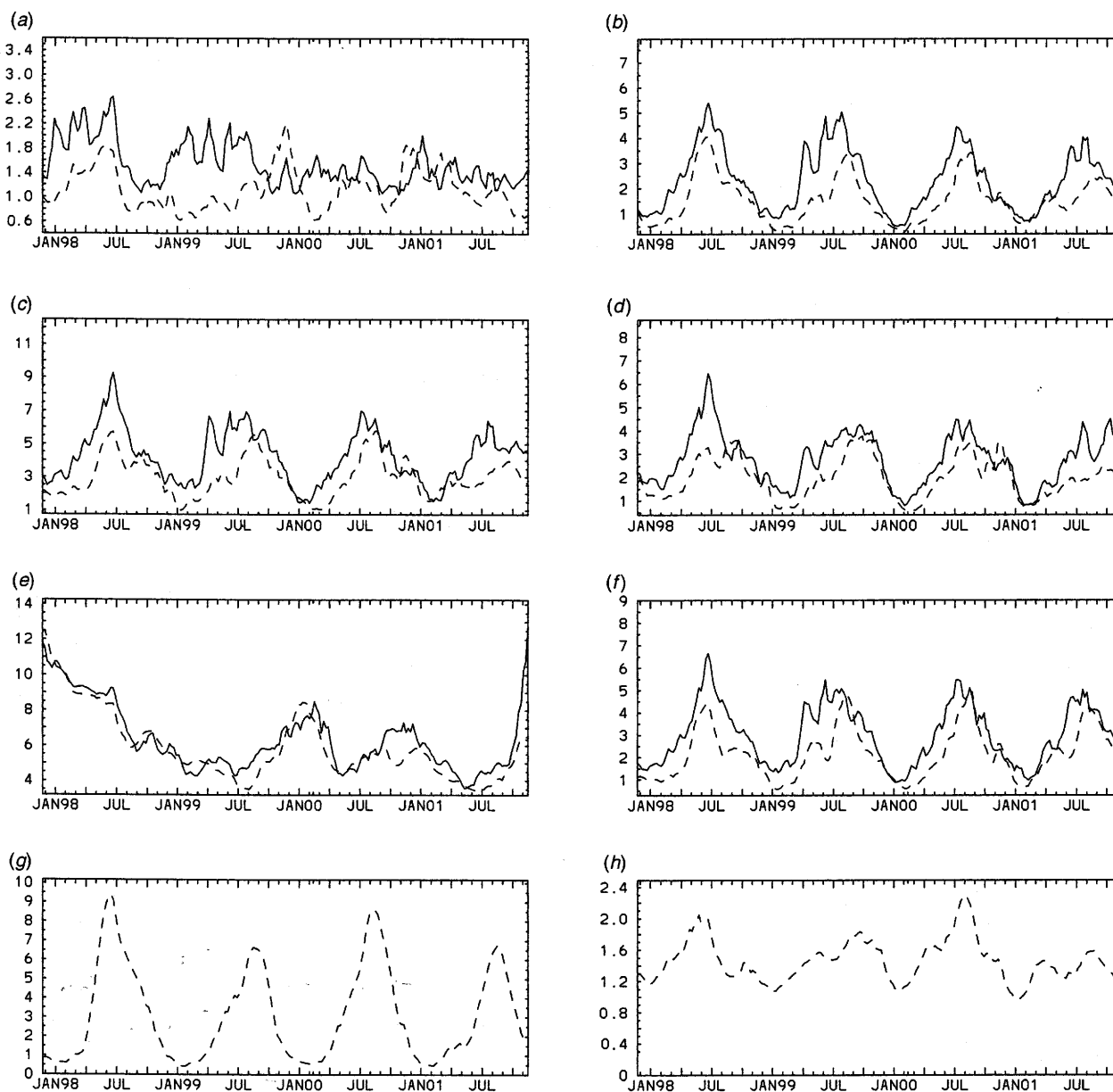


Fig. 9. Seasonal forecast (solid lines) and validation (dashed lines) monthly US West standard deviations: (a) fire weather index; (b) ignition component; (c) burning index; (d) energy release component; (e) Keetch-Byram drought index; (f) spread component; (g) fire counts; and (h) acres burned.

Forecast lead relations

Figure 12 shows the forecast lead relationships for the climate means. Basically, the bias in the indices is a linear function of forecast time, which means that the monthly and seasonal biases can be approximated by the bias at the midpoint of the month (2 weeks for the first month, 6 weeks for the first season). The forecast standard deviations (not shown) are also a function of forecast lag and averaging period, with the smallest variations occurring for seasonal means.

The correlations of the forecast with validations (Fig. 13) show the strong forecast skill at 1 week, followed by a sharp

decline towards zero. However, note that the seasonal and monthly correlations tend to be somewhat larger than the weekly correlations and these correlations may be significant. The greatest skill occurs in predicting the KB, followed by the BI and ER and then the IC and SC. The FWI skill is comparable to the SC, in part because of the stronger dependence of these indices on wind speed. Persistence forecasts are also shown and, while these forecasts are somewhat inferior to the dynamical forecasts, they do indicate that indices with longer persistence are better predicted at seasonal time scales.

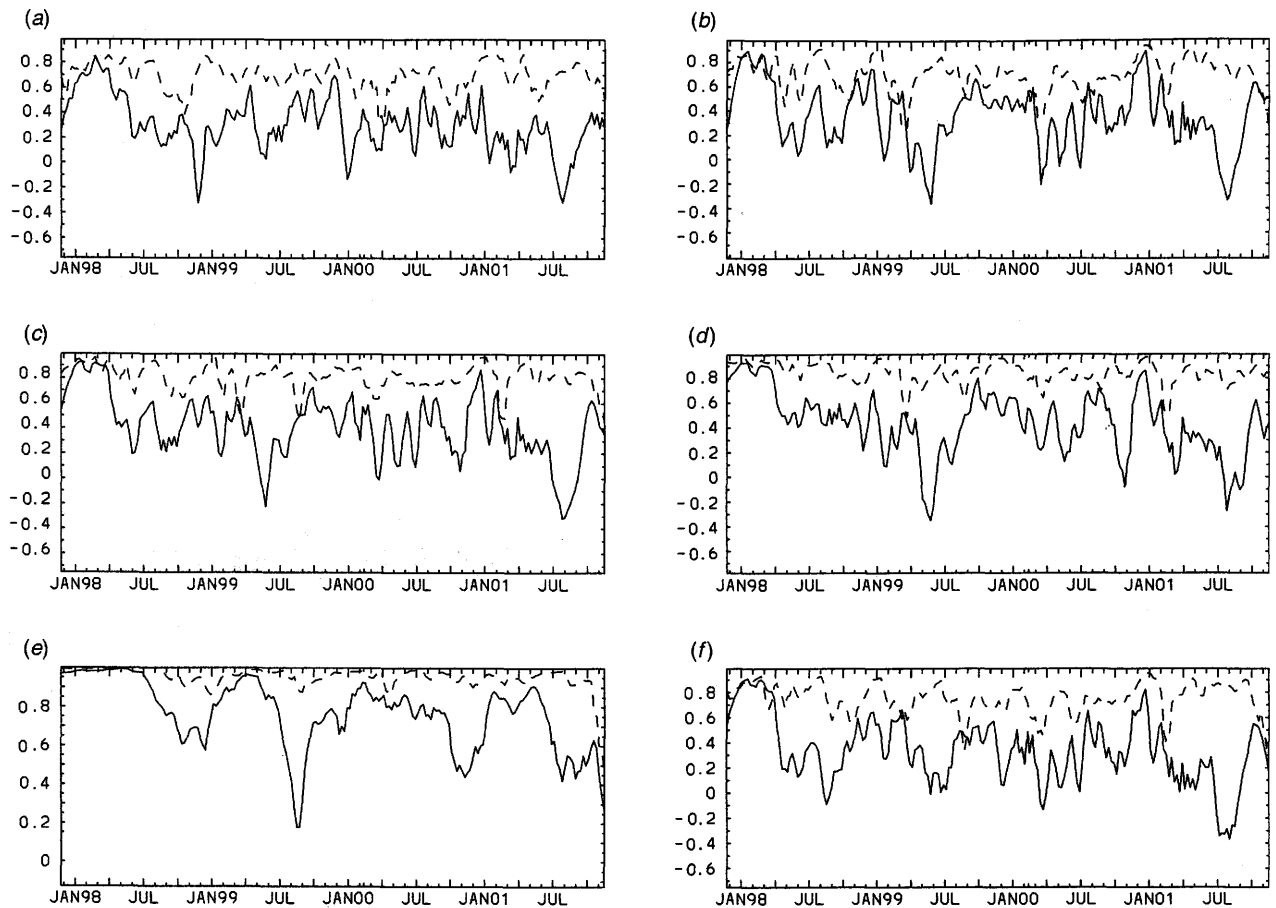


Fig. 10. Seasonal forecast (solid lines) and weekly forecast (dashed lines) US West correlations: (a) fire weather index; (b) ignition component; (c) burning index; (d) energy release component; (e) Keetch-Byram drought index; and (f) spread component.

Despite the significant skill in forecasting the actual fire danger indices (which is summarized in Fig. 14), there is only a weak lead relationship with the forecast and validating NFDRS indices and AC. The strongest relationships occur for seasonal averages. This may be related in part to the use of monthly fire statistics. In fact, as is also shown in Fig. 14, there is little difference between using the validating analysis and the seasonal forecast for these NFDRS correlations with fire extent (AC). Despite this low skill there is an indication that the indices are at least somewhat related to fire occurrence and surprisingly that the simplified FWI is the overall best (but really indistinguishable from the other indices). Finally, we should mention that we believe there is much work that is still needed to develop better indices and better measures of fire extent and occurrence, as well as better analysis and forecast models.

Summary

The goal of this paper was to determine the skill of a dynamical model in forecasting seasonal fire danger ratings. During the past decade, seasonal forecasts have certainly become

more commonplace, although making an explicit connection to the fire danger community has been typically lax. Instead it is commonly assumed that forecasts of standard monthly mean temperature and precipitation should be made and that applications communities would somehow adapt to using these monthly means. Here we have shown that it is quite possible to make the forecasts compatible with what the fire applications community needs: daily time series of a few critical variables (e.g. temperature, relative humidity, wind speed, precipitation) in order to drive fire danger models.

As was shown, there was probably significant seasonal forecast skill for all of the standard fire danger ratings (IC, BI, ER, KB, SC) as well as the FWI, which had previously been used to assess fire danger forecast skill. Persistence forecasts were also shown and, while these forecasts are somewhat inferior to the dynamical forecasts, they did indicate that indices with longer persistence were better predicted at seasonal time scales. These seasonal fire danger forecasts had significant skill even for 1-month forecast leads (12 week averages of weeks 5–16). It would be of interest to determine the forecast skill for one season lead (e.g. out to 7 months),

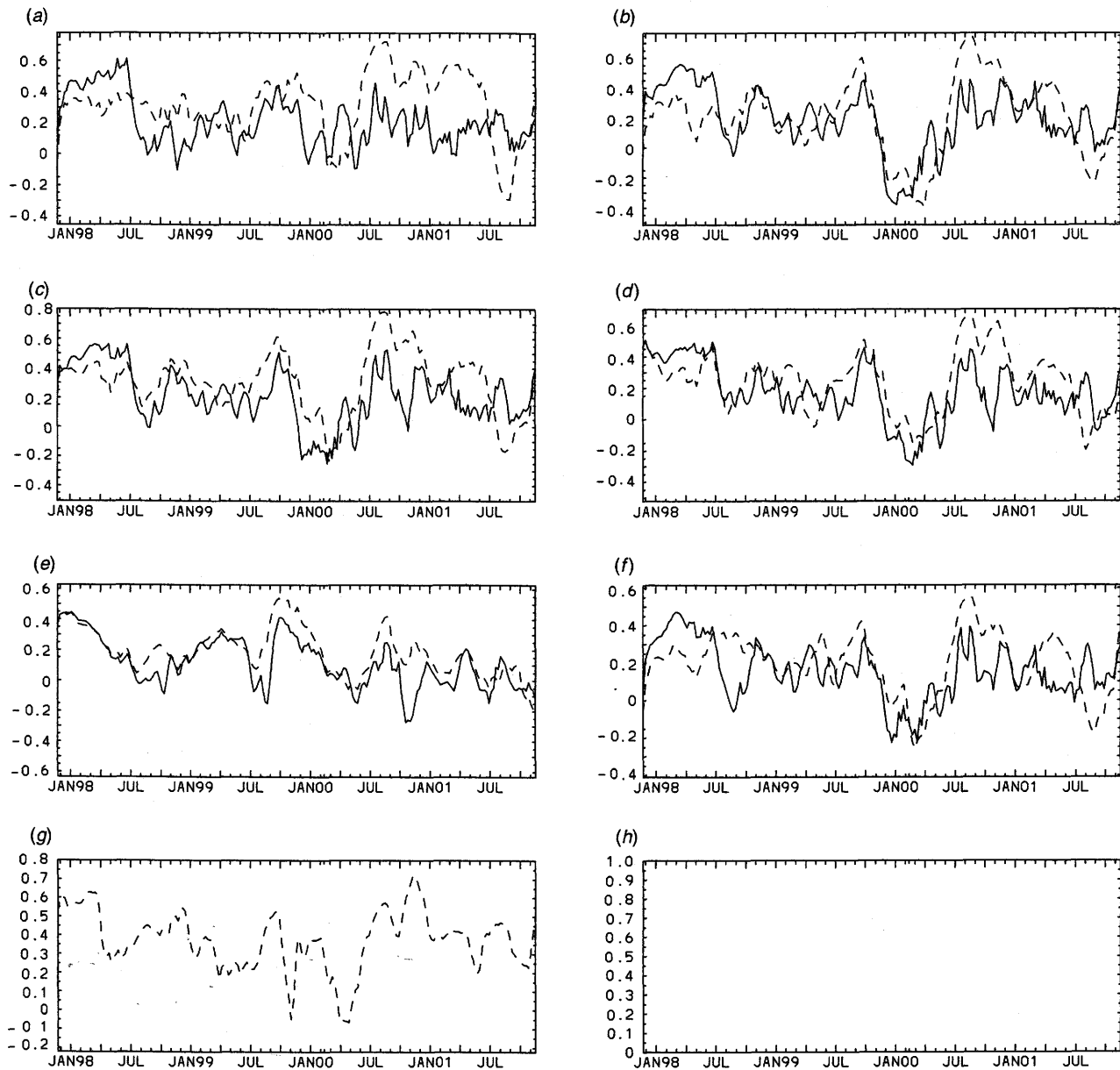


Fig. 11. Summer seasonal mean correlations between forecast (solid lines) and validation (dashed lines) National Fire Danger Rating System indices and acres burned. (a) Fire weather index; (b) ignition component; (c) burning index; (d) energy release component; (e) Keetch-Byram drought index; (f) spread component; (g) fire counts; and (h) acres burned.

although this would mean a substantial increase in computer time and storage (almost double) in order to produce these ultra long-range forecasts.

It was further shown that the fire danger indices are related to fire statistics such as fire counts and acres burned, especially when the validating was used rather than forecast output. Still, the relationships are weak and further improvements should be possible. In fact, fire danger indices are now evolving towards making use of more remotely sensed vegetation characteristics instead of trying to parameterize these

relationships from complex averaging of meteorological input. In this regard the experimental fire potential index (FPI), first investigated in 1996 at the USFS Intermountain Fire Sciences Laboratory and then refined in collaboration with the US Geological Survey EROS Data Center (EDC), is a major augmentation to standard USFS fire danger indices (Burgan *et al.* 1998). The FPI model combines Relative Greenness derived from the normalized difference vegetation index (NDVI) data derived from NOAA AVHRR data to generate 1-km resolution fire potential maps on a daily basis.

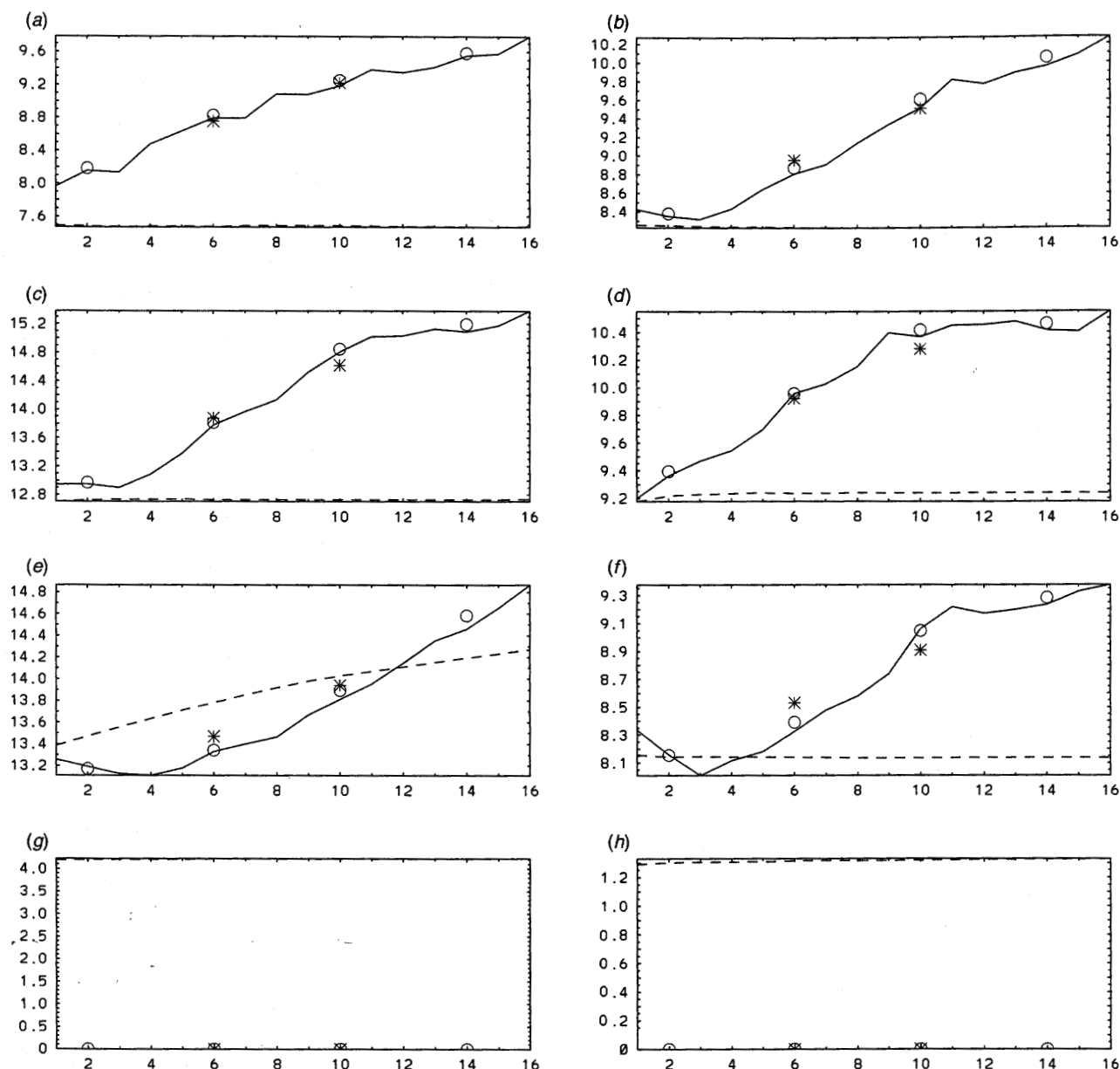


Fig. 12. Forecast weekly means (solid lines), validation weekly means (dashed lines), forecast monthly means (○) and seasonal means (*) as a function of forecast lead-time. Monthly and seasonal forecast values are plotted at the centre points (i.e. week 2 for month 1 and week 6 for first season). (a) Fire weather index; (b) ignition component; (c) burning index; (d) energy release component; (e) Keetch-Byram drought index; (f) spread component; (g) fire counts; and (h) acres burned.

One potential advantage for this index is that fire danger could potentially be assessed for other regions, where good fuel models do not currently exist.

Finally, although NFDRS indices are widely used by the USFS and other agencies (NIFC) to guide decisions involving fire danger, it should be emphasized that the fire danger rating of an area is only one, imperfect, tool to assess 'fire business' decisions. The emphasis is on imperfect because fire danger rating information is not a final answer by itself; it must be considered along with the manager's local knowledge of the

area and consequences of a decision when arriving at the best solution for a particular problem. Given the current low forecast skill, seasoned experienced judgement remains a critical aspect of fire danger forecasts.

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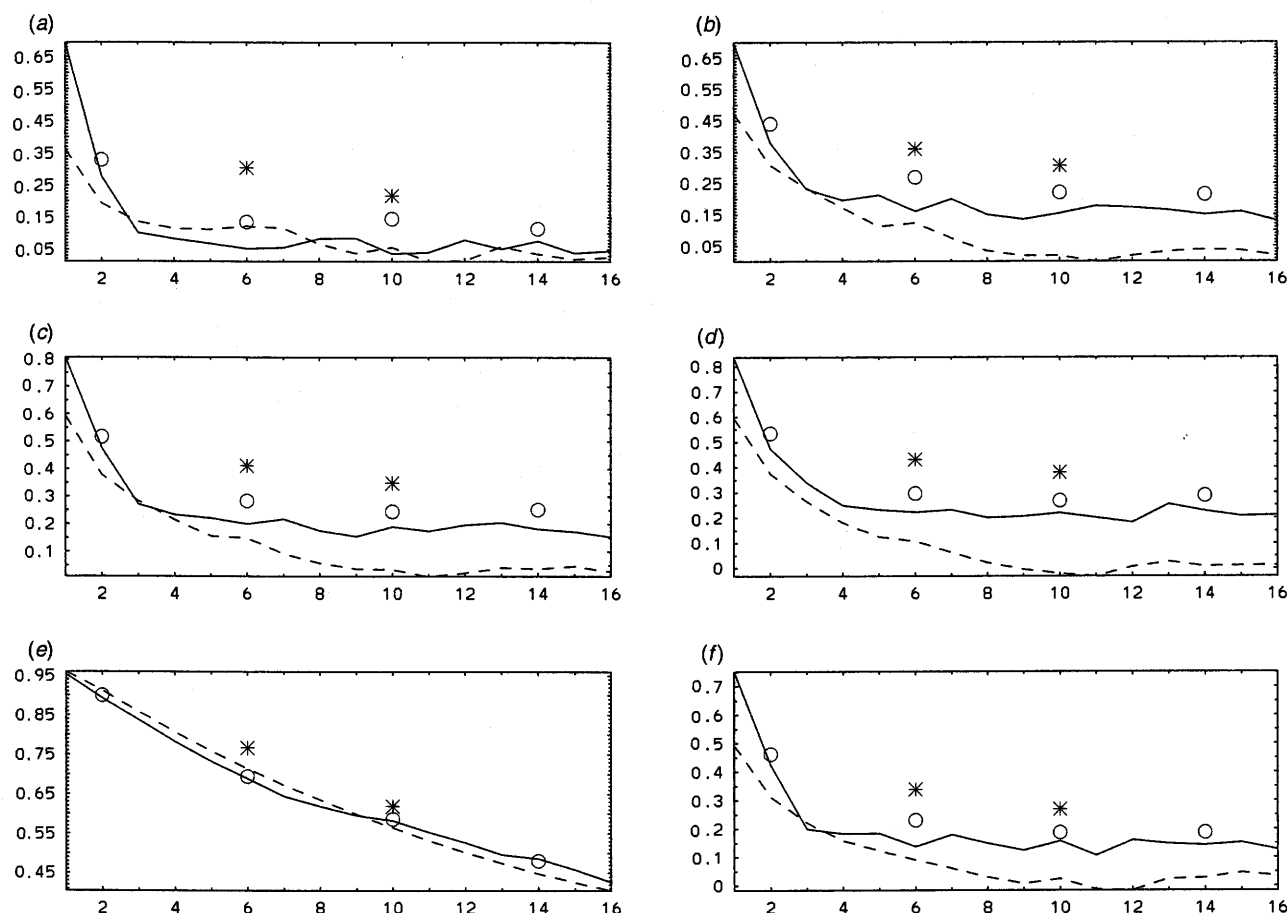


Fig. 13. Forecast weekly (solid lines), persistence weekly (dashed lines), forecast monthly (○) and seasonal (*) correlations as a function of forecast lead-time. Monthly and seasonal values are plotted at the centre points (i.e. week 2 for month 1 and week 6 for first season). (a) Fire weather index; (b) ignition component; (c) burning index; (d) energy release component; (e) Keetch-Byram drought index; and (f) spread component.

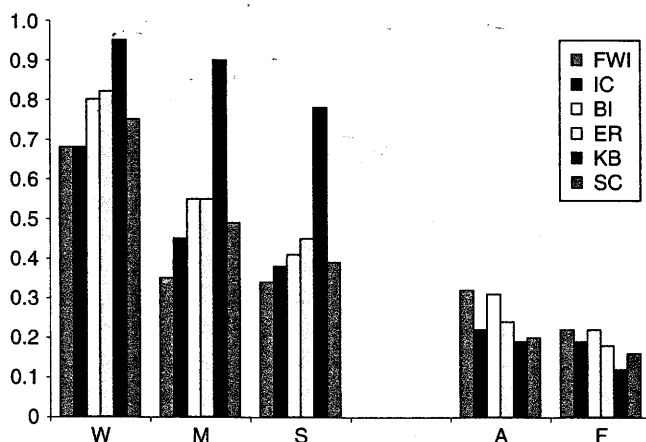


Fig. 14. Summary forecast skill for weekly (W), monthly (M), seasonal (S) forecasts of fire weather index (FWI), ignition component (IC), burning index (BI), energy release component (ER), Keetch-Byram drought index (KB) and spread component (SC) in comparison to the validating analysis (A) and seasonal forecasts (F) for these same indices with the seasonal mean acres burned.

are those of the authors and do not necessarily reflect the views of NOAA or USFS. We are grateful to A. Westerling for providing us with his fire statistics.

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Evaluation of the Experimental Climate Prediction Center's fire danger forecasts with remote automated weather station observations

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Abstract. The Scripps Experimental Climate Prediction Center has been routinely making regional forecasts of atmospheric elements and fire danger indices since 27 September 1997. This study evaluates these forecasts using selected remote automated weather station observations over the western USA. Bias and anomaly correlations are computed for daily 2-m maximum, minimum, average temperature, 2-m maximum, minimum and average relative humidity, precipitation and afternoon 10-m wind speed, and four National Fire Danger Rating System indices—ignition component, spread component, burning index and energy release component. Of the atmospheric elements, temperature generally correlates well, but relative humidity, precipitation and wind speed are less correlated. Fire danger indices have much lower correlations, but do show useful spatial structure in some areas such as Southern California, Arizona and Nevada.

Introduction

The Experimental Climate Prediction Center (ECPC) at the Scripps Institution of Oceanography has been routinely making global spectral model (GSM) and regional spectral model (RSM) forecasts since 27 September 1997 (Roads 2004; Roads *et al.* 2005). These models produce global to regional forecasts of common atmospheric elements including: temperature, relative humidity (RH) and precipitation. In addition to these weather elements, the RSM also calculates fire danger indices for the continental United States. Forecasts are generated once daily for 7 days and once weekly for 16 weeks, and are made available (images and digital output) via the ECPC website (http://ecpc.ucsd.edu/m2s/m2s_ECPC_forecasts.html). The forecasts are directly relevant to fire weather meteorologists, fire and fuel specialists and fire management. In fact, it should be noted that ECPC originally developed these forecasts in collaboration with the United States Department of Agriculture Forest Service (USFS) in order that they might eventually have access to automatic seasonal forecasts that would be useful in making strategic resource (e.g. suppression equipment, personnel) decisions (Roads 2004).

The skill of the meteorological forecasts for the parent GSM (Roads *et al.* 2001), RSM precipitation (Roads 2004) and fire danger indices calculated from the RSM (Roads

et al. 2005) have been examined previously in comparison to the initial state of the model (as well as observed precipitation). The initial state for the model comes initially from the National Centers for Environmental Prediction's (NCEP) global data assimilation system, although the global analysis is then subsequently modified by the RSM 1-day forecasts used for the validation (observed precipitation was used in place of RSM precipitation for these validations).

As shown by Roads (2004), when compared to these validating observations, the US RSM precipitation forecasts had a week-1 anomaly correlation of 0.4 overall, and an anomaly correlation of 0.13 overall for the seasonal forecasts. As further discussed by Roads *et al.* (2005), the National Fire Danger Rating System (NFDRS) burning index and energy release component seasonal forecasts showed the greatest skill when compared to validating observations, with overall week-1 forecast anomaly correlations of 0.8 and seasonal forecast correlations closer to 0.4. The spread component and ignition component indices had slightly lower overall week-1 and seasonal forecast correlations of approximately 0.7 and 0.35 respectively. Despite lower overall seasonal forecast correlations, all NFDRS indices had high seasonal skill over many locations in the western United States (Roads *et al.* 2005). A related index, the fire weather index (FWI), has been assessed (Roads *et al.* 1991, 1997, 2001) and was also found

to be predictable at weekly and seasonal time scales. The importance of the FWI is that it can also be forecast globally and thus provides a fire danger indicator for regions that do not utilize a formal fire danger rating system.

The down-scaled global analysis used previously to evaluate these forecasts may not be representative in many regions of interest to the fire management community, as the analysis is based mainly on airport and urban stations. Also, it is somewhat disquieting to compare the forecast model with itself, even though the forecasts (weeks to seasons) are reasonably far from the initial day used for the validation. Our objective, therefore, was to further assess the accuracy of the ECPC weekly, monthly and seasonal RSM forecasts in comparison to actual observations from remote automated weather stations (RAWS). Our emphasis was placed on the elements most related to fire danger (e.g. maximum, minimum and average daily surface temperature, maximum, minimum and average daily surface relative humidity, surface wind speed, precipitation amount and precipitation duration) and an evaluation of National Fire Danger Rating System (NFDRS) indices, including: burning index (BI), spread component (SC), energy release component (ERC) and ignition component (IC).

This study provides a further assessment of the operational utility of RSM for fire climate forecasts. Land managers and fire weather meteorologists will, as a result of this analysis, be able to identify the spatial regions, forecast periods and time of year that the model is most skillful for, for operational and planning purposes. Another goal of this study was to quantitatively identify strengths and weaknesses of the model to increase operational utilization of the forecasts within the US Interagency Predictive Services. For research meteorologists, this analysis will provide a further performance measure that will be helpful in the fine-tuning and development of the model itself.

The next section gives a more detailed examination of the history and operation of the RSM, including a summary of some of the previous skill tests performed on the model. Then we describe the origin and details of the RAWS, ECPC validation and ECPC forecast datasets. The following section outlines the methodology used in preparing and analyzing the data. We then present the analysis in terms of temporal and spatial variations. Finally we discuss and summarize the results, including the potential value of the seasonal forecasts.

Experimental Climate Prediction Center's global and regional models

The Experimental Climate Prediction Center's GSM, based on the model used for the NCEP/NCAR (National Center for Atmospheric Research) reanalysis (Kalnay *et al.* 1996; see also Roads *et al.* 1999), was the parent model of the RSM (Juang and Kanamitsu 1994; Juang *et al.* 1997) which, like the GSM, is based on the primitive equations and uses similar physical parameterizations. The RSM is used to enhance

regional climatological features (e.g. Chen *et al.* 1999, 2001; Takle *et al.* 1999; Anderson *et al.* 2000, 2001; Roads and Chen 2000; Anderson and Roads 2002; Roads *et al.* 2005; Han and Roads 2004) although boundary conditions must either be provided by a global analysis or by the GSM (or other global model) forecasts.

The GSM is run routinely once a day for 7 days at ECPC starting from NCEP's 00 UTC Global Data Assimilation System, and once every weekend (00 UTC on Saturday) the 7-day forecast is extended to 16 weeks. These global forecasts have skill as shown by Roads *et al.* (2001), Roads and Brenner (2002) and Chen *et al.* (2001) for global, US, European, Middle East and Asian regions. The GSM also provides the needed four-times-daily boundary conditions for the RSM forecasts that are run in parallel at varying degrees of increased resolution over different regions including the USA, south-western USA, California and Brazil. Evaluations of the RSM forecast skill over the USA at 60 km resolution have been described previously (Roads 2004; Roads *et al.* 2005). The US RSM forecast skill is at least as good as the GSM forecast skill and it is expected that model developments under way will eventually show more clearly the added advantage of the increased resolution. Given the equivalent skill of the RSM with the GSM, there has been an increasing emphasis on using the RSM for USA and other regional short- and long-range predictions. For this study only the RSM output was examined.

Data

Three primary datasets were used in this study: RSM forecasts, RSM validation (including observed precipitation) and RAWS observations. The spatial domain for all datasets covers the western United States and includes approximately 100°W to ~125°W longitude, and 30°N to 50°N latitude. All datasets were inclusive of the period from 27 September 1997 through 28 December 2002, which were the first 5 years of ECPC forecasts.

RSM forecast dataset

Every weekend, ECPC makes a 16-week RSM forecast for surface temperature (daily minimum, maximum and mean), relative humidity (daily minimum, maximum and mean), precipitation amount, hours of precipitation, mean afternoon wind speed (defined as 1200 to 1800 h local time), IC, SC, ERC and BI. Model output consists of daily values that are averaged to create weekly means that can then be combined into monthly (4-week) and seasonal (12-week) means. Each extended forecast examined (275 total from 27 September 1997 through 28 December 2002) is evaluated in the form of twelve 1-week means, three 1-month means and one seasonal mean (only the first 12 weeks of each 16-week forecast was examined in this study, although, as shown in Roads *et al.* (2005), there is little difference between the 1–12-week and

5–16-week forecasts). These forecasts have a spatial resolution of approximately 0.6 degrees (~60 km) and comprise 58×97 grid nodes (covering the entire contiguous USA), where each node represents the spatial position of the forecast output elements. Because the majority of RAWS are located in the western USA, only the 43×39 grid nodes over the West were used in this study.

ECPC RSM validation dataset

Every day ECPC makes 1-day RSM forecasts based on the 00 UTC NCEP operational analysis initial conditions. These are referred to as ECPC validation forecasts. The validation forecasts are not identical to the operational analyses, which are based on 6-h forecasts made four times daily and use the latest high-resolution global model, but can be considered a useful approximation (Roads 2004; Roads *et al.* 2005). As with the extended forecasts, the validation forecasts include all 13 atmospheric elements used in this study and an identical spatial grid. They are archived as weekly averages of daily values and essentially function as validating observations. There were 286 weekly averages of validation data used in this study, including the first 11 weeks of data from 2003. Acquisition of the 2003 data was necessary to evaluate the extended forecasts made in late 2002. It should also be noted that, because precipitation observations are readily available (Higgins *et al.* 2000), these were used in place of the RSM 1-day precipitation forecasts. These precipitation observations are originally taken from thousands of individual observations, which are gridded with the additional help of radar observations to distinguish between missing and zero values.

RAWS

Within the USA, land and fire management agencies maintain the RAWS observational network for fire weather related measurements (see <http://www.fs.fed.us/raws>). These stations are typically located in wilderness, forest and rangeland areas where it is desired to monitor fire danger. The hourly observations are transmitted to the National Interagency Fire Center (NIFC) via satellite. These data are forwarded to the Weather Information Management System (WIMS) for agency use, distribution and storage, and also via direct link to the Western Regional Climate Center (WRCC) for historical archiving. The RAWS data used in this study were obtained from WRCC. These data are not yet integrated into the RSM model analyses and thus provide a somewhat independent assessment of the analyses and the forecasts evaluated here.

From this dataset, RAWS surface measurements of daily minimum temperature, daily maximum temperature, daily mean temperature, wind speed, daily minimum relative humidity, daily maximum relative humidity, daily mean relative humidity, precipitation and hours of precipitation were extracted. The WIMS provides site descriptors (fuel model, slope, climate class, etc.) for each RAWS which, combined

with the above atmospheric elements, allows for the calculation of relevant fire danger indices (BI, IC, ERC, SC) used in this study.

There were 262 RAWS sites in the western USA that had sufficient quality data for the years 1997–2003. Sites were chosen based on three criteria regarding completeness of the dataset:

- (1) No more than 2 months of missing data in any year;
- (2) Availability of year-round operational measurements; and
- (3) Availability of historical data for the period September 1997 through March 2003.

Measurements that were clearly in error (e.g. relative humidity over 100%, negative wind speed) were considered missing and excluded from the analysis. There were 477 RAWS with data histories from September 1997 through March 2003 that were missing weeks, months or years of data, mostly due to instrument error or seasonal operation, and therefore were not used in this study.

NFDRS

The four fire danger indices output by NFDRS that are examined in this study are the burning index (BI), the ignition component (IC), the energy release component (ERC) and the spread component (SC), discussed in technical documentation provided by Bradshaw *et al.* (1983), Deeming *et al.* (1972, 1977) and Burgan (1988). The atmospheric and environmental elements used in the calculation of NFDRS indices include temperature, relative humidity, precipitation, wind speed, cloud cover, fuel type and slope. The NFDRS indices are traditionally computed for point locations (such as RAWS), the conditions of which are considered representative of the climate, fuels and slope of a pre-defined area. However, computations of NFDRS indices from RSM output were performed using gridded weather, fuels and topography data. The fuels and topography grids initially have a 1 km resolution, but are then extrapolated to a coarser 100 km grid. Likewise the weather elements output by the RSM (60 km resolution) and observed precipitation (25 km) grid are also extrapolated to a 100 km grid. The actual NFDRS calculations are performed at the 100 km grid resolution, but are ultimately interpolated back to a 60 km grid resolution to match the rest of the RSM output.

Methods of analysis

A quality control (QC) analysis was performed on the RAWS data for 263 stations that met the initial acceptance criteria described above. These data were checked for suspicious values (spikes) through a visual inspection of the time series for each variable and RAWS site. Measurements that were clearly in error (e.g. relative humidity over 100 percent, negative wind speed) were considered missing and excluded from the analysis. One RAWS was removed from the study as a result of this process (leaving 262 total stations for analysis),

as it appears to have been physically moved at some point in the last 5 years. The RSM forecast and validation files did not require a similar QC process, but were visually examined via time series plots to ensure that they had been accurately transferred from ECPC to local disk files.

In order to compare the RAWS observations to the RSM forecast or validation data, it was necessary to match the datasets both temporally and spatially. The RSM forecast output was provided as weekly averages of daily values out to 16 weeks (although this study examined only the first 12 weeks of these forecasts), and these have been produced every weekend from 27 September 1997 to present. The RAWS observations include the 1300 local time observations of temperature, humidity and wind speed needed for the calculation of the NFDRS indices, as well as the daily values for maximum and minimum temperature and relative humidity, precipitation and the mean afternoon wind speed.

The first step in matching the datasets was to transform these daily values into weekly averages of daily values. The 1300 local time observations were used to calculate daily values for the four NFDRS indices (which was also done for the RSM forecasts), while the daily maximums and minimums of temperature and relative humidity were used to calculate daily averages. The weekly mean of these daily values were then calculated with the dates for each weekly mean matching exactly the dates for the RSM forecast weekly means. Weekly means with missing values for three consecutive or four non-consecutive days were considered missing. The weekly means of all three datasets were then transformed into monthly and seasonal means. For the forecasts, monthly means were calculated by taking the means of weeks 1–4, weeks 5–8, and weeks 9–12 of each 12-week forecast. Monthly means for the RSM validation and RAWS datasets were computed in the same manner, but using the weekly means for the 12-week period matching each 12-week forecast period. Likewise, forecast seasonal means were computed by taking the mean of all 12 forecast weeks. RAWS and RSM validation seasonal means were computed by taking the mean of the 12 weeks matching each 12-week forecast.

Once the three datasets were matched on a temporal scale, the next step was to match them spatially. The RSM forecast and validation data are output as a grid of values separated by a distance of roughly 60 km. RAWS observations are spatially dissimilar in that they are spaced in an irregular pattern that is not in any way aligned with the model grid output (see Fig. 1). Extrapolation of the station data to a grid matching that of the RSM output was ruled out due to potential error in the values at extrapolated grid nodes with no nearby RAWS or that are close to RAWS with missing data. In this case, interpolating the forecast grids to RAWS locations seems a more accurate method of comparing the datasets. The values in the RSM validation and forecast grids were bilinearly interpolated to match each of the 262 RAWS locations used in this study. This bilinear interpolation algorithm determines

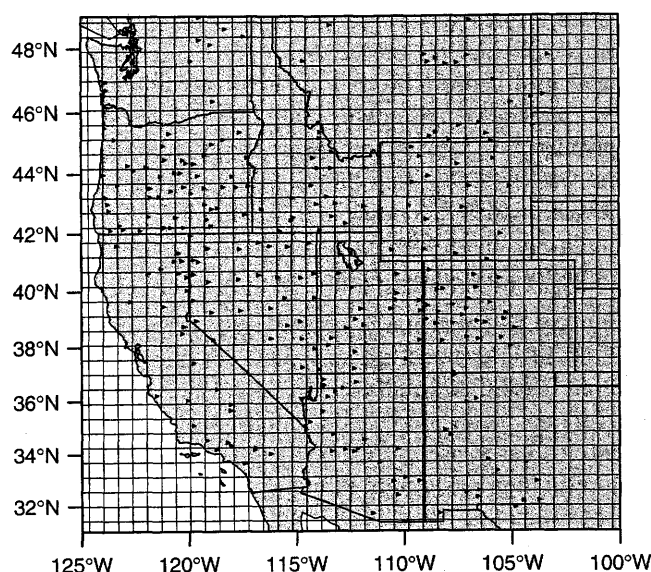


Fig. 1. RAWS locations (black triangles) with RSM output grid overlay (grey lines).

values between grid nodes by calculating a distance-weighted average of values at the nearest four nodes. For the sake of consistency, the ECPC validation v. ECPC forecast comparison uses the forecast and validation grid values as they are interpolated to RAWS sites, rather than making a direct comparison of the forecast and validation output grid nodes. As shall be shown, these evaluations of RSM forecasts with RSM validations at the chosen stations were quite similar to the previous forecast evaluations over the entire RSM grid (Roads 2004; Roads *et al.* 2005).

Appendix 1 describes the statistical measures used to evaluate the forecasts. These measures include bias, which measures the average differences between the forecast climatology and observed climatology; root mean differences, which measure the average squared differences; and anomaly correlations, which measure the consistency of the forecast and observed patterns. Computing the anomalies for each dataset required a calculation of the climatologies for each variable and forecast type in the three separate datasets. However, computing climatologies for use in determining the anomaly correlations was difficult in that the RSM forecast periods did not reset at the beginning of every year because forecasts are made every weekend no matter what time of year it is (in the early ECPC days, maximum computer time was available only on the weekends). This means that the first forecast of one year will not quite match the same dates as the first forecast of another year.

The climatologies for each variable and forecast type were calculated in a manner to solve or at least mitigate this date-matching problem. First, rather than try to compare weekly averages that do not span the same dates from year to year, monthly climatologies for all forecasts beginning in a given

month were computed. Thus, the climatology for any given week or month actually contains 20 values as opposed to just five. For example, a 1-week forecast made in mid-June will be compared to a climatology consisting of all 1-week forecasts made in the month of June during the 5-year period. Likewise, the fifth week of a 12-week forecast would be compared to a climatology consisting of all fifth-week forecasts made in June, a 1-month forecast mean compared to a climatology of all mean 1-month forecasts made in June, and so forth. Attempts to develop interpolated climatologies depending on whether the forecast began at the beginning or end did not seem to make much difference although ECPC does attempt to interpolate the climatologies for displays, which currently only include forecasts at the beginning of the month. Again, the two observational datasets (RAWS and RSM validation) do not contain extended forecasts, but the climatologies for the weekly, monthly and seasonal means matching the forecast dates are computed in the same manner.

Finally, the 12 climatologies (January to December) for a given variable and forecast type were interpolated (in this case linearly) to each individual day within each month. For instance, if the March climatology of daily maximum temperature for seasonal forecasts made in March has a value of 21.11°C and the April value for the same climatology is 26.67°C , each day between 15 March and 15 April increases by a regular amount from 21.11 to 26.67°C . Therefore, the anomaly computed for a seasonal forecast of maximum temperature made on 23 March would then use the smoothed climatology value for 23 March (in this case about 22.59°C).

The performance of the model in summer months is of obvious interest from a fire danger aspect since this is when, for most areas in the West, the fire season occurs. Winter-time performance of the model, while still having relevance to fire danger in areas with warmer climates (i.e. Arizona and southern California), is done here mostly for comparison and contrast with the summer forecasts. These seasonal contrasts also demonstrate the response of fire danger indices calculated in the model with seasonally varying atmospheric elements. In summer cases, the study included all weekly forecasts made in June, July or August (JJA), all monthly forecast means that fall within JJA, and the seasonal forecast for each summer (the seasonal mean from the first forecast week in June). The winter analyses likewise included all weekly forecasts made in December, January or February (DJF), all monthly forecast means that fall within DJF, and the seasonal forecast for each winter (the seasonal mean from the first forecast week in December). Winter analyses will not be graphically shown here due to space constraints, but will be discussed as relevant in comparison to summer season results.

Results

The time series of variability (standard deviation; s.d.) for the RAWS observations (Fig. 2) shows the greatest seasonal

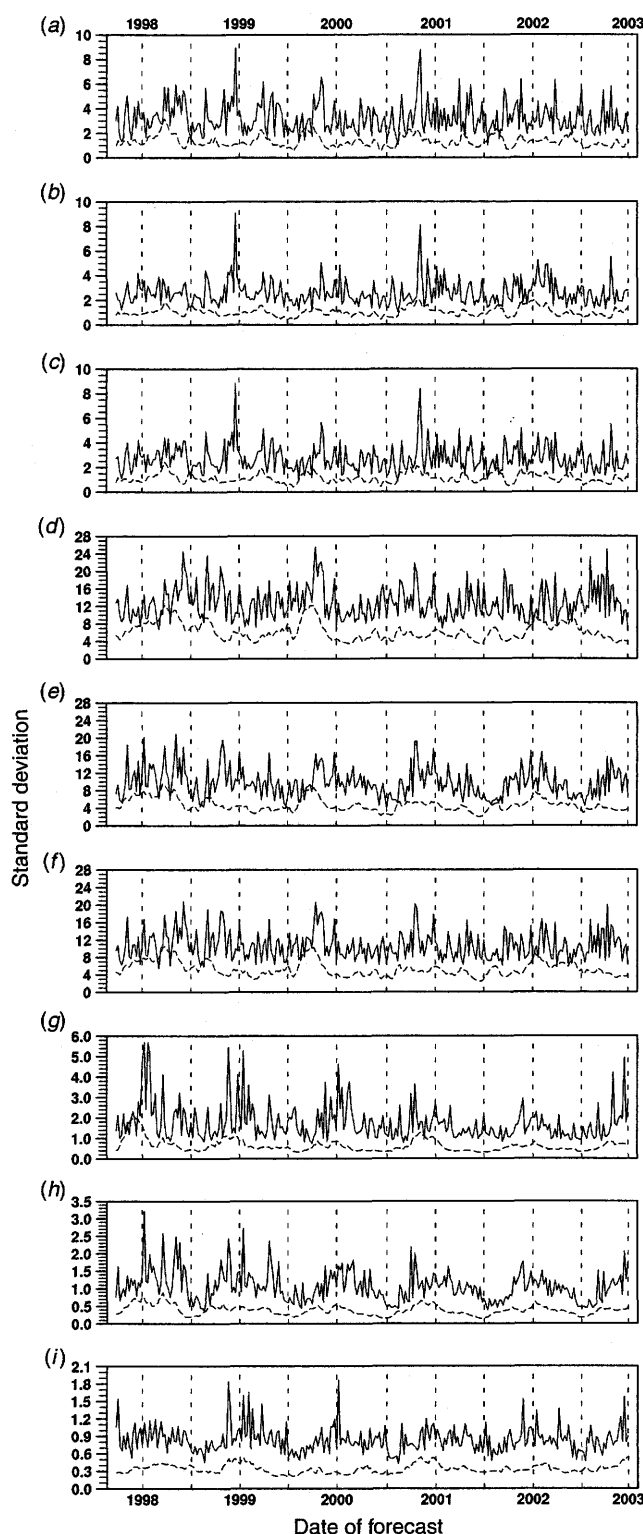


Fig. 2. RAWS seasonal (dashed line) and weekly (solid line) standard deviations (s.d.) for (a) maximum temperature (K); (b) minimum temperature (K); (c) average temperature (K); (d) maximum relative humidity (%); (e) minimum relative humidity (%); (f) average relative humidity (%); (g) precipitation amount (mm); (h) precipitation duration (h); and (i) wind speed (m/s).

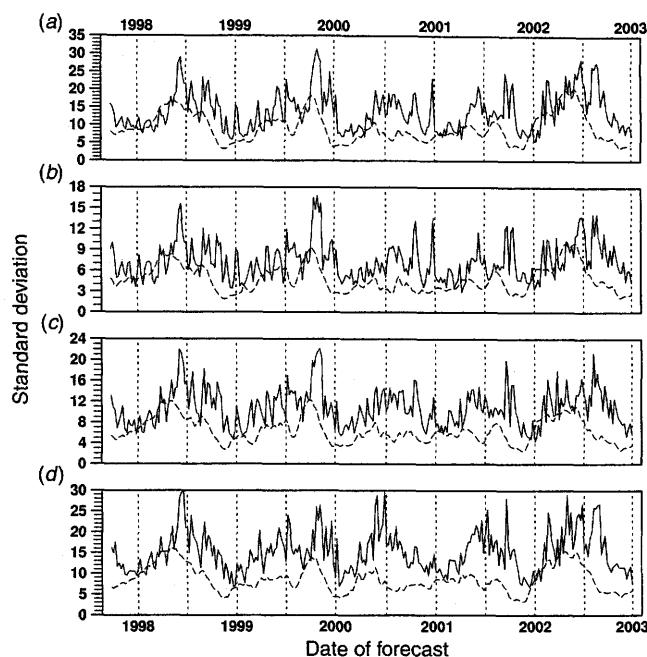


Fig. 3. RAWS seasonal (dashed line) and weekly (solid line) standard deviations (s.d.) for (a) burning index; (b) energy release component; (c) ignition component; and (d) spread component.

variation during the late spring and late autumn for the maximum temperature. However, minimum temperature variability peaks in late autumn and winter. The average temperature profile seems to resemble the minimum temperature in both seasonality and magnitude of variation. Maximum relative humidity variation tends to peak in autumn, with the least variation typically in late winter to early spring. In addition, the variation for maximum relative humidity is somewhat dramatic on a weekly basis. The variation in minimum relative humidity is less extreme than the maximum RH on a weekly basis, but follows a similar seasonal pattern. The average relative humidity reveals little seasonal variation and roughly the same amount of seasonal variation as the minimum RH. The precipitation indices typically show the greatest variation during the winter seasons, of which 1998 and 1999 show noticeably larger variation. The strongest variation in the wind speed typically occurs during the winter as well. The variability in these elements seems to concur with known climate patterns in the West.

The fire danger indices (Fig. 3) all show a similar seasonal pattern. For instance, the burning index variability tends to peak in summer, although the maximum variation in 1999 occurred in autumn. ERC and IC follow a very similar pattern to BI. The SC also peaks in summer, but with greater variation when compared to the other indices. In all cases as expected, the seasonal variability is less than the weekly variability (week to week variations are minimized in a seasonal mean).

RAWS *v.* the ECPC forecasts (RF; Fig. 4) shows a negative bias of about 3 K for both week-1 and seasonal means in the

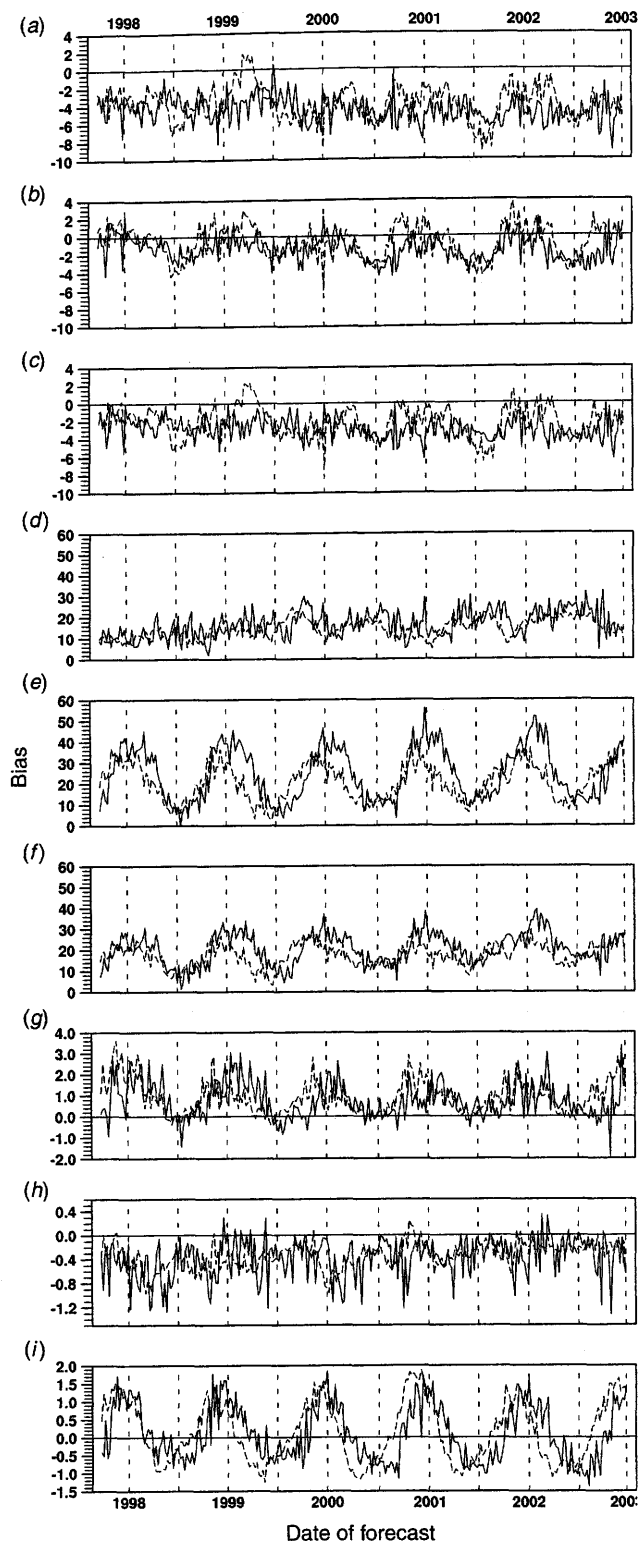


Fig. 4. RAWS *v.* RSM forecast seasonal (dashed line) and week-1 (solid line) bias for (a) maximum temperature (K); (b) minimum temperature (K); (c) average temperature (K); (d) maximum relative humidity (%); (e) minimum relative humidity (%); (f) average relative humidity (%); (g) precipitation amount (mm); (h) precipitation duration (h); and (i) wind speed (m/s).

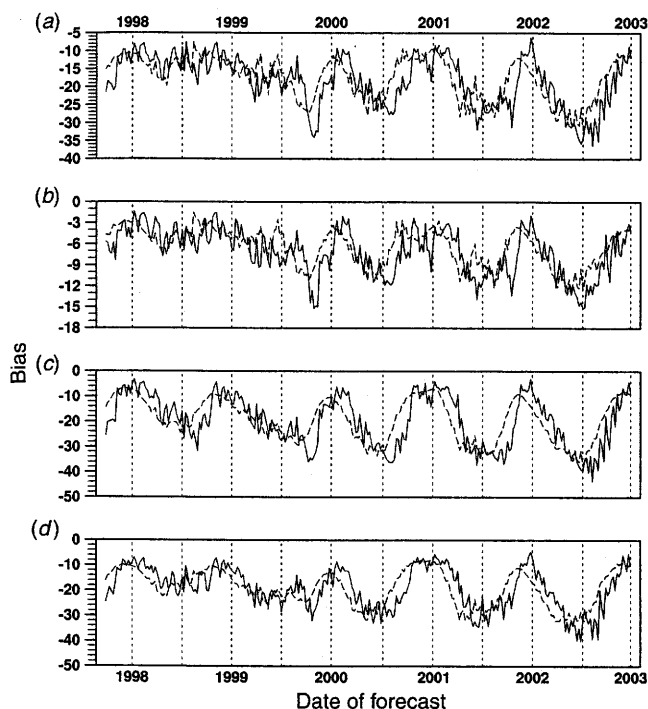


Fig. 5. RAWS *v.* RSM forecast seasonal (dashed line) and week-1 (solid line) bias for (a) burning index; (b) energy release component; (c) ignition component; and (d) spread component.

maximum daily temperature. The negative bias for both week-1 and seasonal means is smaller for minimum and average temperature at approximately -1 K and -2 K, respectively. Seasonal variations in the RF bias of the temperature values are difficult to detect. Maximum daily relative humidity shows a consistent positive bias of about 15% decreasing to around 10% in winter (seasonal variability is difficult to see in the first two years). Minimum and average RH had a strong positive bias of approximately 30% and 25%, respectively. There was clear seasonality in the relative humidity RF bias, with the largest values during winter (40% minimum, 30% average) and lowest values in the summer (5% minimum, 10% average). Precipitation amount had a positive RF bias with a magnitude of 1–2 mm in summer, but much smaller values in winter. Conversely, precipitation duration had a negative bias of one-quarter to one-half hour. Precipitation amount exhibited a seasonal cycle, but none was detected in the duration. There was a seasonal variation for wind speed bias with a positive value of 1.5 m/s in winter and almost no bias in summer. Because there was little differentiation in the magnitude and seasonality of the RF bias for the week-1 and seasonal means, the bias values mentioned above apply to both the week one and seasonal means.

All NFDRS indices (Fig. 5) had a negative RF bias. ERC had the smallest bias with values around -6 . Seasonally, ERC bias was less negative in autumn (-2) than in spring (-9).

BI had a less noticeable seasonal bias than any of the other indices, with an average negative value of -25 . IC and SC had a similar pattern of negative bias, which varies from about -12 to -38 . It should be noted that the bias for all fire danger indices was less negative for 1998.

The RF correlation values in Figs 6 and 7 have been smoothed in order to capture the seasonal cycle. For RF (Fig. 6), the 1-week forecasts show a very high correlation, as might be expected, especially for the temperature and relative humidity indices. There are occasional negative correlations for all six temperature and relative humidity anomalies, but for the most part they have an approximate correlation of 0.6. The precipitation anomalies have a lower but consistently positive correlation of approximately 0.4. The correlation for week-1 wind speed is not as impressive (approximately 0.25), as it is effectively the most difficult element to forecast (as well as measure), but at least the correlation is consistently positive. Seasonal means for all atmospheric anomalies are lower than the 1-week means (typically by 0.4 to 0.2) with the temperature anomalies showing the most dramatic difference. However, all seasonal means remain, on average, positive (most at about 0.2, except for wind speed which correlates at 0.05). In contrast with the atmospheric anomalies (especially temperature and relative humidity), the fire danger anomalies (Fig. 7) show much less, though still positive, correlation (~ 0.2 for week 1 and 0.1 for seasonal). Despite a very low correlation, these anomalies do show a slight seasonal variation, and the correlation tends to improve in late summer and early autumn.

Figures 8 and 9 show for each type of forecast mean the anomaly correlation (AC) performance on a year-round basis of the week-1, month-1 and seasonal mean. The RAWS *v.* forecast (RF), RAWS *v.* validation (RV) and validation *v.* forecast (VF) comparisons are all displayed. Table 1 can be consulted for weeks 2 through 12 and months 1 and 2. All of the weekly means, somewhat predictably, start off with high correlation in week 1, drop significantly by week 2, and level off to a lower, but still positive correlation from weeks 3 through 12 (Table 1). RF atmospheric anomalies (Fig. 8) show a correlation of 0.6 in the first week for the maximum and average temperature. Minimum temperature correlates at a lower value of 0.5. For the relative humidity indices in the first week, minimum and average RH correlate at about 0.475, in contrast to maximum RH, which has a correlation value closer to 0.4. The two precipitation indices correlate similarly, both with values around 0.4. Wind speed correlates less strongly with values closer to 0.24.

In contrast to the RF values, VF atmospheric anomalies (Fig. 8) show high week-1 correlation values above 0.6 for the temperature and RH indices (0.7 or greater for maximum and average temperature). VF precipitation correlation is only a little higher than RF correlation with values of approximately 0.48. Wind speed correlates more strongly as well in the VF comparison with a value of 0.54.

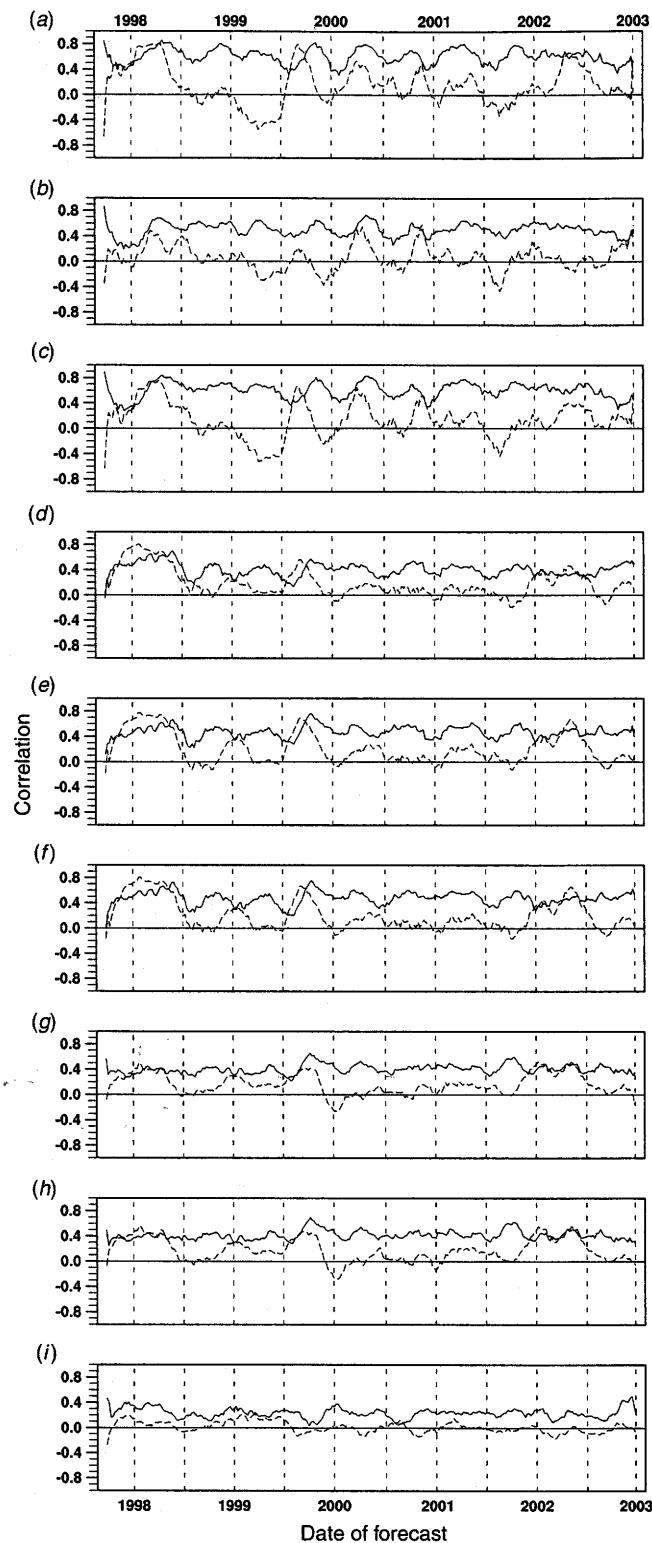


Fig. 6. Smoothed RAWS v. forecast seasonal (dashed line) and week-1 (solid line) anomaly correlations for (a) maximum temperature (K); (b) minimum temperature (K); (c) average temperature (K); (d) maximum relative humidity (%); (e) minimum relative humidity (%); (f) average relative humidity (%); (g) precipitation amount (mm); (h) precipitation duration (h); and (i) wind speed (m/s).

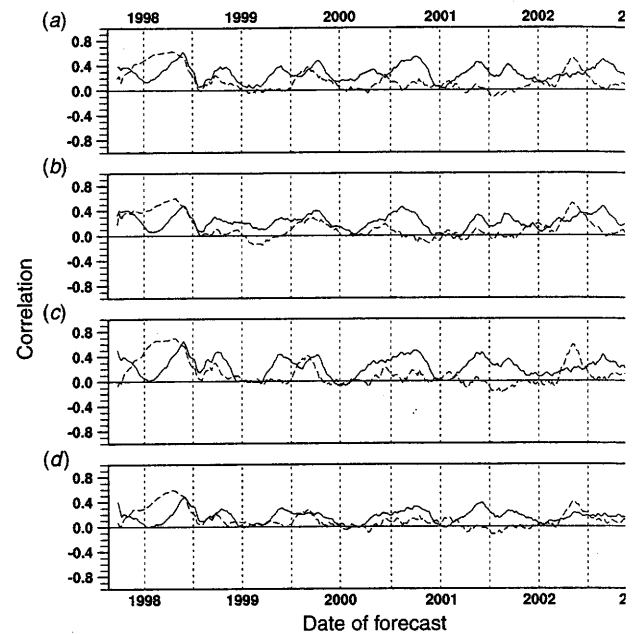


Fig. 7. Smoothed RAWS v. forecast seasonal (dashed line) and week-1 (solid line) anomaly correlations for (a) burning index; (b) energy release component; (c) ignition component; and (d) spread component.

The atmospheric correlation values for the RV comparison (Fig. 8) generally fall somewhere between the RF and VF comparison values for corresponding indices. For instance, the maximum RH correlation has a value of 0.47, compared to 0.4 for RF maximum RH and 0.6 for VF maximum RH. One exception to this is the maximum temperature correlation, which is slightly higher than the RF and VF numbers at 0.7482. The precipitation indices also correlate with higher values than the other two comparisons with values around 0.64.

The greatest disparity between week-1 correlations in the VF and RF comparisons occurs in the fire danger indices. In Fig. 9, none of the RF correlations exceed 0.3. BI is highest with a value of 0.27. The ERC and IC both have values approaching 0.22 at week 1. The lowest correlation is spread component at 0.15. In this comparison, the weekly correlations level out by week 3 with values between 0.06 and 0.1. The RV weekly correlation values (Fig. 9) for the fire danger indices are very similar to the RF values.

In a strong contrast, the VF fire danger indices correlate much higher at week 1 (Fig. 9). The BI and ERC have correlation values of about 0.7. The IC and SC have correlation values closer to 0.6, although these are still much higher than the RF correlation values.

The monthly RF correlations are generally much lower than the week-1 correlations, even for month 1 (Fig. 8). Values for the atmospheric indices are typically close to 0.1 or 0.15 for month 1. In fact, in most cases the RF monthly correlations could probably be approximated by the week

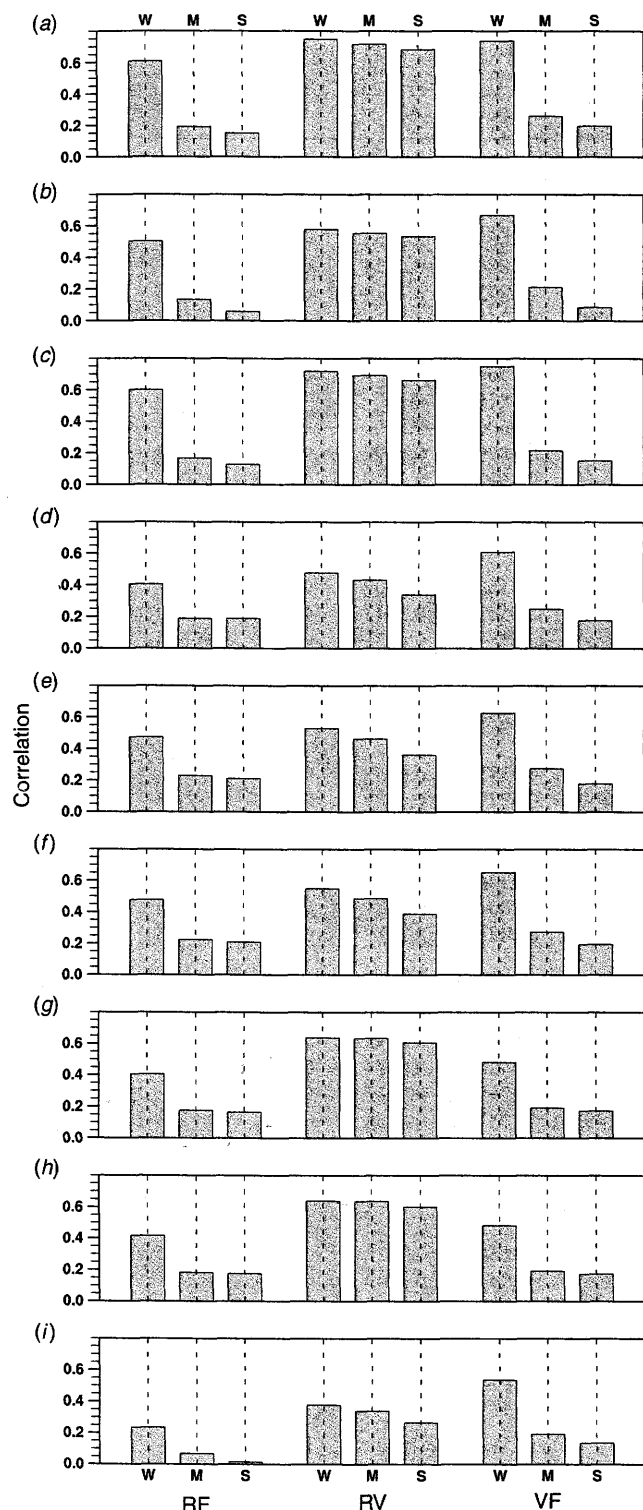


Fig. 8. RAWS v. forecast (RF), RAWS v. validation (RV) and validation v. forecast (VF) week-1 (W), monthly (M) and seasonal (S) anomaly correlations. Atmospheric elements are (a) maximum temperature (K); (b) minimum temperature (K); (c) average temperature (K); (d) maximum relative humidity (%); (e) minimum relative humidity (%); (f) average relative humidity (%); (g) precipitation amount (mm); (h) precipitation duration (h); and (i) wind speed (m/s).

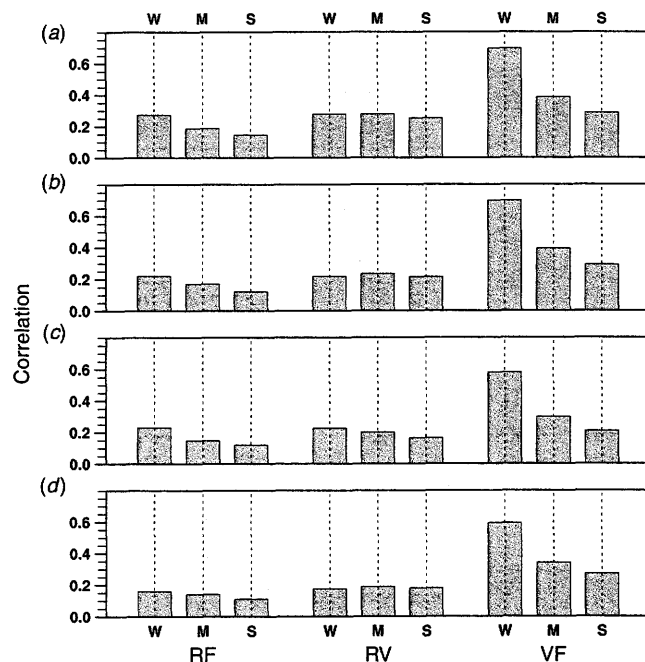


Fig. 9. RAWS v. forecast (RF), RAWS v. validation (RV) and validation v. forecast (VF) week-1 (W), monthly (M) and seasonal (S) anomaly correlations. Atmospheric elements are (a) burning index; (b) energy release component; (c) ignition component; and (d) spread component.

week-6 and week-10 weekly means, although month-3 correlations tend to fall below these values. The VF monthly correlations (Fig. 8) also closely approximate the week-2, week-6 and week-8 values. These correlations have values closer to about 0.2 for the first monthly mean, although VF values for month-3 mean tend to be lower than those correlations for the RF month-3 mean (Table 1). The RV monthly correlation is typically much higher than the other monthly means, and just below the VF weekly correlation value.

RF and VF correlations (Fig. 8) for the seasonal forecast means tend to be lower than the week-1 means, on the order of about 0.15–0.2, although the RF seasonal correlations for minimum temperature and wind speed are less than 0.1. Comparing the RF and VF comparisons, the seasonal VF temperature correlations are higher than the same correlations for all three indices, although the difference is about 0.03. However, the RF RH indices have slightly higher correlations when compared to the same VF values. The seasonal precipitation correlations are approximately equal in both comparisons. The wind speed correlates at what may be a significant seasonal value of 0.14 in the VF comparison, but at a non-significant value of 0.01 in the RF comparison. For the fire danger anomalies (Fig. 9), the RF seasonal means correlate at about 0.1–0.15, compared to values closer to 0.3 for the VF seasonal correlations. RV seasonal mean correlations are very high for the atmospheric anomalies (from 0.35 maximum RH to 0.68 maximum temperature), but drop to

Table 1. Summary of anomaly correlations
Anomaly correlations for weeks 1–4 (W1, W2, W3, W4), months 1–3 (M1, M2, M3) and seasonal (S) forecasts for each element and index described

	W1	W2	W3	W4	M1	M2	M3	S
RAWS v. forecast								
Max. temperature	0.6082	0.1916	0.0302	−0.018	0.1913	0.0647	0.1034	0.1515
Min. temperature	0.5033	0.1447	0.0162	−0.014	0.1333	0.0148	0.0701	0.0584
Av. temperature	0.6011	0.169	0.0173	−0.021	0.1668	0.0502	0.1026	0.1282
Max. RH	0.4054	0.1559	0.0748	0.0646	0.1886	0.0824	0.1238	0.1894
Min. RH	0.4747	0.2025	0.0927	0.0742	0.2293	0.1113	0.1609	0.2108
Av. RH	0.4779	0.19	0.0868	0.0701	0.2246	0.1	0.1485	0.2094
Precipitation								
Amount	0.4067	0.1998	0.1349	0.1252	0.1766	0.1077	0.122	0.1659
Duration	0.4172	0.1939	0.1228	0.1142	0.1826	0.107	0.12	0.1759
Wind speed	0.2354	0.0636	−0.007	−0.017	0.0696	−0.006	−0.011	0.0151
BI	0.273	0.1745	0.1052	0.0875	0.1879	0.0897	0.0787	0.1464
ERC	0.22	0.1632	0.0948	0.0806	0.1699	0.0749	0.0708	0.1211
IC	0.2295	0.1285	0.0766	0.0637	0.148	0.0711	0.0713	0.1201
SC	0.1585	0.1017	0.0736	0.064	0.1414	0.0644	0.0563	0.1107
RAWS v. validation								
Max. temperature	0.7482	0.7482	0.7482	0.7482	0.7169	0.7169	0.7169	0.6824
Min. temperature	0.5777	0.5777	0.5777	0.5777	0.5543	0.5543	0.5543	0.5339
Av. temperature	0.7196	0.7196	0.7196	0.7196	0.6934	0.6934	0.6934	0.6626
Max. RH	0.4792	0.4792	0.4792	0.4792	0.4329	0.4329	0.4329	0.3406
Min. RH	0.5303	0.5303	0.5303	0.5303	0.4639	0.4639	0.4639	0.3625
Av. RH	0.5497	0.5497	0.5497	0.5497	0.4879	0.4879	0.4879	0.3886
Precipitation								
Amount	0.6379	0.6379	0.6379	0.6379	0.6355	0.6355	0.6355	0.6072
Duration	0.6399	0.6399	0.6399	0.6399	0.6379	0.6379	0.6379	0.6025
Wind speed	0.3783	0.3783	0.3783	0.3783	0.3401	0.3401	0.3401	0.2674
BI	0.2781	0.2781	0.2781	0.2781	0.2776	0.2776	0.2776	0.2525
ERC	0.2176	0.2176	0.2176	0.2176	0.2344	0.2344	0.2344	0.2138
IC	0.2259	0.2259	0.2259	0.2259	0.2004	0.2004	0.2004	0.1636
SC	0.1738	0.1738	0.1738	0.1738	0.1877	0.1877	0.1877	0.1796
Validation v. forecast								
Max. temperature	0.7371	0.2227	0.0566	−0.012	0.2595	0.0384	−0.017	0.1967
Min. temperature	0.6675	0.2118	0.0328	−0.025	0.212	0.054	−0.003	0.0857
Av. temperature	0.7492	0.2215	0.0412	−0.028	0.2168	0.0395	−0.013	0.152
Max. RH	0.6091	0.1907	0.0947	0.0542	0.2486	0.0484	0.001	0.1776
Min. RH	0.6269	0.2213	0.0982	0.0515	0.2749	0.0362	−0.033	0.179
Av. RH	0.6525	0.211	0.1028	0.0493	0.2737	0.0358	−0.021	0.1965
Precipitation								
Amount	0.4827	0.1999	0.111	0.1233	0.1923	0.0965	0.0553	0.1743
Duration	0.4827	0.1999	0.111	0.1233	0.1923	0.0965	0.0553	0.1743
Wind speed	0.5373	0.1202	−0.006	−0.013	0.1935	0.0314	−0.009	0.138
BI	0.6955	0.3476	0.1739	0.1413	0.3868	0.0825	0.0014	0.2867
ERC	0.6987	0.3143	0.193	0.156	0.3941	0.0749	0.0004	0.2916
IC	0.5797	0.2632	0.1402	0.1225	0.2975	0.0342	−0.021	0.2083
SC	0.5931	0.3108	0.1524	0.1235	0.3399	0.0661	−7E−04	0.2705

Av., average; BI, burning index; ERC, energy release component; IC, ignition component; Max., maximum; Min., minimum; RH, relative humidity; SC, spread component.

the level of the RF seasonal mean correlations in the IC and SC indices. The RV ERC and BI seasonal correlations are closer to those of the VF comparison at about 0.2.

Discussion and conclusion

The goal of this study was to use RAWS observations to evaluate the skill of a dynamic forecast model, and to assess from

a skill perspective seasonal forecasts of weather elements and fire danger rating indices for potential fire management uses. All biases between the RAWS observations and RS forecasts and validations for both atmospheric elements and a fire danger index had a greater magnitude than the biases between the elements of the RSM forecasts and RSM validations. Overall, the relative humidity biases were the largest. Most bias is likely due to inadequate parameterizations a

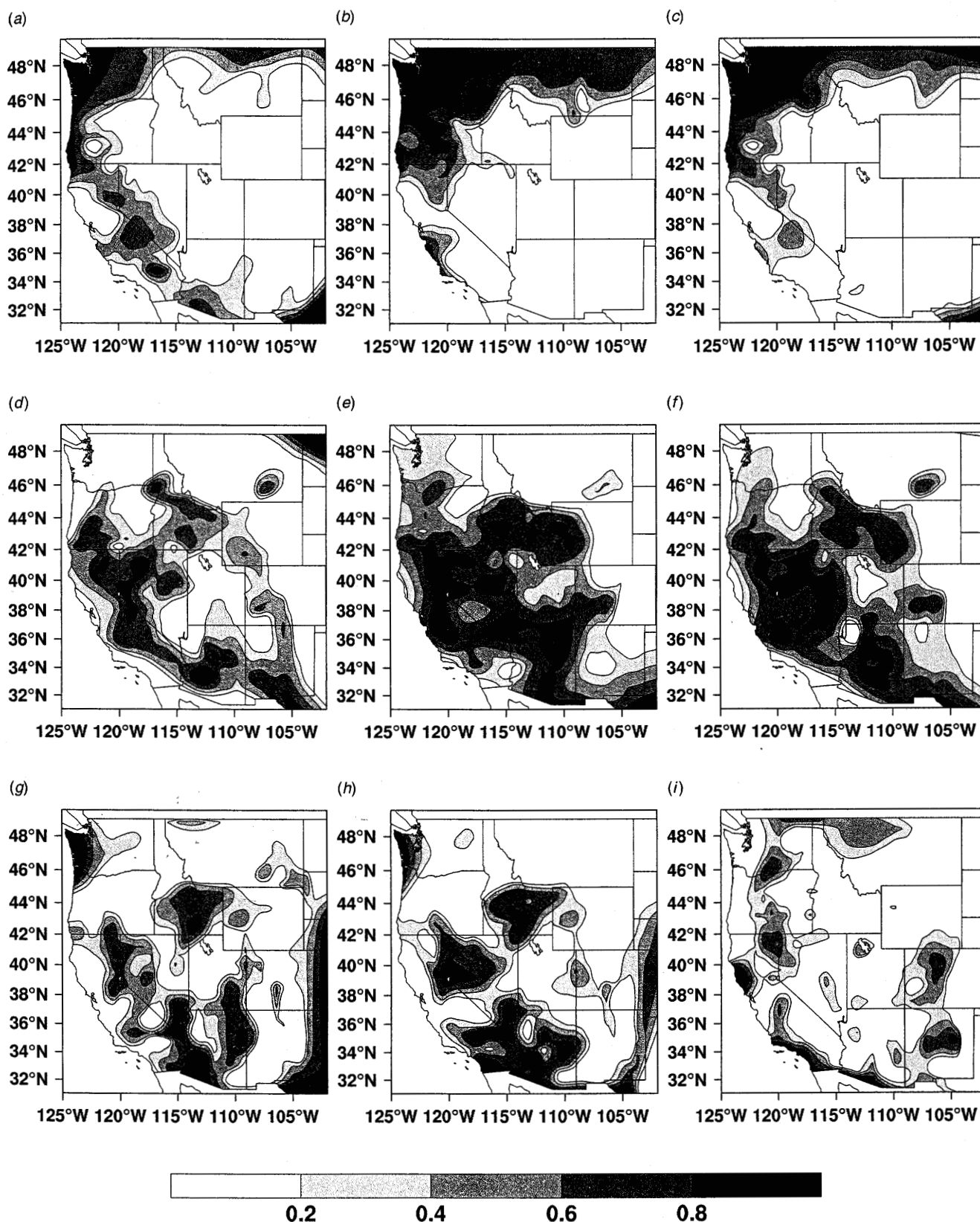


Fig. 10. RAWS v. forecast summer (JJA) season mean correlations for (a) maximum temperature (K); (b) minimum temperature (K); (c) average temperature (K); (d) maximum relative humidity (%); (e) minimum relative humidity (%); (f) average relative humidity (%); (g) precipitation amount (mm); (h) precipitation duration (h); and (i) wind speed (m/s).

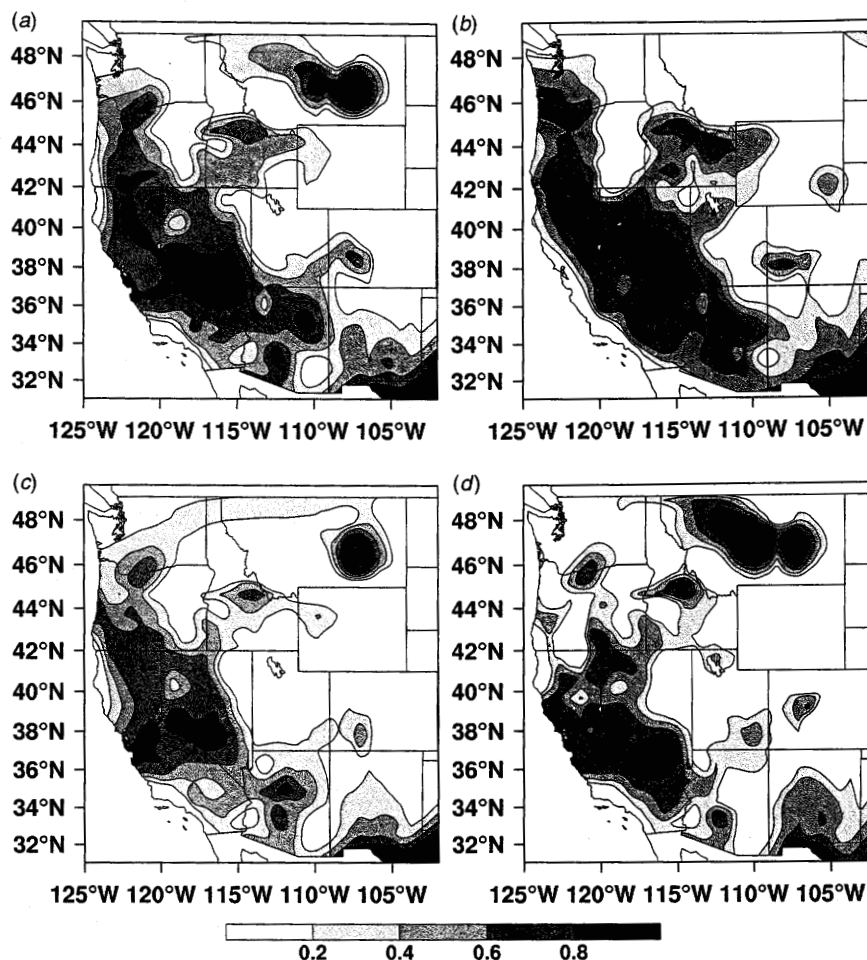


Fig. 11. RAWS v. forecast summer (JJA) season mean correlations for (a) burning index; (b) energy release component; (c) ignition component; and (d) spread component.

model errors, but some of this bias may be attributable to other factors. For the nine atmospheric elements assessed, additional bias may come from differences between RAWS observations and the NCEP data used to create the validation files. RAWS are generally located in forest, wilderness and rangeland areas while atmospheric soundings and surface data incorporated into the NCEP analysis grids are gathered from locations which may be unrepresentative of where wild-fire actually occurs. Additional sources of bias may stem from the fact that the RSM validations (which are 1-day forecasts) are themselves only close approximations of the NCEP data (Roads 2004; Roads *et al.* 2005), and that the RSM surface grid was interpolated to RAWS sites for this study. Using a more advanced interpolation algorithm, reducing output grid size or increasing the number of RAWS in the study might be beneficial in minimizing overall bias, but would not remove it completely. Bias in the fire danger indices may also be due to differences between RAWS and ECPC site descriptions (fuel model, slope) that were used in the NFDRS equations. Again, the ECPC fire danger indices were calculated on 100 km

fuel model and slope grids, and then interpolated to 60 km grids for comparison with the atmospheric RSM indices. The fire danger indices calculated from RAWS observations were based on the WIMS descriptions for each RAWS site, not grids.

Biases are assumed consistent, and could be removable after the fact. What is more important, therefore, is the accuracy of anomaly forecasts with respect to the model climatology in comparison to anomalies with respect to the observed climatology. In addition the seasonal cycle can be large, and not removing it can make the forecasts seemingly have high correlations without actually having substantial skill. Correlations between datasets were therefore computed using anomalies (deviations from the appropriate climatology) rather than the original forecast or observational values. The results of the anomaly correlations for all three studies are summarized in Table 1. As noted in the previous section, the VF (ECPC validating observations v. ECPC forecasts) comparison correlations were generally higher than the RF (RAWS observations v. ECPC forecasts) comparison at

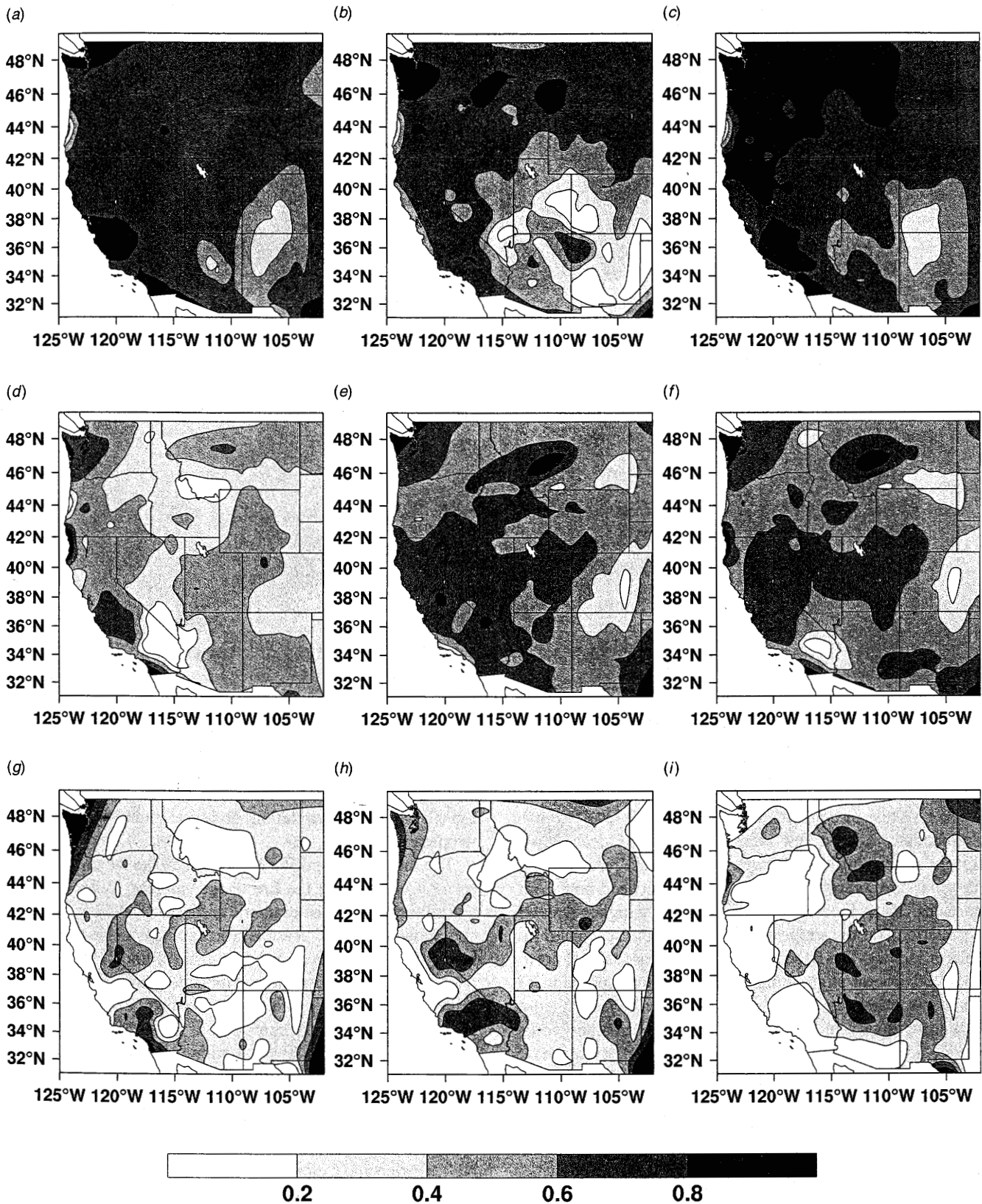


Fig. 12. RAWS v. forecast summer (JJA) week-1 mean correlations for (a) maximum temperature (K); (b) minimum temperature (K); (c) average temperature (K); (d) maximum relative humidity (%); (e) minimum relative humidity (%); (f) average relative humidity (%); (g) precipitation amount (mm); (h) precipitation duration (h); and (i) wind speed (m/s).

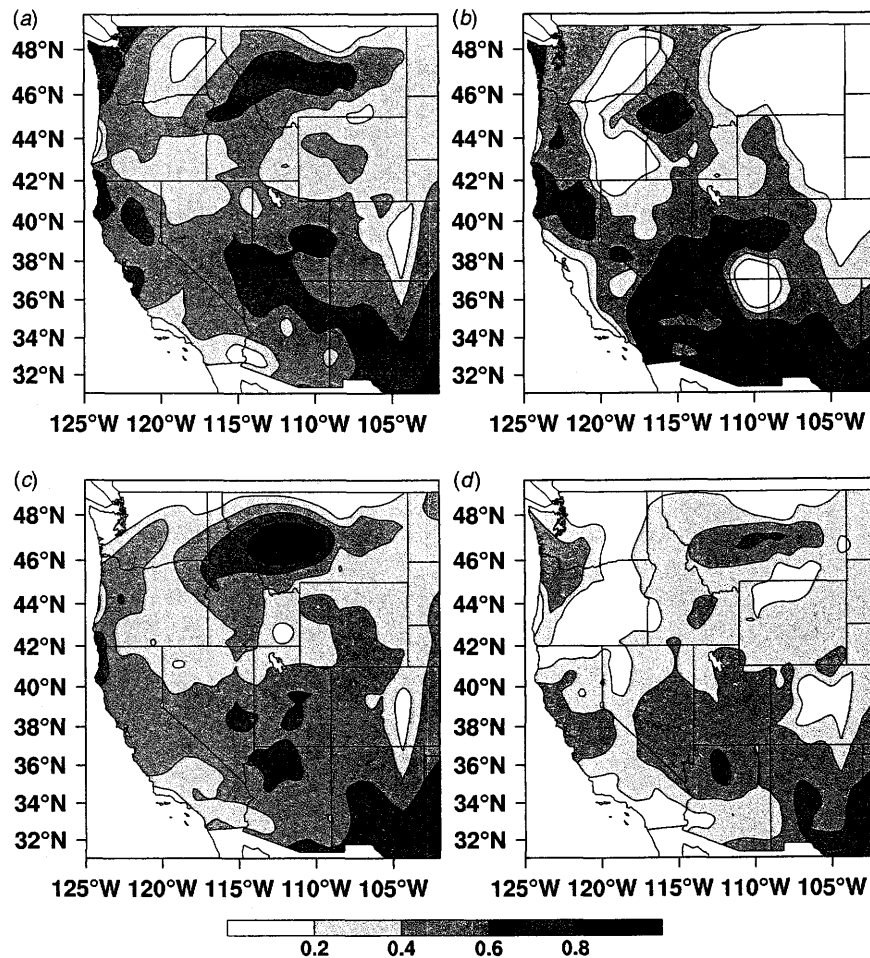


Fig. 13. RAWS v. forecast summer (JJA) week-1 mean correlations for (a) burning index; (b) energy release component; (c) ignition component; and (d) spread component.

week 1. While in both cases the weekly skill level dropped dramatically after a few weeks, the rate of decline was not noticeably different between the two comparisons. The seasonal correlations were very similar between both comparisons. The VF correlations were slightly higher than RF correlations for the temperature indices, wind speed and fire danger indices. The RF comparison had seasonal RH correlations that were slightly higher than the corresponding VF comparison indices. The precipitation indices correlated at about the same value for both comparisons. In other words, the precipitation forecasts compare equally well to the RAWS precipitation and the Higgins gridded dataset. While the seasonal correlation values were low for all indices in both the RV (RAWS observations v. ECPC validating observations) and VF comparisons, they were higher than the extended weekly values (with the exception of RF wind speed, which had a seasonal correlation close to zero). The RV comparison had relatively high correlation values for the atmospheric indices at all time scales (as might be expected when comparing what are essentially two different sets of observations),

especially the precipitation and maximum and average temperature.

The most striking difference between the RF and VF correlation values is in the fire danger indices. While week-1 values show the most dramatic difference between the two comparisons, the VF seasonal forecasts correlate noticeably better than the RF seasonal forecasts, although all correlation values are high enough to indicate some skill in model forecast. Of the four indices, BI typically correlated best, especially in RF and RV comparisons. In the VF comparison, BI and ERC have similar correlation values.

In any event, indices with low overall correlation showed some forecast skill in specific locations, especially during a given season. For instance, seasonal wind speed correlated with values at or above 0.6 in parts of Washington, Oregon, New Mexico and Colorado during the summer (Figs 10, 11). Figures 10–15 show the spatial distribution of skill for the summer (JJA) seasonal forecasts. In general, the week-1 forecasts (Figs 12, 13) show broader areas of high skill compared to month-1 (Figs 14, 15) and seasonal forecasts. It is also

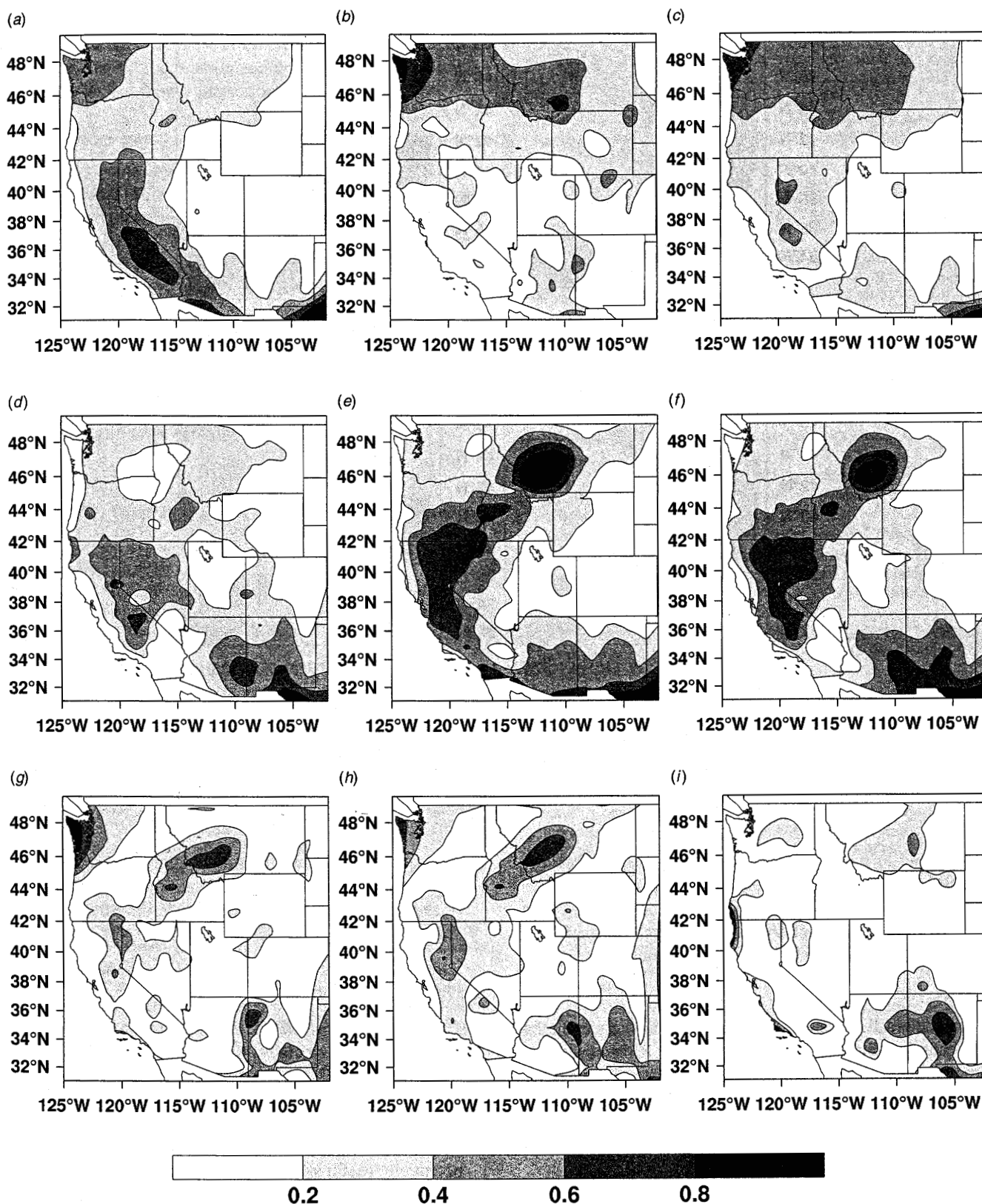


Fig. 14. RAWS v. forecast summer (JJA) month-1 mean correlations for (a) maximum temperature (K); (b) minimum temperature (K); (c) average temperature (K); (d) maximum relative humidity (%); (e) minimum relative humidity (%); (f) average relative humidity (%); (g) precipitation amount (mm); (h) precipitation duration (h); and (i) wind speed (m/s).

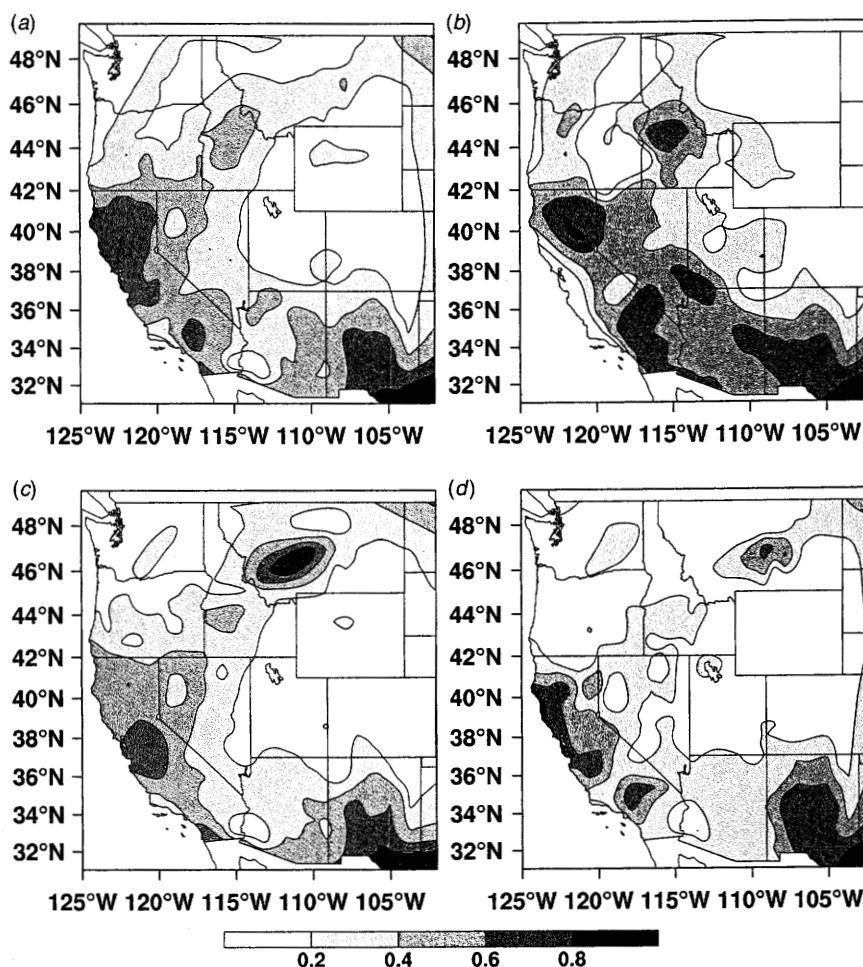


Fig. 15. RAWs v. forecast summer (JJA) month-1 mean correlations for (a) burning index; (b) energy release component; (c) ignition component; and (d) spread component.

worthwhile to note that the results for the RSM forecasts v. RSM validation comparison were comparable to studies performed at ECPC for the fire danger indices (Roads 2004), despite the reduced number of RAWs sites in comparison to the number of RSM grid points used previously.

Specific recommendations

For modelers, these results provide some encouragement that fire danger indices can be forecast by this RSM at seasonal time scales, even if the current skill is not high. These results also serve as a useful contrast of RSM skill when compared to RAWs observations rather than the validating observations. The following suggestions are made:

- Re-evaluate the algorithm used to output fire danger indices, focusing on elements like fuel moisture and carry-over values. This will help ensure the NFDRS algorithm used by ECPC matches NFDRS calculations for individual RAWs as closely as possible.

- Incorporate RAWs into the initialization of the model (perhaps even as part of the NCEP/NCAR operational analysis or reanalysis); this would likely aid in improving model skill with reference to RAWs.

Fire management should consider the potential value of the RSM forecast for their specific site:

- The skill of most of the week-1 forecasts of atmospheric indices (especially temperature) is very high. These values could be useful in making short-term management decisions.
- While seasonal skill as an overall average is low, especially for the fire danger forecasts, most indices still show some potentially useful skill in some regions.

Future work on this topic might include a more in-depth examination of the large biases. Also, performing a similar study between the ECPC GSM forecasts and RAWs observations would be helpful, given that most seasonal forecasts

currently use global rather than regional models. Upgrading the GSM and RSM to the latest versions of each may improve the forecasts of precipitation and wind speed; with possible increases in fire danger forecast skill as a result.

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Appendix 1. Statistical measures

The purpose of forecast verification is to determine the quantitative accuracy of the forecast. As described previously by Roads *et al.* (2005) the statistical methods employed in this study as a means of forecast verification were bias, root mean square error, anomaly correlation and standard deviation. Due to space constraints, the root-mean square error and standard deviation are not discussed here, but these measures can be found in Reinbold (2003).

Bias is a simple calculation of forecast minus observation (Wilks 1995), or

$$\text{bias} = f - o, \quad (1)$$

where f is the forecast value and o is the value of the observation. When shown in graphical form, this calculation has the benefit of revealing under what situations the model is over- or under-forecasting and by how much. It is also useful in determining potential seasonal characteristics in the errors between the datasets.

Anomaly correlations are commonly used to evaluate extended forecasts. This measure is designed to reflect good forecasts in the pattern of an observed field, but does not effectively measure the magnitude of the values (Wilks 1995). There are two different equations, representing the two types of anomaly correlation used in this study. The first is for judging the spatial variation and correlation of the anomalies (Roads 2004; Roads *et al.* 2005) and is not shown here. The

second is better described as temporal variations in spatial correlations (equation 3; Roads *et al.* 2005). Anomalies are first computed by taking the difference between the total forecast (either by region or for the entire Western USA) and the climatological monthly means. In other words,

$$A = f - C_f, \quad (2)$$

where A is the anomaly, f is the forecast and C_f is the climatological mean for that forecast type (weekly, monthly or seasonal mean).

Given that A is a forecast anomaly of any type (weekly, monthly or seasonal mean) and that B is the validating anomaly from observation, the temporal variations in the spatial anomaly correlations (AC, sometimes known as pattern correlation) are calculated using

$$AC = \frac{\sum_{m=1}^M [A_m B_m]}{\left[\sum_{m=1}^M A_m^2 \sum_{m=1}^M B_m^2 \right]^{1/2}}, \quad (3)$$

where M is the total number of RAWs in the current region ($M = 262$ for the western USA). Missing values in the observational anomalies for equation (3) reduce the anomaly summations for both datasets by the number missing (the missing RAWs values and matching interpolated validation or forecast values are removed).

Modeling interactions between fire and atmosphere in discrete element fuel beds

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Abstract. In this text we describe an initial attempt to incorporate discrete porous element fuel beds into the coupled atmosphere–wildfire behavior model HIGRAD/FIRETEC. First we develop conceptual models for use in translating measured tree data (in this case a ponderosa pine forest) into discrete fuel elements. Then data collected at experimental sites near Flagstaff, Arizona are used to create a discontinuous canopy fuel representation in HIGRAD/FIRETEC. Four simulations are presented with different canopy and understory configurations as described in the text. The results are discussed in terms of the same two discrete locations within the canopy for each simulation. The canopy structure had significant effects on the balance between radiative and convective heating in driving the fire and indeed sometimes determined whether a specific tree burned or not. In our simulations the ground fuel density was the determining factor in the overall spread rate of the fire, even when the overstory was involved in the fire. This behavior is well known in the fire meteorology community. In the future, simulations of this type could help land managers to better understand the role of canopy and understory structure in determining fire behavior, and thus help them decide between the different thinning and fuel treatment strategies available to them.

Additional keywords: fire propagation; FIRETEC; HIGRAD.

Introduction

Wildfires comprise a complex set of physical and chemical processes, some of whose interactions depend on the coupling between atmospheric flows, the fire and the vegetation structure. This coupling affects many aspects of the wildfire, such as the balances between different modes of heat transfer, the turbulent mixing which brings gaseous reactants together, and the shape of the fire perimeter.

The influence of plant canopies on atmospheric flows is a well-developed field of study. Much of this work has been focused on characterizing either the average effects of the integrated canopy, the turbulent mixing that is induced by the canopy, or the integrated effects of the canopy on the shear layer above the fuel bed (i.e. Brunet *et al.* 1994; Finnigan and Brunet 1995; Shaw *et al.* 1995). Scaling laws have been used to relate the geometry of obstacles that make up a canopy to the boundary layer that they influence. Work by Belcher *et al.* (2003) describes numerous relationships that characterize the integrated effects of arrays of obstacles on net flow and roughness lengths both within and above the obstacles.

This work is focused on spatially averaged time-mean flows. A closure scheme for flow around tree structures has recently been developed by Calogine and Séro-Guillaume (2002). This work uses a fractal porous media model and averaging techniques to generate a tree-like obstruction that acts as an obstacle for air flow in direct numerical simulations. From these simulations a closure model is generated for the Navier-Stokes equations when applied to flow around a tree. This detailed look at flow around and through tree-like structures represents an important step in being able to better simulate airflow through irregular canopies including vegetation of different sizes and shapes.

There has also been a variety of research done on wildfire propagation of a crown fire as described in Albini and Stocks (1986), Rothermel (1991), Van Wagner (1993), Call and Albini (1997) and Cruz *et al.* (2002). In general, much of this research is focused on the gross aspects of fire behavior such as average spread rate and mean fuel consumption as a fire moves through a canopy. These works have provided critical information for operational fire behavior prediction,

usually based on balances between averaged canopy-scale processes. However, they have not focused on explaining the balances in physical processes at the tree-scale or the manner in which the heterogeneities of canopy structure change these balances.

The ongoing development of the HIGRAD/FIRETEC wildfire modeling system presents a new tool for the investigation of coupled atmosphere–fire–fuel interactions. This modeling framework is composed of an atmospheric model, HIGRAD (Reisner *et al.* 2000a), which is designed to capture high gradients in quantities such as temperature and velocity, and a wildfire model, FIRETEC (Linn 1997). FIRETEC is based on conservation of mass, momentum, species and energy. HIGRAD/FIRETEC has been used to examine fire behavior and the interaction between the fire and atmosphere in several idealized simulations, including idealized grass fires and homogenized understory and canopy simulations. These idealized simulations show strong interaction between fire, atmosphere and vegetation structure. The details concerning the formulation of FIRETEC or HIGRAD or their numerical implementation will not be discussed as most of these details can be found in other publications (Linn 1997; Reisner *et al.* 2000a, 2000b; Linn *et al.* 2002).

Here we describe an initial attempt to incorporate discrete element fuel beds into HIGRAD/FIRETEC. We address questions such as how a fire responds to changes in canopy arrangement or understory fuel load, and how radiative and convective heat transfer work together to determine which trees will burn and which will not. The structure of the vegetation in these simulations is based on individual tree data measured at experimental sites near Flagstaff, Arizona, which have been arranged as a representation of discontinuous canopy vegetation that resembles ponderosa pine forests. These simulations allow us to examine the interaction of wind and fire in discontinuous fuel beds and, more specifically, to investigate the role that the canopy structure plays in such fires.

In this text we introduce some of the conceptual models that we have used to translate the measured tree data into discrete fuel elements that can be employed in FIRETEC. We describe some simulations that have been performed with these discrete fuel elements (we shall refer to the fuel elements as trees even though they are in fact crude representations of trees), and report on some of the output produced by these simulations. Specific results that demonstrate interaction between the fire and the atmosphere are highlighted as an indication of the potential benefits of using HIGRAD/FIRETEC to investigate wildfire behavior in discontinuous canopies.

Fuel representation

HIGRAD/FIRETEC is a finite volume computer program that uses a structured three-dimensional grid to describe

evolving, spatially varying quantities such as temperature, the velocity of the gaseous species and various characteristics of the fuel. In its current implementation it cannot represent individual leaves, needles or twigs as done by other researchers using fractal (Calogine and Séro-Guillaume 2002) or other methods. Instead, the multiphase stochastic approach used in FIRETEC represents vegetation as a porous medium that provides bulk momentum and heat exchange terms between gas and solid phases. Fuel is described by assigning mean or bulk quantities such as fine fuel surface area per unit volume, moisture ratio and density to each cell in the three-dimensional grid. This methodology potentially allows HIGRAD/FIRETEC to simulate complex fuel beds that are vertically and horizontally non-homogeneous. The approach is described in detail in Linn (1997) and has similarities to the multiphase approach used by Morvan and Dupuy (2001).

Researchers from the USDA Forest Service (led by Carleton Edminster) and Northern Arizona University (led by John Bailey) measured canopy vegetation characteristics on several 20 m × 50 m plots near Flagstaff, Arizona. Data were collected for each tree within the plots as a part of the Fire/Fire Surrogate program, which is funded by the Joint Fire Sciences Program with additional funding from the Rocky Mountain Research Station. The data that were collected included tree height, height to live crown, height to dead crown, diameter of the crowns in two directions, and diameter of the trunks at breast height for each tree within the plots.

After this extensive sampling effort, the challenge was to convert the collected data into three-dimensional fuel distributions that could be used in HIGRAD/FIRETEC to simulate a fire burning through trees similar to those measured. For this initial effort, we chose to work with a simple bulk averaged structural model for a ponderosa pine tree. In the future, we hope to explore other structural models in collaboration with other researchers. It was assumed that the majority of the fine fuels form a shell around the outside of the canopy that can be described with paraboloids, and that the density of the fine fuels declines from the outside of the tree to the center of the tree (especially near the bottom of the canopy). We also chose to treat the trees as if they were axisymmetric.

With these ideas in mind, we used a series of parabolic profiles to generate a function that describes the distribution of the mass of an idealized tree as shown in Fig. 1. From these profiles and the locations of the trees, we can establish the bulk density of fine vegetation within each cell of the three-dimensional HIGRAD/FIRETEC mesh. Each cell in the mesh can contain part of one tree, parts of more than one tree, or be in a location where there are no trees.

The first step in resolving each tree is to define the locations that are inside the tree perimeter. The formulation that we used to define the interior of the tree crown is shown in equation (1) in terms of a height above the ground, z , and a

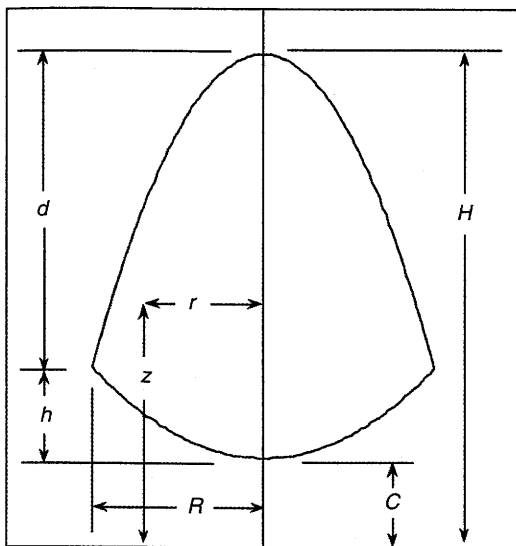


Fig. 1. This diagram shows a two-dimensional cross section of the axisymmetrical three-dimensional shape that we use to describe the crowns of trees in our simulations. The black outline of the tree canopy consisting of two parabolas bounds a region where bulk density is non-homogeneously distributed. A location inside the crown is specified by z and r . C , R and H are specified from the field data.

distance from the center of the particular tree, r :

$$\frac{h}{R^2}[r^2] + C \leq \text{crown}_{\text{interior}} \leq -\frac{d}{R^2}[r^2] + H. \quad (1)$$

The variables are illustrated in Fig. 1. The variables h and d are used to scale the parabolic shapes that form the top and bottom of the crown and, when summed, they equal the difference between the height of the tree, H , and the height to crown, C . An estimated value of $h = d/4$ is currently being used based on a limited number of observations. R is the crown radius at its widest point. The density in the interior of the tree, ρ_{canopy} , is approximated by equation (2):

$$\rho_{\text{canopy}} = \left[\frac{z - C + \frac{d}{R^2}r^2}{H - C} \right] \rho_{\text{max}}. \quad (2)$$

By requiring the integral of this function over the interior of the crown to be equal to the volume of the crown multiplied by an average crown bulk density (bulk density of the crown of a single tree), we get the expression in equation (3):

$$\rho_{\text{max}} = \frac{6}{5} \rho_{\text{avg}}. \quad (3)$$

The average crown density for individual trees was chosen to be 0.4 kg/m^3 based on discussions with forestry researchers. These expressions provide a method for converting standard tree canopy measurements for ponderosa pine trees into a spatial distribution of vegetation bulk density suitable for use in HIGRAD/FIRETEC.



Fig. 2. A cut-away cross-section of a portion of the fuel bed used in the patches simulation. The image shows an isosurface of fuel density at a value of 0.3 kg/m^3 looking across a section roughly $100 \text{ m} \times 100 \text{ m} \times 20 \text{ m}$.

Because the sampled plots were only $20 \text{ m} \times 50 \text{ m}$ in size and we wished to simulate fires over much larger regions, we chose to distribute the trees randomly in several adjacent $20 \text{ m} \times 50 \text{ m}$ domains. This methodology allows us to generate large stands of trees without having the periodic arrangement that would occur if we used the same locations for the trees within each $20 \text{ m} \times 50 \text{ m}$ plot.

The analytical expressions in equations (2) and (3) describe the distributions of bulk density within the canopy volume outlined by equation (1). Because most of the trees in the database are between 10 and 18 m tall and between 2 and 6 m in diameter, the trees will at least partially occupy multiple computational cells that have horizontal dimensions of 2 m and vertical dimensions of $\sim 1.5 \text{ m}$ near the ground. In order to use the idealized trees described above, we must convert the analytically described, continuously varying fuel distributions into discrete values that represent average bulk properties for each cell that contains part of a tree. This is done by numerically integrating the density over the volume of canopy that lies within each cell (as defined by the tree's geometrical data and location) and dividing it by the cell volume as described in equation (4):

$$\bar{\rho} = \frac{\int \rho_{\text{canopy}} dv}{\text{cell volume}}. \quad (4)$$

Figure 2 displays a cut-away cross section of a fuel bed that was generated using the methodology described above. In Fig. 2, many different sizes and shapes of trees are illustrated by drawing an isosurface at the locations where the bulk density of the vegetation was 0.3 kg/m^3 .

The current model for ground fuel includes a load for grass and a load for litter. We make the assumption that the amount of grass fuel falls off as a function of the mass of canopy above it, normalized by the maximum expected mass for the average tree canopy depth, s , while the litter load increases as a function of the amount of canopy above it. This model is described by equation (5), in which m_{ground} is the mass per square meter of the combined grass and litter, m_{grass} is the mass per square meter of the grass in the lowest cell when there is no canopy above it, m_{litter} is a assumed maximum litter mass per square meter on the ground under maximum canopy cover, and c_g and c_l are non-dimensional proportionality constants.

$$m_{\text{ground}} = m_{\text{grass}} e^{-c_g s} + m_{\text{litter}} (1 - e^{-c_l s}). \quad (5)$$

The values of m_{grass} , m_{litter} , c_g and c_l , decided in consultation with forest service researchers, were chosen to be 0.7 kg/m^2 , 33 kg/m^2 , 5 kg and 5 kg respectively for the set of simulations described in this text. It should be noted that the litter density almost never gets above 0.25 as s seldom reaches 1.0. The bulk density of ground fuels is added to any bulk density that low canopy might contribute to that cell. One trait of this representation of the fuel is that the size of the grid cells can greatly affect the bulk density of the fuel in the cell but does not change the mass of the fuel in the cell that is available to burn.

Simulations

Some specific questions that arise when trying to examine wildfire responses to fuel arrangement and structure are:

1. Does the distribution of tree shapes and sizes within a canopy change the likelihood of a particular tree being burned?
2. Does the position of a tree within a canopy change the likelihood of that specific tree being burned?
3. Is the canopy structure the most important factor in determining spread rate or is the speed of the ground fire a critical factor?

We performed four simulations with different canopy and understory configurations. In each of the simulations we ignited a fire near a 60 m wide fuel break (area with no fuel). The fuel break was included to eliminate the effects of variations in canopy structures upwind of the ignition location. The fire was blown with a 6 m/s wind across a 20 m grass area toward a forested region. The forested region was 220 m wide, with a 20 m area of grass on the far side of the forest. The total horizontal domain of the simulations was $320 \text{ m} \times 320 \text{ m}$, including the forest, the grasslands and the fuel break. The horizontal resolution near the ground was 2 m, and the vertical resolution was $\sim 1.5 \text{ m}$ at the ground. The FIRETEC time step in these simulations was 0.002 s. The simulations were allowed to continue until the fire neared the boundary of the mesh (150–220 s after ignition).

In the first simulation, which we will refer to as the ‘full’ forest simulation, the canopy was created using the methodology described above over a $220 \text{ m} \times 320 \text{ m}$ region. This simulation is depicted in Fig. 3. The canopy in this simulation consisted of over 5000 trees. The canopy bulk density over the forested region was $\sim 0.24 \text{ kg/m}^3$.

The forest canopy in the second simulation was derived from the canopy in the first simulation by removing every tree that had a trunk diameter of less than 28 cm at breast height. Trunk diameter was one piece of the field data that was collected for each tree. The diameter threshold was chosen arbitrarily for testing purposes, but is a variable often employed in thinning strategies. An image from this simulation is shown in Fig. 4. There were only about 1000 trees in this forested area. The canopy bulk density over the forested

region was $\sim 0.055 \text{ kg/m}^3$. We will refer to this simulation as the ‘thin’ simulation for the rest of the text.

The third simulation, hereafter called the ‘patches’ simulation, contained a canopy derived from that used in the full forest case. In this simulation, trees were removed to leave patches of varying-sized trees. The patches were placed arbitrarily, with the mean spacing between patches in the order of 50 m and the average radius of the patches $\sim 15 \text{ m}$. An image from this simulation is shown in Fig. 5. The canopy bulk density over the forested region was $\sim 0.066 \text{ kg/m}^3$.

The canopy used in the fourth simulation is the same as the canopy in the patches simulation, but the grass in the understory has been cut in half (by reducing both the height and mass of the fuel). An image from this simulation is shown in Fig. 6. We refer to this simulation as the ‘patches, low ground fuel’ (PLGF) simulation. The four simulations are summarized in Table 1.

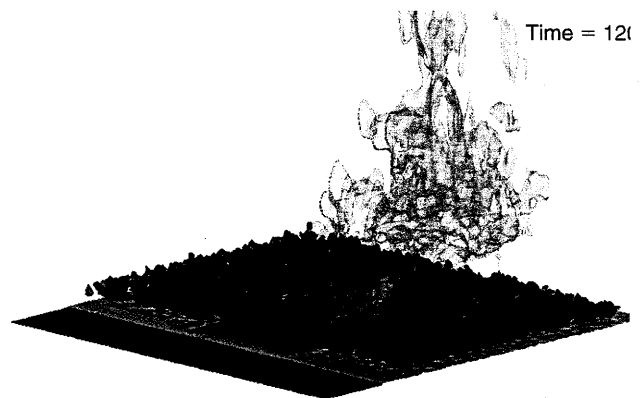


Fig. 3. Image from the full simulation 120 s after ignition. The colors on the horizontal plane indicate bulk ground fuel density, with black being no fuel and bright green being 1 kg/m^3 (0.7 m tall grass). The brown areas are locations where there is less grass, but some fuel litter. The dark green isosurfaces show the tree locations, and the black isosurfaces show parts of the canopy that have been significantly burned. The orange, red, and grey isosurfaces indicate regions of hot gases.

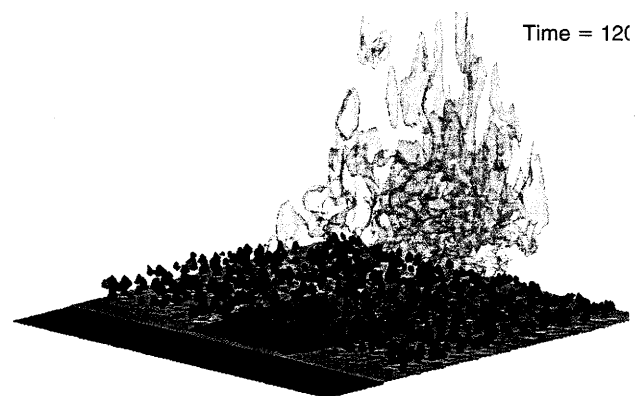


Fig. 4. Image from the thin simulation 120 s after ignition. The colors in this image are similar to those in Fig. 3.

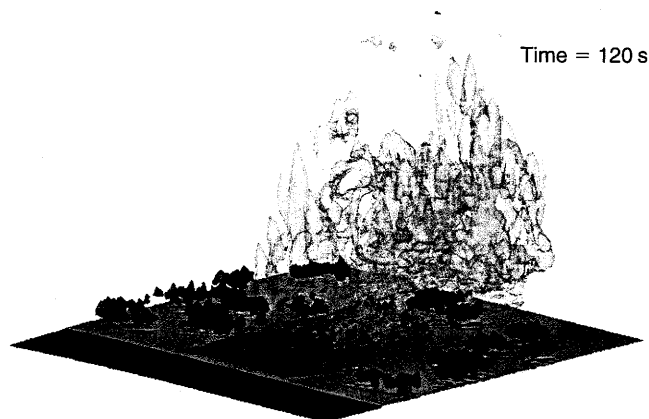


Fig. 5. Image from the patches simulation 120 s after ignition. The colors in this image are similar to those in Fig. 3.



Fig. 6. Image from the patches, less ground fuel (PLGF) simulation 120 s after ignition. The colors on the horizontal plane indicate bulk ground fuel density, with black being no fuel and brown being short grass and litter. The dark green isosurfaces show the tree locations, and the black isosurfaces show parts of the canopy that have been significantly burned. The orange, red, and grey isosurfaces indicate regions of hot gases.

Results and discussion

The reason for performing the four simulations described in the previous section is to provide a basis to demonstrate that HIGRAD/FIRETEC can be used to examine the effects of vegetation structure on fire behavior. Currently, we are examining these properties by comparing magnitudes and qualitatively correlating the evolution of various properties with each other. In the future, we hope to be able to compare these types of simulations to data from real crown fires. One facet of fire behavior that is easily compared between simulations is the distance that the fire front travels, also known as the spread rate.

Figure 7 shows the downwind propagation distances of the fire from the ignition point. From the data shown in this

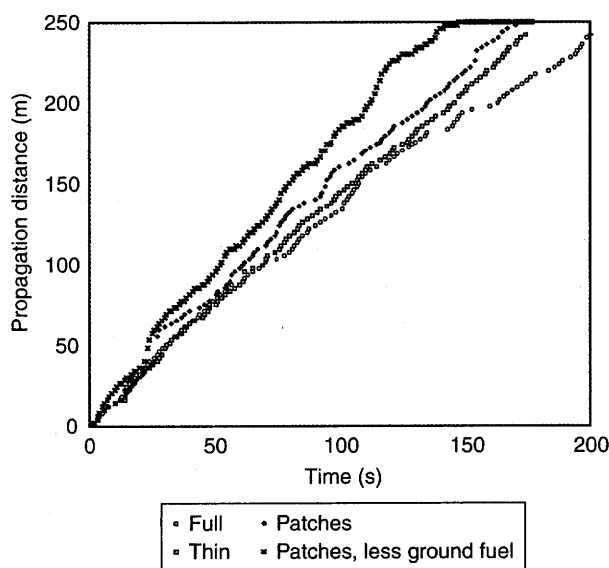


Fig. 7. This figure shows the downwind propagation of the fire front for the four simulations.

Table 1. Summary of the four simulations performed as a part of this study
PLGF, patches, less ground fuel

Simulation	Bulk canopy density (bulk grass density) (kg/m ³)	Canopy description
Full forest	0.24 (0.464)	A canopy that was produced by assembling adjacent 20 m × 50 m regions of trees. Each of these 20 m × 50 m regions includes 81 randomly arranged trees of different shapes and sizes resembling those in a set of data that was measured on a 20 m × 50 m plot near Flagstaff, Arizona.
Thin	0.055 (0.464)	A canopy produced by removing all trees with a trunk diameter at breast height (dbh) less than 28 cm from the canopy used in the full forest simulation.
Patches	0.066 (0.464)	A canopy produced by thinning trees from the canopy used in the full forest simulation to leave patches of uneven-sized trees with spaces in between.
PLGF	0.066 (0.232)	Same as patches, but the grass bulk density has been cut in half.

figure, it is possible to calculate the average rate of spread (slope of the lines shown) of the fires in each simulation. The spread rates range from 1.17 m/s for the full forest simulation to 1.52 m/s for the PLGF simulation. The thin and patches simulations produced mean spread rates of ~ 1.47 m/s and 1.44 m/s respectively. The spread rates of the fires in the simulations with similar ground fuels are grouped closely, while the spread of the fire in the lighter grasses is faster. This indicates that, under this specific range of conditions, the ground fuel density plays a critical role in determining the overall spread rate even when the fire is burning the canopy. It is commonly observed in real fires that if the crown fire outpaces the ground fire, it will stall and 'wait' for the ground fire to catch up. This is consistent with the overall spread rate of a fire being determined by the ground fuel density and composition. We intend to investigate the reason for this apparent lack of sensitivity of the spread rate to canopy structure in future research.

The spread rate of the fires in these simulations is heavily influenced by the wind that blows through the canopy, understory grasses and stem space. Figure 8 shows average vertical profiles of downwind velocity from a 6000 m² domain within the forested regions in the four simulations. These plots contain data derived from spatial averages at an instant in time 6 s after the beginning of the simulations. This particular time was chosen because the winds had had a chance to adjust from the initial uniform wind conditions to reflect the drag caused by the vegetation, but the fire was not yet strongly influencing the flow. The purpose of this figure is to show the average vertical wind shear present in the different fuel distributions in the absence of fire, which is a major difference between the simulations. Once the fire is active in the simulation it dominates many aspects of the wind field (including the average vertical wind shear) and, due to its dynamic nature in our non-steady-state simulation, there is no average state to be discussed. Other aspects of canopy dynamics such as rotational flow are also captured in our simulations (as can be seen in Fig. 9, representing 80 s into a simulation) but are not intended to be illuminated by Fig. 8.

The four plots in Fig. 8 are associated with the full forest, thin, patches and PLGF simulations, and show average downwind velocities for various heights above the ground. In addition, the plots show the mean velocity plus average positive fluctuations and mean velocity less the average negative fluctuations at various heights. The average positive fluctuation is the mean value of all the fluctuations that are greater than zero, and the average negative fluctuation is the mean value of all the fluctuations that are less than zero. These two quantities convey the magnitude of the individual fluctuations above and below the mean, and help differentiate between instances where there are a few large positive or negative fluctuations as opposed to many smaller fluctuations. An example of this can be seen in plots (c) and (d) of Fig. 8, where the negative velocity fluctuations are much

greater than the positive fluctuations in the middle of the overstory fuel layer. We can infer from these values that there are a greater number of positive fluctuations than negative fluctuations. Each of the plots in Fig. 8 also shows the mean bulk fuel density at different heights (measured against the axis on the top of the plots), as well as the mean density plus the average positive density fluctuations and the mean density less the average negative density fluctuations.

The impact of the canopy and understory drag on the winds can be seen by comparing the bulk density values and the average velocities. In all of the plots, the velocity profiles show effects of the canopy, stem space and ground fuel. The stem space velocities, which are nearly the same magnitude as the velocities above the overstory, result from the fact that there are virtually no ladder fuels, the drag of the tree trunks is not accounted for (we are currently working on including this effect), and the sampling area is within 60 m of the leading edge of the forest. At the leading edge of the canopy the flow separates and produces increased velocities above and below the canopy. Because there are no ladder fuels to damp the flow through the stem space, the air funnels into the stem space and is only slowed as the drag effects from the ground fuel and overstory diffuse into the jet of air in the stem space. We would expect the stem space velocities to be less if we had some ladder fuels (even just at the leading edge of the forest) or if we were sampling much farther downwind in the canopy.

When we compare the four plots in Fig. 8, the mean velocities inside the canopy of the full forest simulation are the lowest of the four simulations. The mean velocities of the thin simulation are lower than those of both patches simulations, even though the mean bulk densities are comparable. This difference results from the fact that the scattered large trees in the thin simulation do not provide channels as large as those found between the tree clusters in the patches and PLGF simulations. Also, the scattered nature of the trees in the thin simulation allows the moving air to feel a drag effect from each tree represented, whereas in the patches simulations the clustered nature of the trees shelters some of the trees from the wind. Those sheltered trees therefore do not provide as much drag to the overall flow. The average negative fluctuations in the patches and PLGF simulations are large compared to those of the other simulations. This is, again, because the wind can find channels to quickly move through (shown in Fig. 9), but the clusters form significant obstacles that redirect the wind and limit the flow within them. This also explains why the mean velocity plus the average positive fluctuation is nearly as large as the velocity above the canopy.

The average positive and negative fluctuations are of similar magnitude in the full and thin simulations. This is due to the relatively homogeneous spatial distribution of the trees in these simulations. The lower magnitude of the fluctuations in the full simulation as compared to the thin simulation reflects that the spaces between trees in the full simulation are smaller and the canopy acts more like a continuous

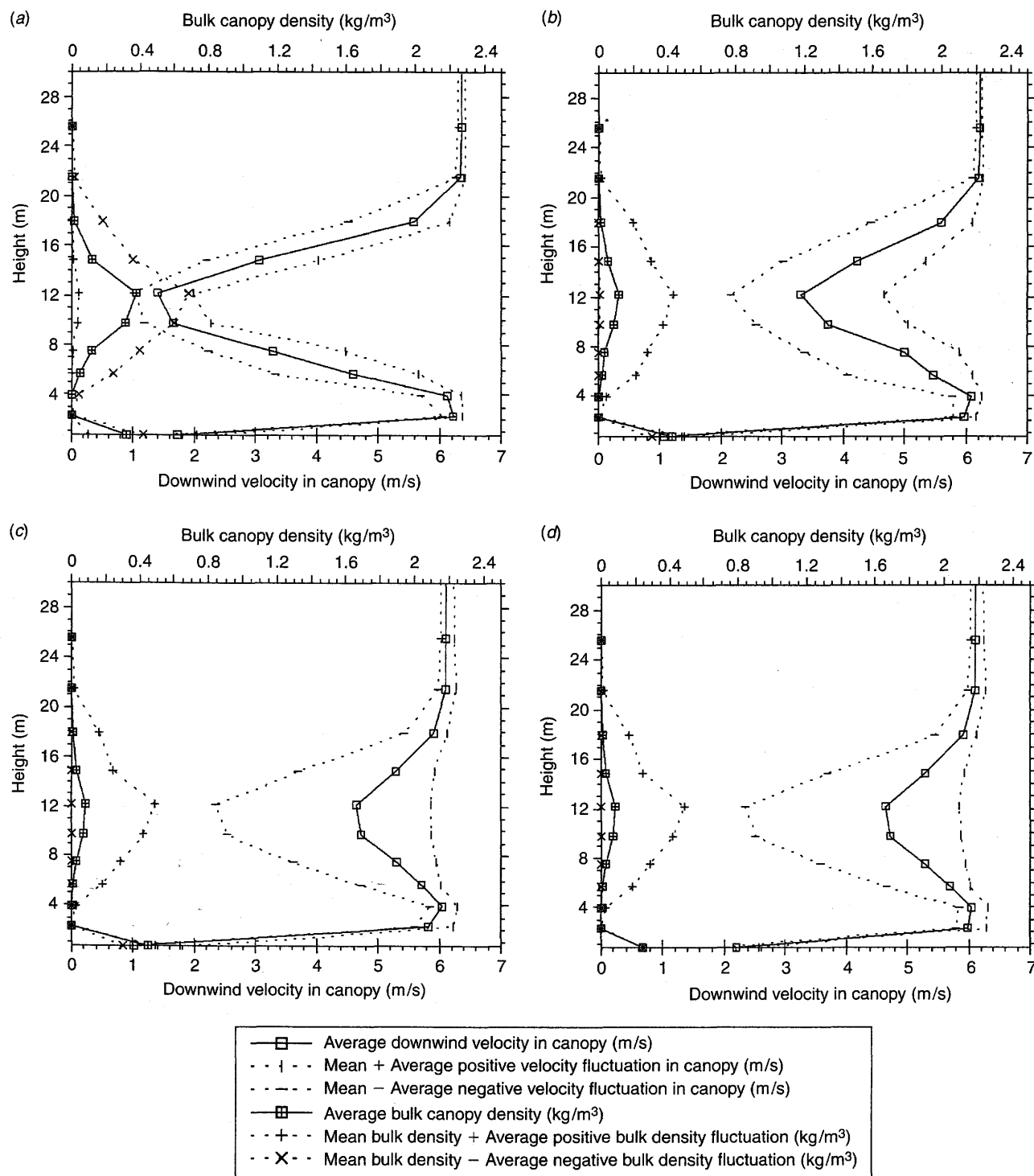


Fig. 8. These four plots associated with the full (a), thin (b), patches (c) and patches, less ground fuel (d) simulations show vertical profiles of average downwind velocities and bulk vegetation densities 6 s after the beginning of the simulations. The plots also show the mean values plus the average positive fluctuations and the mean values minus the average negative fluctuations for the velocities and densities.

homogeneous porous medium, which provides a spatially homogeneous drag. The average bulk ground fuel densities shown in Fig. 8 are of similar magnitude in all of the simulations except PLGF, which is half that found in the patches

simulation. As a result, the average velocity in the grass layer in the PLGF simulation is more than double the average velocity in the grass layer in the patches simulation as seen in Fig. 8c,d. The grass layer velocity in the full simulation is

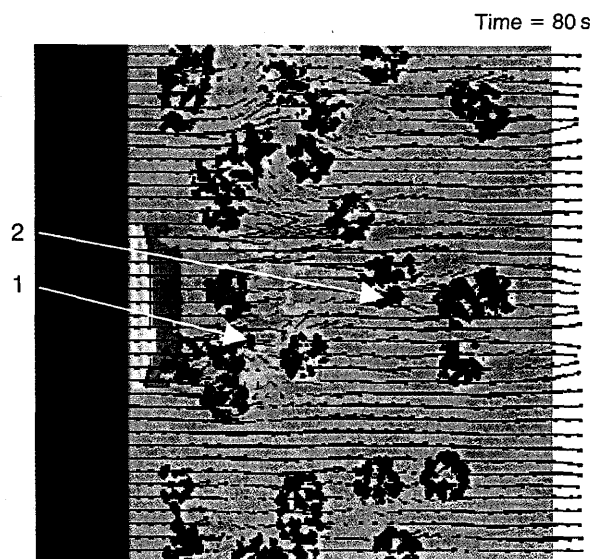


Fig. 9. This image, from the patches simulation, uses vectors to show the flow field 12 m from the ground. The ambient winds at the upper and lower left corners of the image are 6 m/s. The white arrows indicate the two points (trees) that are referred to in the text. The black region is an area where there is no fuel, and the light green areas are the locations of the tall grass. The dark regions are isosurfaces of trees.

significantly larger than in the thin simulation, even though the bulk ground layer vegetation densities are similar. This difference reflects the fact that the full simulation has more litter and less grass than the thin simulation. Thus, the average relative compactness of the ground fuel is greater in the full simulation and causes less drag.

In order to examine the impact of tree arrangement on fire behavior with regard to specific trees, two identical trees at two different specific locations in each of the simulations were chosen. By selecting two identical trees that exist in the same two locations in all four simulations, the impact of location within a non-homogeneous canopy and the impact of various average canopy structures on the way a fire will burn particular trees can be investigated. The trees that were selected have a trunk diameter (at breast height) of 29.4 cm, with a crown radius of 2.75 m. The height of the selected trees is 14.1 m, and the height to live crown is 8.5 m. The locations of these trees are shown in Fig. 9, which is a representation of the patches simulation. The exact same locations were used in the other three simulations, though the trees around them are different in the full forest and thin simulations.

A specific location ~12 m from the ground was chosen in these trees, and the evolution of various characteristics of the fuel, wind and energy transfer at these points is shown in Figs 10–13. The purpose of these figures is to explore and demonstrate the types of information that HIGRAD/FIRETEC can provide, and to incite curiosity about the potential questions that could be addressed with tools such as HIGRAD/FIRETEC. They are meant to

show trends and illustrate interesting differences in the fire–atmosphere–fuel interaction with different canopy structures and at different locations within the canopy, not to show details.

The two graphs in Fig. 10 show the mass loss of the vegetation as a function of time. Each of the graphs is associated with one of the two selected points, and contains eight different sets of data values with interpolated curves drawn through them. These curves show the reduction of bulk density of the wood and the reduction of the water from the vegetation as determined by the mixing-limited burning model and moisture evaporation models in FIRETEC. These plots are included to show the evolution of the wood and water quantities as well as the extent of their depletion at the two locations. In both plots, we can see that the water begins to leave the vegetation at a faster rate than the wood. In the full simulation at point 1, and in the full, thin and patches simulations at point 2, the water is driven out completely. Approximately 60% of the wood mass is lost from the full forest at points 1 and 2, and ~65% in the thin forest. There is less than 20% of the wood lost from either of the points in the patches and PLGF simulations.

The two plots in Fig. 11 show the solid (for this purpose, ‘solid’ includes wood and liquid water) temperature as functions of time. The vegetation at point 1 in the full forest is able to begin burning actively. The temperature reaches nearly 1200 K and remains above 1000 K for approximately 20 s. The temperatures at point 1 in the other simulations never get above 400 K. The temperatures at point 1 in the two patches simulations rise and fall off slowly, while the temperature at point 1 in the thin simulation oscillates at a higher frequency. This oscillation in temperature is believed to be associated with wind oscillations. The temperatures at point 2 in the thin and full forest simulations indicate active burning as they reach temperatures of nearly 1100 K and 1000 K respectively. Point 2 in the patches simulation has a rise in temperature to approximately 475 K, then a decline in temperature and finally a second small rise and fall (peaking at ~350 K). The temperature at point 2 in the PLGF simulation grows slowly to approximately 400 K and declines slowly. These observations are consistent with the trends seen in Fig. 10.

The graphs in Fig. 12 show the bulk convective heating rates as a function of time. In the full simulation at point 1, there is a strong heating period associated with the hot plume being driven over the cooler vegetation, and then a strong cooling period associated with cooler ambient winds being driven over the burning tree after it is ignited. There are smaller amplitude heating and cooling periods for point 1 in the other simulations. The convective heating at point 2 in the PLGF simulation precedes the heating in the other simulations because the ground fire reaches this location faster, as seen in Fig. 7. The convective heating of the vegetation at point 2 in all of the simulations is less than in the full case at point 1. This is probably due to the fact that the temperature

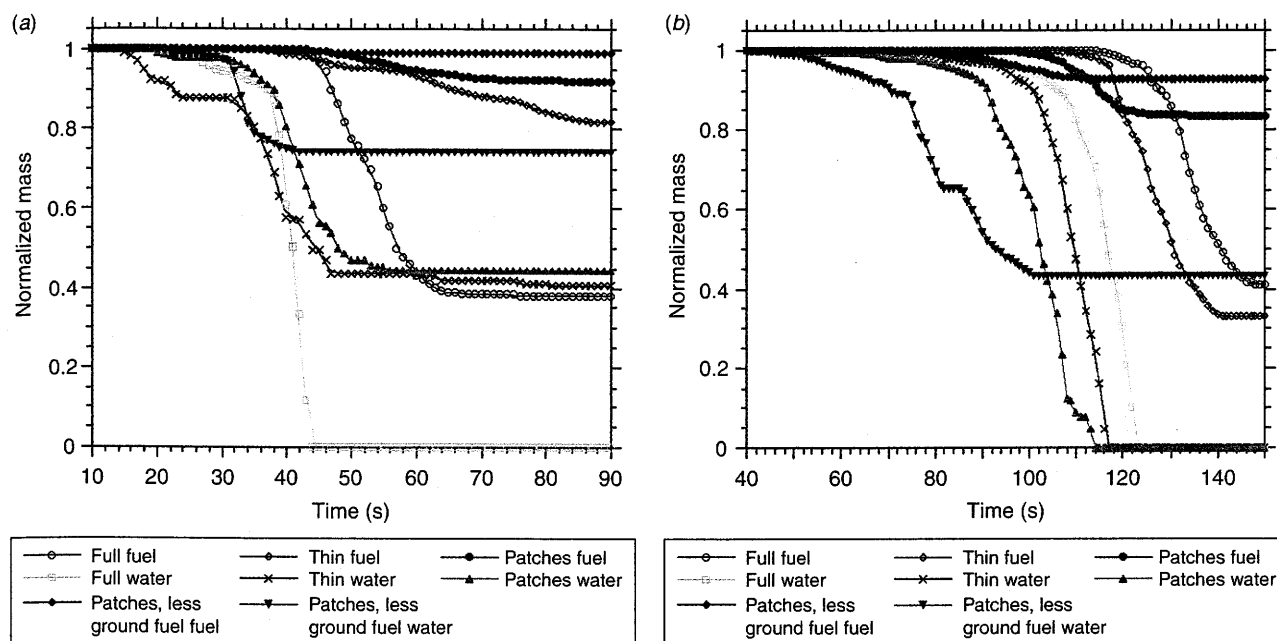


Fig. 10. Plots of vegetation mass loss at points 1 (a) and 2 (b) as a function of time since ignition.

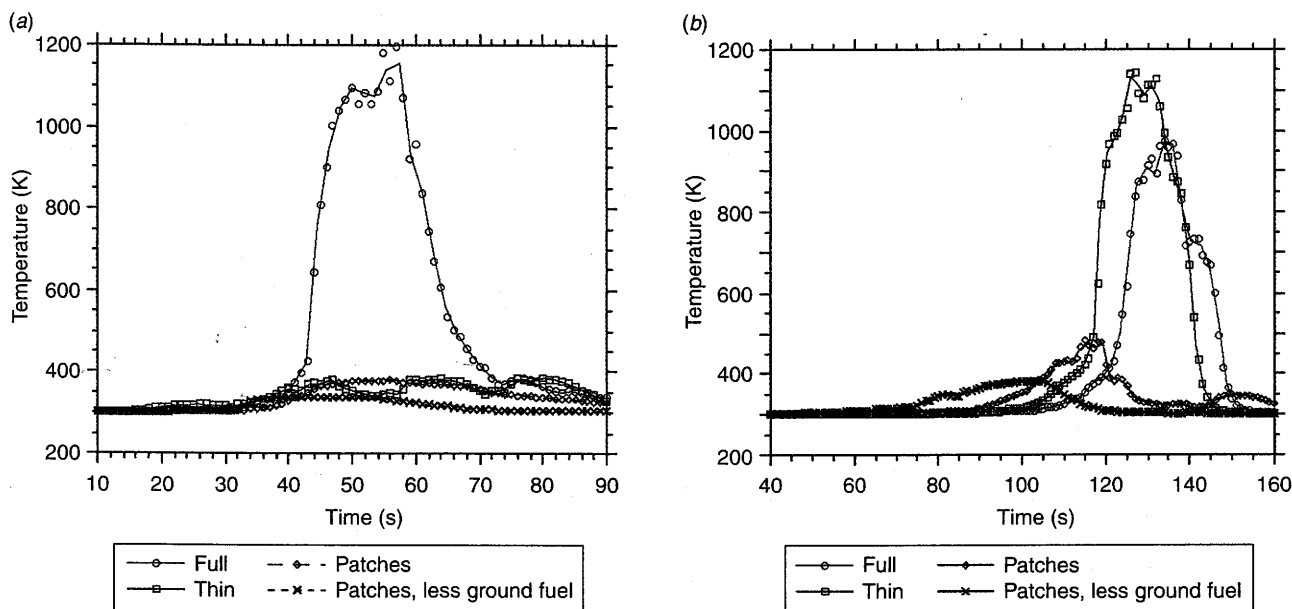


Fig. 11. Plots of the solid temperature at points 1 (a) and 2 (b) as a function of time since ignition.

of the solid rises at nearly the same rate as the air around it. In these simulations, as the plumes from the ground fires pass by point 2 and the cooler ambient air blows on the vegetation, there is a strong convective cooling effect. This contributes to the incomplete burning in the full and thin simulations (seen in Fig. 10) and the intermediate cooling in the patches simulation (seen in Fig. 11).

The final figure, Fig. 13, shows the net bulk radiative heating rate of the solid as a function of time. In this text, the net

bulk radiative heating rate is defined as the energy absorbed by the solid from thermal radiation minus the thermal radiation that is emitted by the solid. The peak heating rate at point 1 in the thin simulation is significantly smaller than the rates for the other simulations at this point. This difference can be attributed to the fact that the trees in the thin simulation are much more dispersed, and so do not have as much radiative heating contribution to each other. The heating rate increases at point 1 faster than it falls off. This indicates that the ground

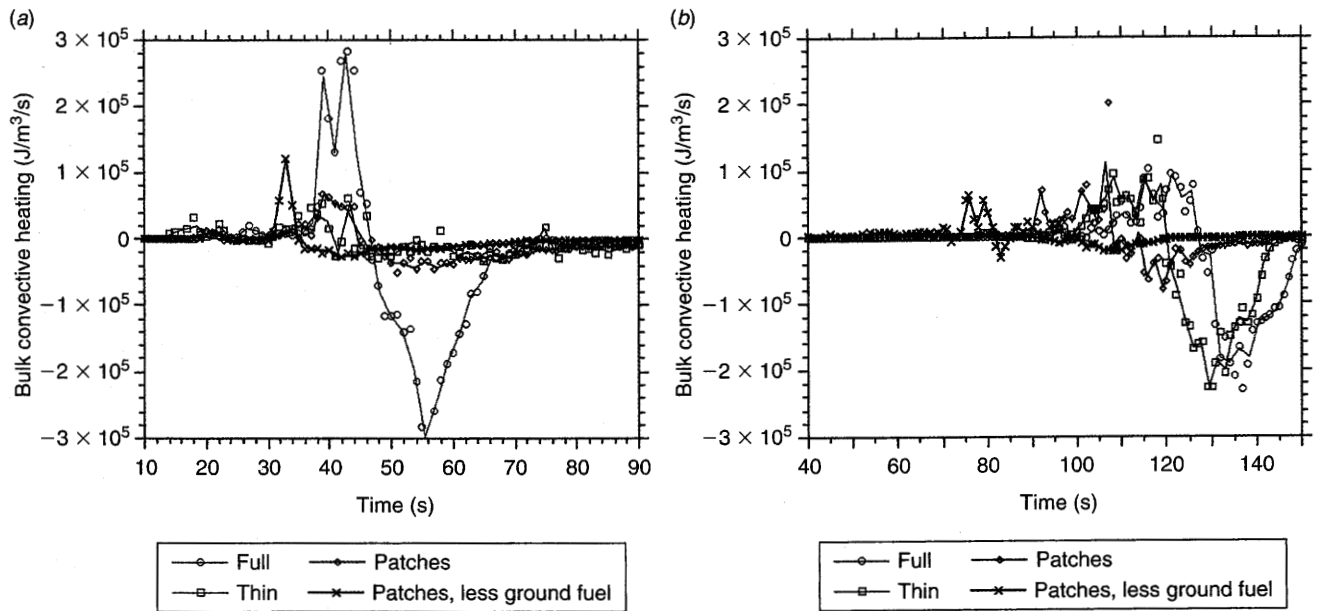


Fig. 12. Plots of the bulk convective heating rate of the solid at points 1 (a) and 2 (b) as a function of time since ignition.

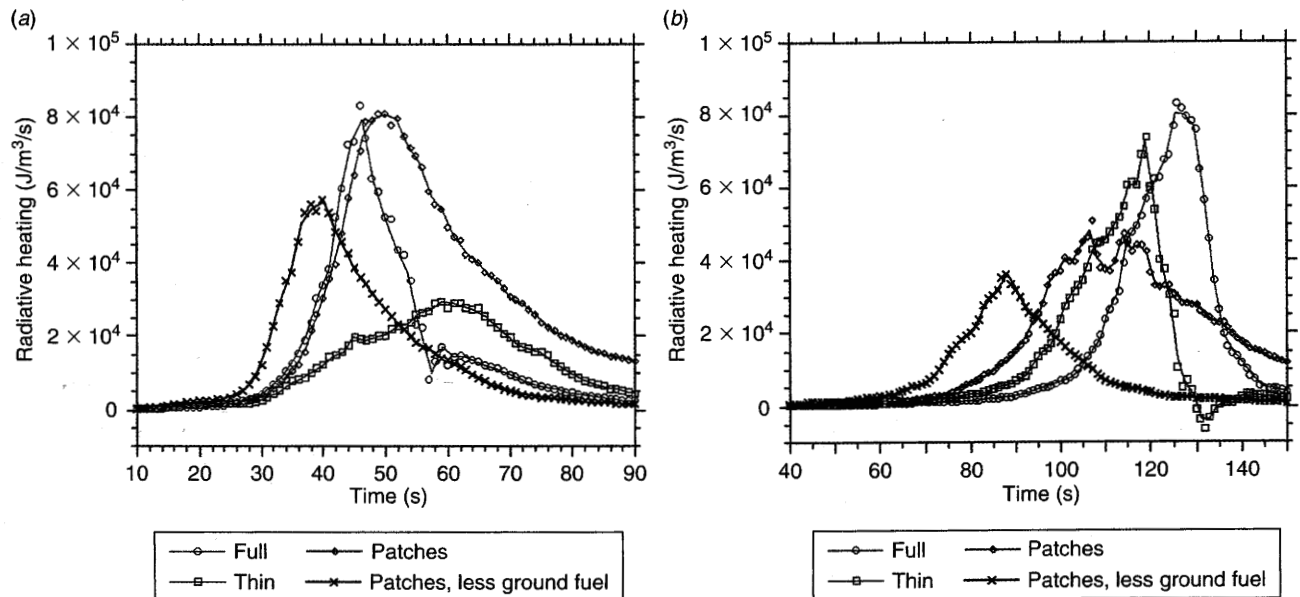


Fig. 13. Plots of the bulk radiative heating rate of the solid at points 1 (a) and 2 (b) as a function of time since ignition.

fires are still increasing in intensity as they approach point 1 but are fully developed at point 2. Another contributing factor could be that the increase in the heating rate is controlled solely by the movement of the fire front whereas the decrease in the radiative heating rate also depends on the slower rate at which the fuels under point 1 cool down or burn out. The net bulk radiative heating rate in the full case at point 2 is similar in magnitude to that at point 1. The heating rate in the thin simulation at point 2 is much higher than at point 1. In this simulation more trees were ignited around point 2 than

around point 1, and so the effect of having the trees dispersed was lessened. The reduction in the radiative heating rate for the two patchy simulations at point 2 could be due to the fact that there are no trees just upwind of the patch of trees where point 2 is located, and so the majority of the trees in the patch itself do not burn significantly. This reduces the tree-to-tree radiation effects. An additional interesting feature shown in Fig. 13 is the negative values in the thin simulation radiative heating rates at point 2. These values are associated with the fact that the solid temperature at point 2 is high, even after

the ground fire has passed. As a result, the tree is actually radiating more energy than it is receiving for a short period of time (potentially igniting other trees). These qualitative descriptions show the effects of different canopy structures on fire behavior. Any one of these examples could be fleshed out in a further series of simulations.

Conclusion

The results described in the previous section illustrate one way that HIGRAD/FIRETEC can be used to compare macroscopic fire behavior (such as propagation distance and spread rate) within different vegetation stands. HIGRAD/FIRETEC was also used to investigate the impact of position or arrangement of individual trees on the chance of a particular tree burning. The simulations provide an illustration of the interaction between the physical processes that drive wildfires and the evolution of the vegetation itself.

The comparison of the PLGF simulation to the patches and other simulations shows that, under some conditions, it is the density of the ground fuel that determines the overall spread rate of the fire even when the overstory is involved in the fire. This might only be true for a select range of wind conditions or canopy densities, so future research will focus on determining under what envelope of conditions this is true.

The simulations show that the position of a tree and the arrangement of the trees around it influence whether that tree will burn or not. By looking at the reported results for the vegetation at point 1 we can see the manner in which the evolution of the physical properties of the vegetation (bulk density, moisture content and temperature) are related to the convective and radiative heating for the various simulations. Clearly, the canopy structure had a significant effect on the significance of the convective heating at point 1 but less effect on the radiative heating. In the full forest simulation the peak convective heating rate is significantly greater than the peak radiative heating rate, whereas the peak radiative and convective heating rates in the other simulations are much more comparable. The more significant convective heating in the full forest case seems to have resulted in the active burning at point 1, which was not present in the other simulations. This is consistent with the idea that the heat is being carried vertically from the ground fire to the canopy, and the effect of the horizontal flow at point 1 is to cool the crown vegetation. In the full forest at point 1, the reduced horizontal winds in the canopy (shown in Fig. 8a) allow for the buoyancy-driven vertical winds to be stronger and concentrate more heat. In the other canopy structures, the increased horizontal wind allows for more entrainment of cool air into the rising air and dampens the convective heating of the crown vegetation.

At point 2, the peak convective and radiative heating rates are comparable in all of the simulations. The convective heating rate in the full forest simulation at point 2 is much lower than at point 1. This is due to a combination of events including

the reduction of upwind canopy fuels allowing more horizontal wind to reach point 2. Another factor is the longer period of radiative preheating that occurs at point 2 because of the additional radiative contributions of the trees that are already burning. The longer duration of radiative preheating reduces the temperature difference between the solid vegetation and the hot air when it reaches point 2. The comparison of the contribution of convective heating at point 1 and point 2 shows that there are different heat transfer balances occurring at different times and different locations, especially with regard to fires in non-uniform fuel beds such as those being simulated. In fact, this type of study could shed additional insight into the difference between the processes that sustain an existing crown fire as opposed to those required for the transition from a ground fire to a crown fire (point 2 v. point 1).

Another significant difference is the active burning at point 2 in the thin simulation, which is not present at point 1. This difference appears to be tied to the higher rate of radiation heat transfer at point 2. This larger rate has been attributed to the additional strength in the ground fire as well as additional trees on fire near point 2. Any increase in the number of radiation sources can impact point 2 in the thin simulation more strongly than it does in the full forest simulation because of the longer effective optical path lengths in the thin canopy. There are fewer trees to shield the radiation from reaching point 2. In the patches simulation the radiative heating begins sooner because there is a wide unobscured space upwind of point 2, but the peak radiative-heating rate is smaller because there are no burning trees just upwind of point 2. This result shows the combined impact of the location with respect to the ignition point and the bulk and local properties of the canopy.

These are a few of the observations and bits of qualitative insight that can be gained from this set of four simulations. Without a significant amount of validation and comparison with experiments, it is very premature to put emphasis on the actual values produced by these types of simulations. They do, however, provide a valuable tool for looking at the interactions between the physical processes that influence wildfires and the fire's behavior. With sufficient validation, they could eventually play a roll in helping managers to decide between different thinning and fuel treatment strategies at specific locations.

Summary

In this text, we have introduced the use of HIGRAD/FIRETEC for simulating fires as they move through forests made of individual treelike fuel elements. There are many ways to approximate the distribution of ground fuels under the canopies. We chose a simplified algorithm for this text but anticipate the inclusion of more sophisticated algorithms in the future. We hope to draw on the experience of Forest Service researchers to help inform the development of this

algorithm. HIGRAD/FIRETEC is not a high fidelity combustion model, and does not attempt to capture the details of the combustion process or the fine-scale details of the vegetation. HIGRAD/FIRETEC does, however, model the physics of many of the important processes that drive a wild-fire, and captures many of the complex interactions between fire, atmosphere and fuel.

To illustrate the use of this model for simulating forest canopies, we chose two different trees that each existed identically in four different simulations. By examining a set of characteristics of the atmosphere and solid fuel at a location 12 m above the ground in each of these trees in the various simulations, we can see indications of different phenomena occurring as a result of a tree's location with respect to the ignition point, other trees, and the understory vegetation.

The number of simulations presented here is not sufficient to provide conclusive results about all of the interdependencies that occur in the simulations, but they are sufficient to inspire many new questions and to guide future research directions.

In order for information from this type of model to be of true value, we must find ways to compare the results with experiments and real fires. By making these comparisons, we can gain confidence in the aspects of the model that are working correctly and obtain critical information about the deficiencies of the model that need improvement. By recognizing differences between the model and real fire behavior, researchers will learn more about the interaction between the driving processes in wildfires.

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Simulation of the Big Elk Fire using coupled atmosphere–fire modeling

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Abstract. Models that simulate wildland fires span a vast range of complexity; the most physically complex present a difficult supercomputing challenge that cannot be solved fast enough to become a forecasting tool. Coupled atmosphere–fire model simulations of the Big Elk Fire, a wildfire that occurred in the Colorado Front Range during 2002, are used to explore whether some factors that make simulations more computationally demanding (such as coupling between the fire and the atmosphere and fine atmospheric model resolution) are needed to capture wildland fire parameters of interest such as fire perimeter growth. In addition to a Control simulation, other simulations remove the feedback to the atmospheric dynamics and use increasingly coarse atmospheric resolution, including some that can be computed in faster than real time on a single processor. These simulations show that, although the feedback between the fire and atmosphere must be included to capture accurately the shape of the fire, the simulations with relatively coarse atmospheric resolution (grid spacing 100–500 m) can qualitatively capture fire growth and behavior such as surface and crown fire spread and smoke transport. A comparison of the computational performance of the model configured at these different spatial resolutions shows that these can be performed faster than real time on a single computer processor. Thus, although this model still requires rigorous testing over a wide range of fire incidents, it is computationally possible to use models that can capture more complex fire behavior (such as rapid changes in intensity, large fire whirls, and interactions between fire, weather, and topography) than those used currently in the field and meet a faster-than-real-time operational constraint.

Introduction

Wildland fire modeling encompasses a wide range of complexity. The simplest conceptual models are based on experience and intuition from past fires and can be used to anticipate imminent fire behavior. At the next level of complexity are tools such as BEHAVE (Andrews 1986) and BEHAVE Plus (Andrews *et al.* 2003), which are based on the semi-empirical Rothermel (1972) equations for fire spread rate, flame length and intensity. These equations are founded on the work of Fons (1946) and Emmons (1963) and formulate the quasi-steady equilibrium spread rate for a surface fire on flat ground in no-wind conditions. They are calibrated using data from piles of sticks burned in a flame chamber/wind tunnel to represent other wind and slope conditions for the fuel complexes tested. Similar relationships have been developed for the Canadian Fire Behavior Prediction system (Forestry Canada Fire Danger Group 1992) and others based on Australasian fuel complexes (Noble *et al.* 1980; Cheney *et al.* 1993; Catchpole *et al.* 1999) for quick estimation of fundamental parameters of interest such as fire spread rate, flame length and fireline intensity of surface fires at a point for specific fuel complexes, assuming a representative

point-location wind and terrain slope. Asenio and Ferragut (2002) presented a two-dimensional (horizontal and vertical direction) combustion model based on conservation laws that include radiation, convection, pyrolysis and reactions of volatile gases. Its limitations are that it currently lacks coupling with the Navier-Stokes equations for modeling fluid motion and that parameters must be calibrated with data.

Because the shape of the fireline does not come from the physics of these models, certain assumptions must be made to shape the fire growth. Richards (1994, 1995) and Richards and Bryce (1995) developed a set of equations that describe the propagation of a fire front assuming an elliptical shape that expands based on Huygens' Principle of wave propagation and can be used to represent the shape of the spreading fire. Two-dimensional fire growth models such as FARSITE (Finney 1998) and Prometheus, the Canadian wildland fire growth model designed to work in Canadian fuel complexes, have been developed that use Huygens' Principle and apply semi-empirical spread rate formulas such as those of Rothermel (1972) and other relationships regarding ground-to-crown transitions to calculate fire spread and other parameters along the surface.

Although more sophisticated applications (Roe *et al.* 2001; Fujioka 2002; Jones *et al.* 2003; Zeller *et al.* 2003) used a three-dimensional numerical weather prediction system to provide inputs such as wind velocity to a fire growth model, this was a passive input and the feedback of the fire on the atmospheric wind and humidity were not accounted for.

More complex physical models join computational fluid dynamics or numerical weather prediction models with a wildland fire component and allow the fire to feed back on the atmosphere. The atmospheric models solve the prognostic Navier-Stokes equations of fluid motion for two- or three-dimensional components of momentum, a continuity equation to assure fluid mass is conserved, a prognostic thermodynamic equation, and may include equations for cloud microphysical quantities, as well as a wildland fire component that represents either the spread rate and intensity of the fire using parameterized empirical formulas or some parameterization of the combustion processes in the fire. These models include Clark *et al.* (1996a, 1996b) Los Alamos FIRETEC, (Linn *et al.* 2002), the Fire Dynamics Simulator (Evans *et al.* 2003; McGrattan 2004) and, to some degree, the two-dimensional model PIF97 (Dupuy and Larini 1999; Porterie *et al.* 2000). These tools have different emphases and have been applied to better understand the fundamental aspects of fire behavior, such as effect of fuel inhomogeneities on fire behavior (Linn *et al.* 2002), feedbacks between the fire and the atmospheric environment as the basis for the universal fire shape (Coen *et al.* 2001; Clark *et al.* 2004), and fire spread at the community scale (Evans *et al.* 2003). The cost of added physical complexity is a dramatic increase in computational cost, so much so that a full three-dimensional explicit treatment of combustion in wildland fuels (so-called direct numerical simulation (DNS)) at scales relevant for atmospheric modeling does not exist and is beyond current supercomputers. So, even these more complex models parameterize the fire in some way—for example, Clark *et al.* (1996a, 1996b) use the Rothermel (1972) equations to calculate local fire spread rates using fire-modified local winds and, although FIRETEC and FDS carry prognostic conservation equations for the reacting fuel and oxygen concentrations, the computational grid cannot be fine enough to resolve the reaction rate-limiting mixing of fuel and oxygen, so approximations must be made concerning the subgrid-scale temperature distribution or the combustion reaction rates themselves. The limitations on fire modeling are not solely computational. At this level of complexity, the models encounter limits in knowledge about the composition of pyrolysis products and reaction pathways, in addition to gaps in basic understanding about some aspects of fire behavior such as fire spread in live fuels and surface-to-crown fire transition.

Thus, while more complex models have value in studying fire behavior and testing fire spread in a range of scenarios as research tools, FARSITE and Palm-based applications of BEHAVE have shown great utility as practical in-the-field

tools because of their ability to provide estimates of fire behavior in real time. While the coupled fire-atmosphere models can represent the ability of the fire to affect its own local weather and many aspects of the explosive, unsteady nature of fires and extreme fire behavior that cannot be incorporated in current tools, it remains a challenge to apply these more complex models in a faster-than-real-time operational environment, primarily because fine resolution is assumed to be necessary to better capture the physical processes of the fire. Also, although they have reached a certain degree of realism when simulating specific natural fires, they must yet address issues such as identifying what specific, relevant operational information they could provide beyond current tools, how the simulation time could fit the operational time frame for decisions (i.e. they must run substantially faster than real time for real-sized problems), what temporal and spatial resolution must be used by the model, and how they will estimate the inherent uncertainty in numerical weather prediction in their forecast. If these models are to go beyond use for scientific understanding to application in an operational setting, these 'operational constraints' must be used to steer model development.

This work addresses two primary factors that increase the physical complexity and the computational requirements of coupled atmosphere-fire models, specifically the feedback between the fire and the atmosphere and the atmospheric grid resolution. It describes the numerical model and its input, application to a sample incident, results and computational performance over a range of atmospheric grid resolutions, and the implications for doing this type of simulation in an operational (and therefore faster-than-real-time) manner, posing questions and issues that can be used to guide future work.

Weather and fire behavior

The Big Elk Fire ignited on 17 July 2002 at 1449 LST near Pinewood Springs, Colorado (ignition point: 40°17'2"N, 105°22'59"W) and burned in heavy timber in rough terrain, with daytime temperatures 27–32°C in historic extremes of low fuel moisture conditions due to prolonged drought. On 17 July, the fire burned aggressively throughout the afternoon into the evening in closed canopy conifer on the eastern- and south-eastern-facing slopes, with steep slopes of 50 to over 70%. Fire behavior included intense surface fire runs, active backing, torching and sustained crown fire runs upslope to the north-west toward Kenny Mountain at the fire's head. Environmental winds were light and slope-driven with light upslope winds of 2–3 m s⁻¹ by day and weak downslope winds at night. Minimum relative humidities of 10–15% were common for the first 5 days of the fire. The fire behavior subsided at night, although relative humidity recovery was weak. Fire behavior was similar on 18 July 2002 (Fig. 1) with intense crown runs and spotting 400–600 m. Much of the downwind woody fuel was burned completely. A cool, moist air ma-

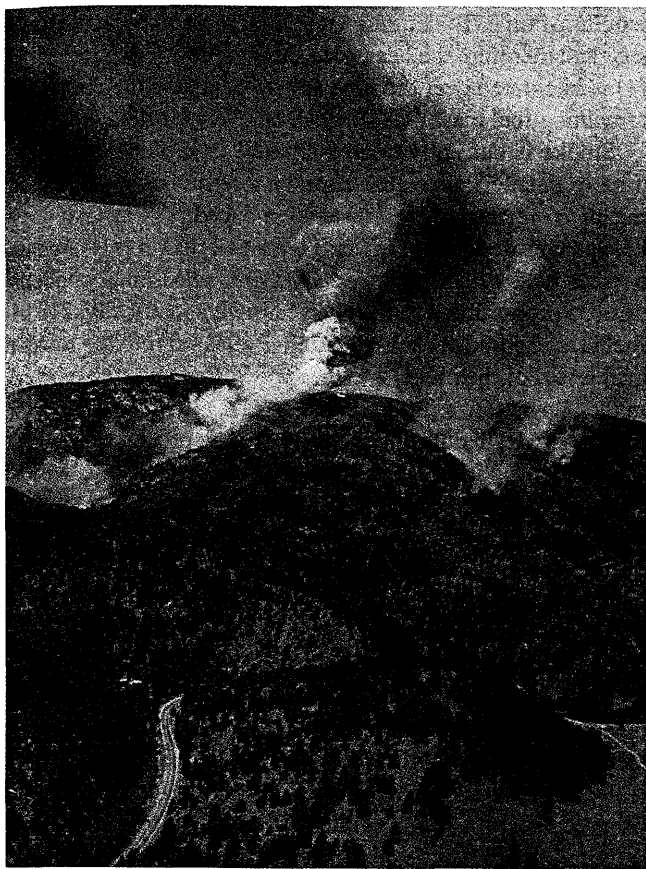


Fig. 1. Photo of the Big Elk Fire at 1057 LST on 18 July 2002, showing the terrain, fuel distribution and fire head as it crests Kenny Mountain. The photographer is facing north-north-west. (Photo courtesy of Kelly Close (Close 2002).)

arrived 20–21 July, slowing fire growth. When contained, area burned was estimated at 1780 ha.

Numerical model

The National Center for Atmospheric Research's coupled atmosphere–fire model is described in detail in Clark *et al.* (1996a, 1996b, 2004). An atmospheric prediction model has been coupled with an empirical fire spread model such that sensible and latent heat fluxes from the fire feed back to the atmosphere and affects the temperature, pressure and, subsequently, the winds around the fire, while these modified winds around the fire in turn drive the fire propagation. This wildfire simulation model can thus represent the complex interactions between a fire and local winds.

Atmospheric model component

The meteorological model (Clark 1977, 1979; Clark and Hall 1991, 1996) is a three-dimensional non-hydrostatic numerical model based on the Navier-Stokes equations for fluid momentum, a thermodynamic equation, and a continuity equation that assures the conservation of mass. Vertically

stretched terrain-following coordinates are used to improve airflow over mountainous topography. Large-scale atmospheric motions (like cold fronts, low pressure systems and pressure troughs) can be introduced to the modeling domain using any large-scale gridded atmospheric analysis data or forecast to initialize the outer domain and update lateral boundary conditions. Two-way interactive nested modeling domains telescope down to the meter-sized fine dynamic scales of vortices in the fireline through horizontal and vertical grid refinement. The increased interception of solar heating by slopes oriented toward the sun is accounted for using the scheme developed by Clark and Gall (1982). Cloud microphysical processes are treated using a bulk two-species (cloud droplets and rain) parameterization for liquid hydrometeors and a three-category (ice crystal, pristine snow and graupel/hail) parameterization for ice particles.

Fire model component

The wildland fire component of the model consists of surface fire, crown fire, the gradual consumption of fuel complexes once ignited, and the release of heat, water vapor and smoke fluxes into the atmosphere. Local fire spread rates depend on the modeled wind components through an application of the Rothermel (1972) fire spread rate formula or, for Australian fuel complexes, those presented by Noble *et al.* (1980). A BURNUP-type algorithm (Albini 1994) characterizes the consumption of fuel mass with time. At the earth's surface, each atmospheric grid cell is subdivided into fuel cells. Four tracers, assigned to each fuel cell, define a quadrilateral that identifies burning areas of fuel cells and defines the fire front, the interface between burning and unignited fuel. A contour advection scheme assures fireline consistency between cells and avoids any ghosting effects (Richards 1994).

The surface fire may dry and then, if a threshold heat flux is exceeded, ignite the canopy above, if present. The surface fire and canopy fire are currently collocated—the model does not currently allow for a running crown fire, where the crown fire spreads independently of a surface fire. A simple treatment that approximates the distribution of heat into the atmosphere by radiation is used; it characterizes the effect of the fire on the atmosphere over the fire using an exponential decay with height, distributing the sensible and latent heat fluxes from both the surface and crown fire over an e-folding length corresponding to the lowest grid points of the atmosphere. For the canopy fire, the heat fluxes are released at and above the canopy height. The e-folding length used here is 50 m; previous work by Clark *et al.* (1996a) shows that the model results are not very sensitive to this parameter. (Also, infrared imagery observations of crown fires (Coen *et al.* 2004) estimate the e-folding length to be about 50 m.) This parameterized treatment of surface fire spread and canopy ignition removes the need for explicitly modeling the radiative transfer that would otherwise be needed to represent the heat transfer from the fire to the fuels and atmosphere. These

heat fluxes modify the temperature, pressure, and the winds around the fire. These modified winds are, in turn, used in the fire spread formula. The fire component is discussed in more detail in Clark *et al.* (2004).

Nelson (2002) and McAlpine *et al.* (1991) presented methods for determining a representative wind speed for input to fire growth models. Such models often employ a 'mid-flame wind speed', U . In coupled atmosphere–fire modelling, it is not practicable to identify U because the mid-flame velocities are primarily vertical, so $U \approx 0$. A single background wind cannot adequately represent the environment as the fire dramatically alters the winds in its environment. Using fine grids and interpolation, one may calculate the wind speed normal to the fireline very close to the line itself; however, because the fireline is a point of convergence of winds from ahead and behind the fire, the horizontal wind is effectively zero. Thus, the user must select a distance behind the fireline (along a line normal to the local fireline front) and a height at which wind speeds spread rate calculation will be taken; here, 2 m behind the line and fuel height respectively. As noted earlier, the Rothermel (1972) formulas for surface fire spread rate that this model and other tools use are calibrated using laboratory experiments that cover a limited parameter space. The sensitivity of the simulated behavior to these parameters reflects the range of experiments used in the development of the model. In particular, extrapolation of the formula to high wind speeds, for example, produces unnatural fire spread rates. Thus, like other applications of this formula, the maximum wind speed input is capped, here, at 10 m s^{-1} .

Having calculated the flaming front spread rate, the rate at which fuel is consumed once ignited is described using a mass loss parameterization, where the mass remaining as a function of time was assumed to decrease exponentially, an approximation to the general mass loss curves produced by the BURNUP algorithm, according to the formula:

$$1 - F = \exp\left(\frac{-t}{W}\right), \quad (1)$$

where F is the fraction of fuel that has been burned, t is time since ignition, and W is a weighting factor assumed representative of each fuel model that determines how fast the fuel mass is consumed. Equation (1) means that recently ignited fuel loses mass rapidly, while later the fractional mass loss rate is slower. Flashy fuels such as grass are characterized by W on the order of 10, corresponding to their rapid burn-out time, while fuel complexes composed of larger fuel elements are better characterized by W on the order of 500, indicating a prolonged heat release. The fire 'goes out' when the mass remaining after ignition approaches an infinitesimally small number and the heat flux from that fuel grid cell approaches zero.

In each time step, the fireline (the interface between burning and unignited fuel) spreads partially through a fuel cell, so locations within the cell will have been burning different

lengths of time. To determine the total mass loss over a time step and therefore the heat released into the atmosphere, eqn (1) is integrated over the time step to calculate the currently remaining fuel mass (see Clark *et al.* 2004 for details).

Smoke is currently treated crudely, as a placeholder for a more detailed, speciated emissions representation, and is meant to indicate where smoke is produced, how thick it may be, and how deeply and where it is transported. It is treated as a tracer field with no fall velocity that is released at concentrations proportional to the mass burned; specifically, 2% of fuel mass burned at each time step is released into the lowest grid levels of the model, whence it is lofted and transported in the buoyant, fire-modified winds.

This model has the potential to represent processes beyond what other models that passively ingest atmospheric winds can. The feedback between the fire and atmosphere makes it possible to evaluate the cause of rapid changes of wind in the fire environment, the formation of large, burning fire whirls like those that occurred during the Missionary Ridge fire of 2002, and blow-ups. However, the possibility of these feedbacks raises questions. Must models capture the two-way feedback between a fire and the atmosphere to simulate fire behavior accurately? If the fire feedbacks do cause local perturbations in the environment, how fine a spatial resolution is necessary to model fires? Still, some important stochastic processes governing wildland fires such as long-range spotting cannot be captured deterministically by this or any other model; they may be represented only in a probabilistic sense. No representation of long-range spotting is currently included.

Numerical experiment design

This section describes the input for simulations of periods during the Big Elk Fire. This case is used as a demonstration of this process. The three environmental factors that influence fire behavior are the primary inputs to the model.

Topography

Three arc second North America topography data are used. Terrain in this location is steep with slopes of 70%. The location of domains and local topography is shown in Fig. 2

Weather

The large-scale atmospheric environment can be introduced into the model from either a single atmospheric upper air sounding or three-dimensional gridded large-scale mode data (either the analyses for a post-incident study, or the forecast from a meso- or synoptic-scale numerical weather prediction model for predictions). In this case, a locally run daily MM5 forecast (<http://rain.mmm.ucar.edu>) for the period from 17 July 2002 1200 UTC to 18 July 2002 1800 UTC is used to initialize the finer-scale atmosphere–fire model. The 3-hourly output of the 10 km MM5 modeling

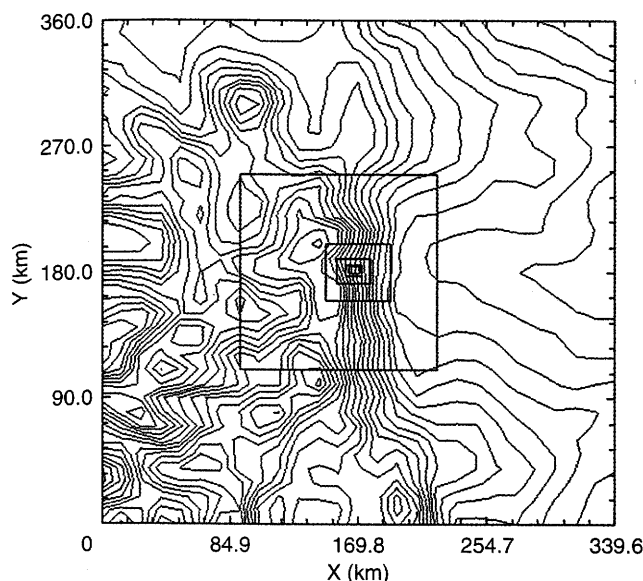


Fig. 2. Location of the six nested domains. Contours show model topography every 160 m.

domain is used to initialize the outermost computational domain of the coupled atmosphere–fire model and update the boundary conditions, thereby introducing mesoscale and synoptic scale features into the simulation. A surface fire is then ignited in the coupled atmosphere–fire model at 1449 LST at the observed ignition point, and the evolving atmosphere and fire behavior simulated until 2200 LST on 17 July. The heat from both surface and crown fires is deposited into the atmosphere over a 50 m extinction depth. Previous studies (Clark *et al.* 1996a) show little sensitivity to this parameter. Solar heating plays an important role in local circulations in this area, with sensible heat fluxes on clear days of 400 W m^{-2} . In the model, the surface experiences a heating due to solar radiation depending on its orientation relative to the sun, thus east-facing slopes warm in the rising sun.

Fuel

At the surface, atmospheric grid cells are horizontally subdivided into fuel cells. The fuel characteristics required for the Rothmel (1972) surface fire growth equations are required for each cell, include the surface fuel model, fuel load, fuel depth, surface area-to-volume ratio, fuel moisture content of extinction, fuel density, mineral content and the heat content, as well as the canopy height. In contrast with Clark *et al.* (2004) and previous applications of this model, the surface fuel category and canopy coverage if any varies spatially from fuel cell to fuel cell. Currently, for simplicity but also because of the lack of quantitative data on the spatial variability of fuel loading, the set of fuel characteristics is assigned the most appropriate of the 13 Anderson (1982) fuel models for surface fire behavior. The fuel models are mapped based on the

land cover type data and vegetation genera of the Colorado GAP analysis (Thompson *et al.* 1996), which is based primarily on Landsat Thematic Mapper images (30-m pixels), with guidance from the incident report (Close 2002). Thirty-meter grids were used for this study. A thorough discussion of the technique used for deriving fuel models and the atmosphere–fire model sensitivity to spatial variability in the fuel models will be presented separately. As discussed by Sandberg *et al.* (2001), these stylized fuel models were meant to approximately represent natural fuel bed properties, and accurate quantitative fuel properties must await databases and fuel classification systems such as those of Ottmar *et al.* (2003).

Fuel characteristics used in this experiment are given in Table 1. For this Colorado fire, the typical surface Anderson fuel models are grass with understory (category 2) and the timber litter categories 8 and 10. Canopy loads are estimated based on the tree genera. The process for developing the fuel characteristics used here and the sensitivity of results to the fuel characteristics will be described elsewhere.

The fuel moisture treatment also differs from Clark *et al.* (2004) and other uses of this model. Rather than applying a fuel moisture that remains constant throughout the simulation, but lacking specific onsite fuel moisture measurements, this simulation represents temporal variability in the surface fuel moisture $m(t)$ using an approximation to the diurnal variability illustrated by Viney and Catchpole (1991):

$$m(t) = \bar{m} + B \sin[\omega(t - t_o)], \quad (2)$$

where $\bar{m}$ is the average daily fuel moisture (in percent of oven dry mass), B is the amplitude of the daily variability, $\omega = \pi/12 \text{ h}^{-1}$, and t_o represents the displacement of the fuel moisture cycle from the local 24-h clock. Due to the extremely dry conditions, lack of precipitation or condensation, or other complicating atmospheric impacts, this approximates the variability due to the primary process, the adsorption and desorption of water vapor from the fuel. Based on local surface stations recording maximum relative humidities at approximately 0330 LST, minimum relative humidities at 1530 LST, and assuming no significant lag in this representative fuel moisture, $t_o = 21.5 \text{ h}$. The variables $\bar{m}$ and B are set to 7% and 2, respectively, producing a range of fuel moisture that peaks at 9% at 0300 LST and reaches a minimum of 5% at 1530. For comparison, estimates of 1000-h dead fuel moisture are given in the incident report for this period as 8–11%, with live fuel moistures given as 95–103%.

Development of a generalized treatment for fuel moisture that responds to the modeled atmospheric conditions including temperature, humidity, precipitation and ventilation by winds is in progress. Ultimately, the fuel categories, fuel loading and state (fuel moisture) must be derived using a combination of remote sensing techniques (e.g. Roberts *et al.* 1999), surface sensor networks and *in situ* ground data to capture quantitatively their spatial and temporal variability.

Table 1. Some characteristics assigned to the fuel categories in this experiment

Surface fuel category	Surface fuel mass (kg m^{-2})	Surface area-to-volume ratio (m^{-1})	Mass loss weighting factor, W (s^{-1})	Fuel depth (m)	Canopy fuel mass (kg m^{-2})	Canopy: initial moisture/dry mass ratio (%)
2: Timber (grass and understory)	0.897	2784	7	0.305	0	—
8: Timber–light litter	1.121	1889	900	0.061	0.121	100
10: Timber–heavy litter	2.694	1764	900	0.305	0.121	100

Table 2. Summary of numerical experiments for the 17 July 2002 Big Elk Fire

All experiments use the same MM5 forecast for large-scale data initialization and fuel category based on the primary vegetation genera from a 50-m gridded Colorado GAP land cover database. Column heads are: N , number of nested domains; Spacing, horizontal grid spacing of finest domain; Height, height of lowest vertical grid point above ground; Feedback?, does fire feed back into atmospheric dynamics?

Experiment	N	Spacing (m)	Height (m)	Feedback?
Control	6	41	18	Yes
No feedback	6	41	18	No
3DOM	3	1111	83	Yes
4DOM	4	370	37	Yes
5DOM	5	123	18	Yes

Simulation configuration

These idealized experiments examine a fire simulated using 4–6 nested domains. The outermost domain has 10 km horizontal grid spacing, corresponding to the MM5 domain that is used to initialize it (28×38 ; 50 vertical grid points), while inner domains nest down at a 3 : 1 nesting ratio giving domain 4 a horizontal grid spacing of 370 m (50×50 ; 52 vertical grid points), and domain 6 a grid spacing of 41 m (128×128 ; 40 vertical grid points) (Table 2). The stretched vertical grid is also nested providing finer resolution in inner domains.

In addition to simulating the fine-scale meteorology and fire behavior in the early periods of the Big Elk fire, several sensitivity tests were performed (Table 2). First, to determine the resolution needed for simulating wildfires, experiments varied the finest atmospheric grid resolution. This was done by running simulations with three (finest horizontal grid spacing: 1111 m) to six (finest horizontal grid spacing: 41 m) inner domains. The fire was contained in the innermost domain of each simulation, which had 15×15 fuel cells within each atmospheric grid volume. Coarser resolutions require fewer grid points and can be run much faster, but do not resolve fine-scale fireline features, which may be an important part of fire propagation (Coen *et al.* 2004).

Second, experiments neglected the feedback of heat and water vapor from the fire into the atmospheric model to test whether it is necessary to include the atmosphere–fire coupling that creates locally strong winds. This feedback is

what allows fires to ‘create their own weather’. Without this feedback, the fire is purely driven by ambient atmospheric winds and is essentially what fire spread calculations made using mesoscale model winds (e.g. Zeller *et al.* 2003) would produce if they could be run at these fine spatial scales. Comparing this No-feedback simulation with the Control simulation forecasts (which do not include the fire) shows whether models may err significantly by not including this effect.

Results

Several frames from the simulation of the first 7 h of the Big Elk Fire using six domains (experiment Control in Table 2) are shown in Fig. 3a–d. The actual fire progression is shown in Fig. 4. The general progression of events, captured in the simulation, was the production of upslope winds due to solar heating of the mountain slopes (Fig. 3a) and ignition of fire in the valley (Fig. 3b). The surface fire dried the canopy and, due to the intense burning and surface fuel loads, produced a large enough heat flux to ignite the canopy overhead. The fire spread rapidly upslope to the north-west (Fig. 3c) under the influence of upslope winds and fire-induced convective winds. The fire spread slower later in the day as fuel moisture increased, as winds subsided, and as the fire reached gentler slopes near the top of Kenny Mountain (Fig. 3d). It is important to emphasize that the simulated fire is driven by the small-scale, solar-heating-driven winds (weak 3 m s^{-1} upslope) and the convective heating produced by the fire. Mesoscale models do not capture this valley flow, as the valley itself is narrower than one typical grid length and much steeper than most such models can use without severe filtering. These small-scale winds are also an important component of the nocturnal lay-down, as the fuel moisture increase alone (in reality, a response to decreasing atmospheric temperature and increasing specific humidity) cannot completely represent this effect.

As pointed out in past work (Coen *et al.* 2001; Clark *et al.* 2004), the feedback from the fire itself into the atmosphere creates the ‘universal fire shape’, including head, flank and backing regions of the fire. Figure 5 shows this effect on this fire. The fire spreads slowly unless local winds are directing it into unignited fuel. As the fire expands and ignites fuel at the head, the buoyancy from the fire forms an updraft ahead of the fire that draws air in at its base, drawing winds toward

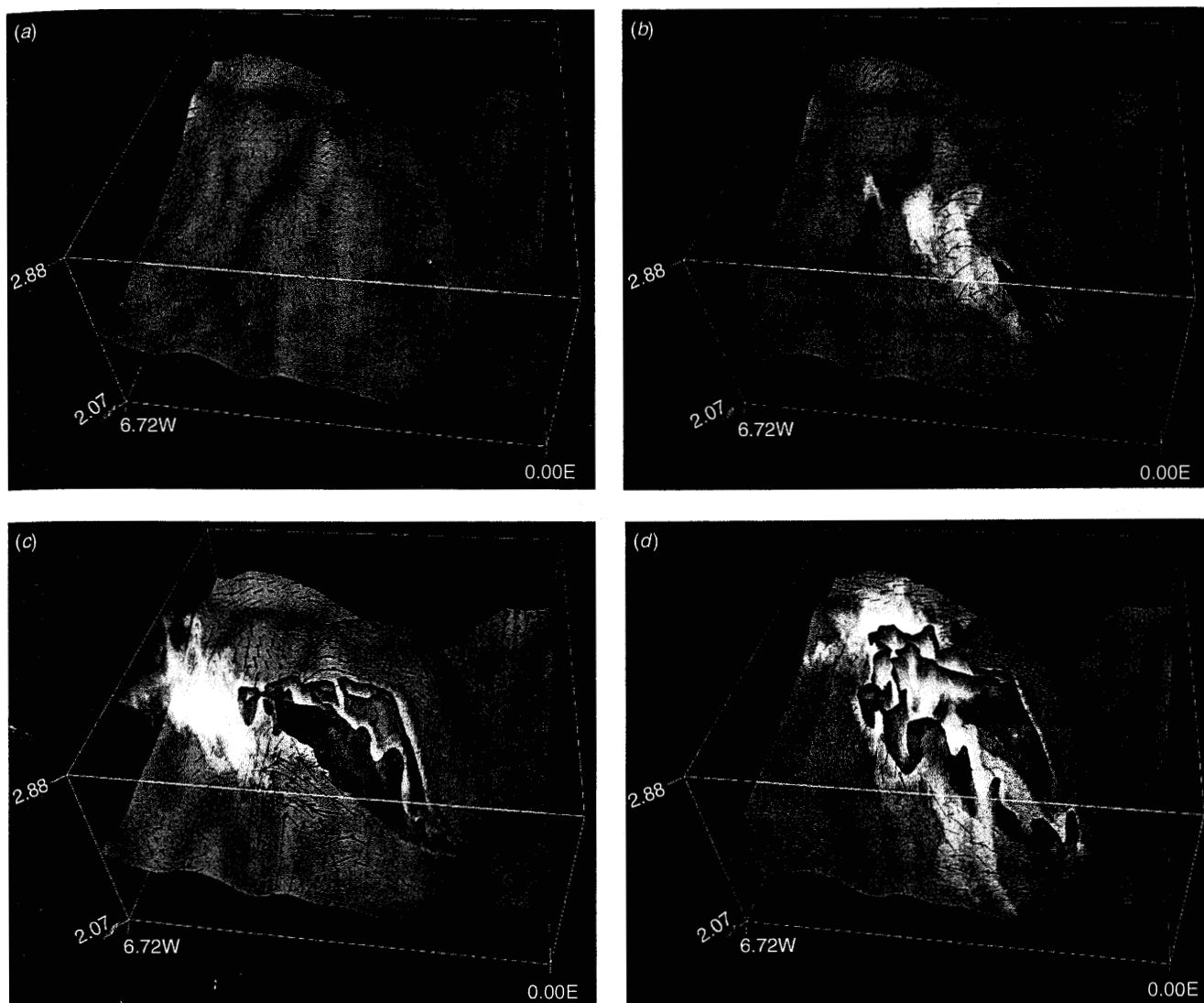


Fig. 3. Four time periods in the first 4 h of the Big Elk Fire. The fire was ignited in the valley, then rapidly climbed the steep slope driven by small-scale up-valley winds. The domain is 6.78 km on the horizontal side and 2.88 km tall. The red isosurface shows where air has a buoyancy of 10°C , the misty white field shows smoke, and the vectors are the winds near the surface.

the head, and leading the head of the fire into unignited fuel. This self-perpetuating fire-affected wind pattern creates the rapidly spreading head, the flanks where the wind blows along the fireline (neither driving nor opposing its growth into fresh fuel), and the creeping backing region, where the wind blows against the fireline, opposing its spread into unburned fuel. Figure 6 shows the difference between the Control and No-feedback simulation in the vertical velocity field and the wind vectors at 18 m above surface, i.e. the 'feedback effect'. The primary effect is a wide updraft region of $2\text{--}8\text{ m s}^{-1}$ at and ahead of the fire, accompanied by confluence of low-level horizontal winds into the base of the updrafts, fed by down-drafts of $2\text{--}3\text{ m s}^{-1}$. Once one accepts that the fire releases heat into the atmosphere, these flow effects are required by the principles of fluid dynamics. (This further supports the

claim made earlier that there is not a representative 'background wind' near a fire that is not affected by it.) However, Fig. 6 also shows that the fire affects the winds farther from the fire and, in fact, throughout the 6.7 km^2 domain, perturbing horizontal and vertical wind components by $2\text{--}4\text{ m s}^{-1}$. Figure 7 shows that including this feedback of the fire on the environmental winds (Fig. 7a) produces a much different-shaped fire than neglecting this effect (Fig. 7b), because the effect of the feedback is to concentrate fire growth at the head and create winds parallel to the interface between the fire and fresh fuel at the flanks, thereby limiting the growth there.

Simulations over a range of atmospheric grid resolutions (Table 2: experiments 3DOM–5DOM and Control) suggest that, while such studies provide interesting insight into fire behavior, they are not essential for capturing many important

Legend

-  BigElk_perim_0717
-  BigElk_perim_0718_1330h
-  BigElk_perim_0718_1800h
-  BigElk_perim_0719
-  BigElk_perim_0720_gps_1200h
-  BigElk_perim_0721_p
-  BigElk_perim_0722b_p

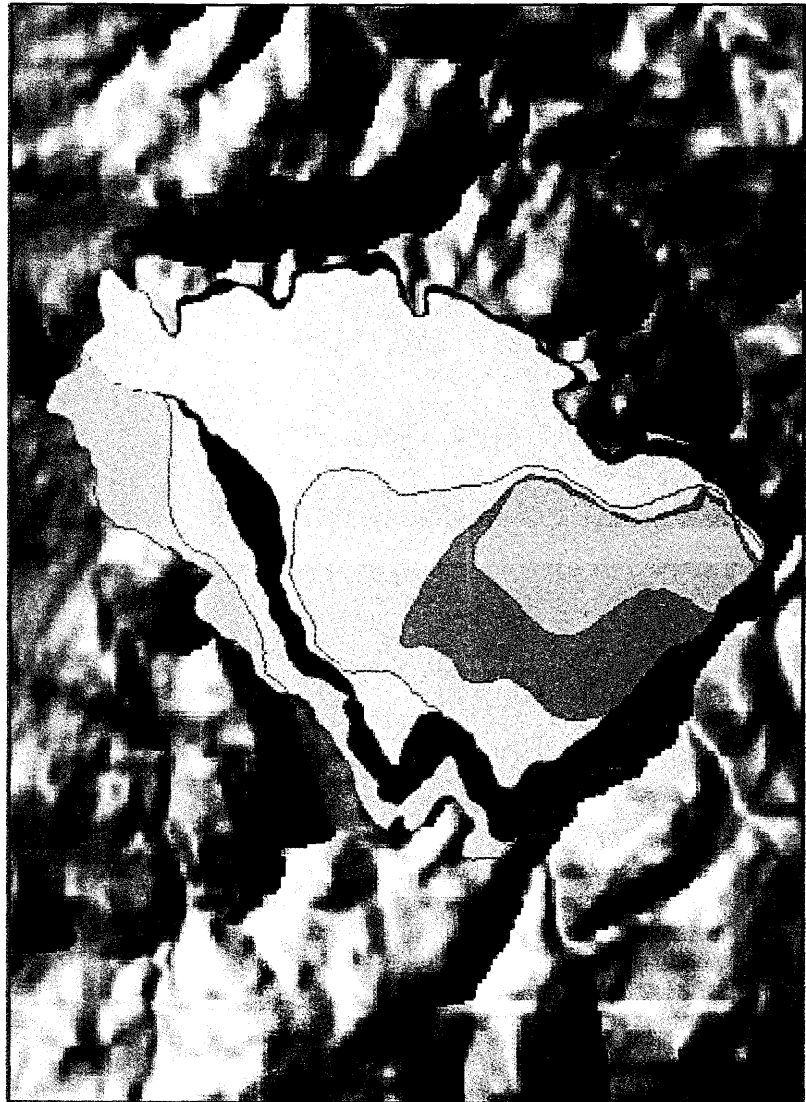
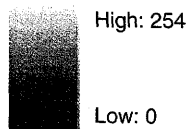
Value

Fig. 4. Fire perimeter progression of the Big Elk Fire at times: 17 July 2200 LST, 18 July 1330 LST, 18 July 1800 LST, 19 July 2200 LST, 20 July 1200 LST, 21 July 2200 LST, and 22 July 2200 LST. Perimeters through 19 July are estimated based on information from line personnel. The first perimeter derived from GPS information was 20 July (Close 2002).

aspects of fire behavior, such as fire growth (spread rate, direction, crown fire ignition, fire shape, etc.). Simulations testing the resolution of the atmospheric grid spacing show that, surprisingly, the four-domain simulation captured the overall spread of the fire nearly as well as the six-domain simulation. The primary difference is that the coarse simulations overestimated the spread rate of the fire (Fig. 8) relative to the simulations that nest to finer resolution. This is because, with coarser atmospheric grid spacing, wind velocities both in the horizontal and vertical directions must be interpolated from grid points farther away. Comparing horizontal and vertical grid refinement, the vertical grid refinement near the surface appears more important, because, as shown in Fig. 8, 5 DOM and Control (six domains), which have the same vertical grid, produce essentially the same fire growth rate. In addition, the

perturbation due to the fire is more diluted when applied over a larger grid volume. Thus, the fire does not focus the winds near the fire so sharply at the head.

Conclusions

These simulations are designed to guide further model development toward useful tools that may give advance knowledge of wildland fire behavior and spread. They show that the coupling between the fire and atmosphere is shown to be an important force that affects the winds at least 5 km from the fire and shapes the overall fire perimeter growth and therefore must be included. Although this effect will always be present, the magnitude and range of this effect will vary from case to case. However, fairly coarse simulations (by detailed fire model standards, although still very fine by numerical

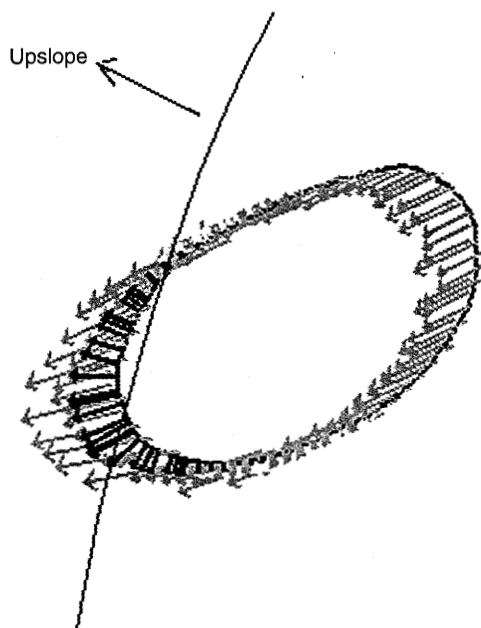


Fig. 5. A snapshot of fire spread during the simulation shown in Fig. 3. The thin black line shows the fire perimeter, the light gray vectors show the wind (affected by the fire) near the fire, and the black vectors show the calculated fire spread.

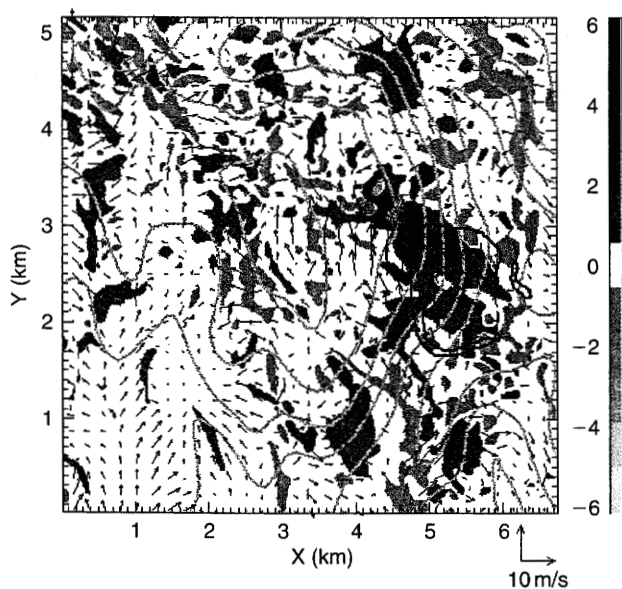


Fig. 6. The difference in the vertical velocity (filled contours, m s^{-1}) and near-surface (18 m AGL) horizontal wind vectors between Control and No-feedback experiments at 1658 LST. Topography contours at 100 m intervals are light gray. Black contours enclose the areas where the fire produces a heat flux greater than 10 kW m^{-2} .

weather prediction standards) on the order of 100–500 m horizontal grid spacing can capture valuable fire behavior information such as spread rates, spread direction and some elements of extreme fire behavior (such as the occurrence

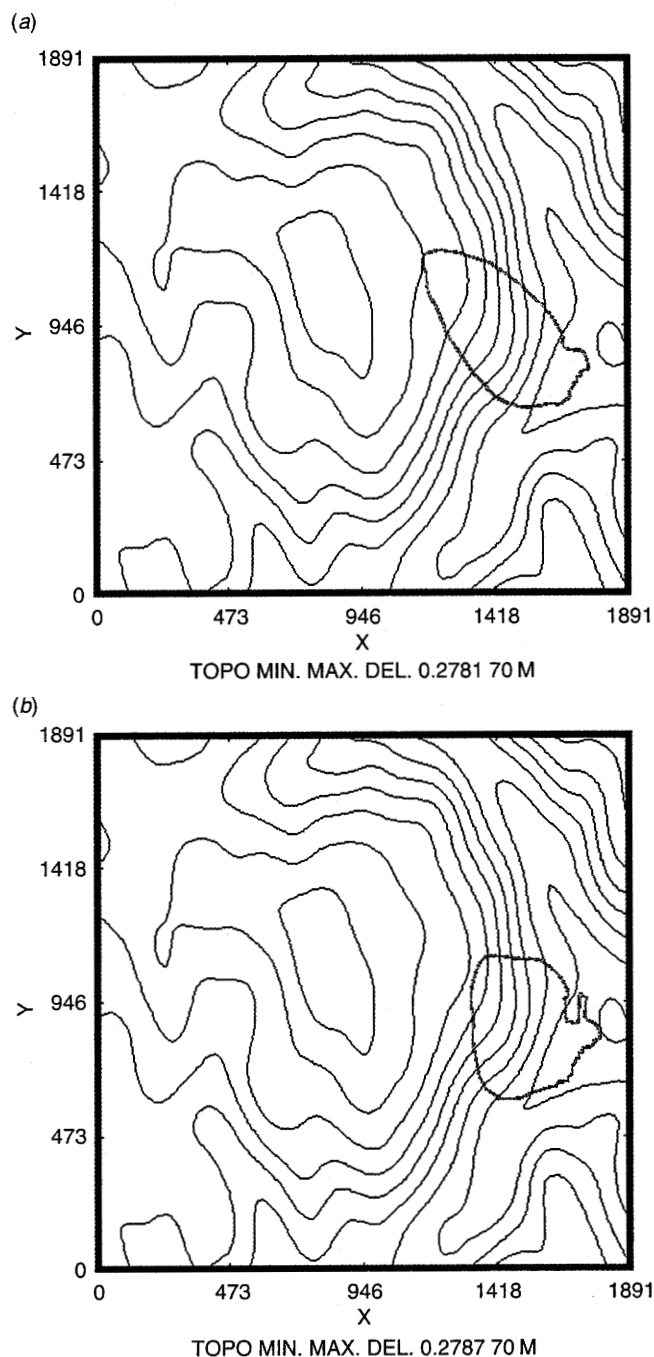


Fig. 7. The simulated fire perimeter (thick contour) in the sixth (inner-most) domain from the (a) Control and (b) No-feedback experiments. Thin contours represent the topography every 70 m. The x- and y-axes indicate the fuel cell number.

of crown fires, although transitory phenomena such as small-scale fire whirls cannot be resolved at these scales).

Past simulations (often at 10–20 m horizontal grid spacing over 2–4 km wide domains) (Clark *et al.* 1996a, 1996b) have shown a great deal of intense, very small-scale bursts, vortices and runs associated with the fireline. To apply

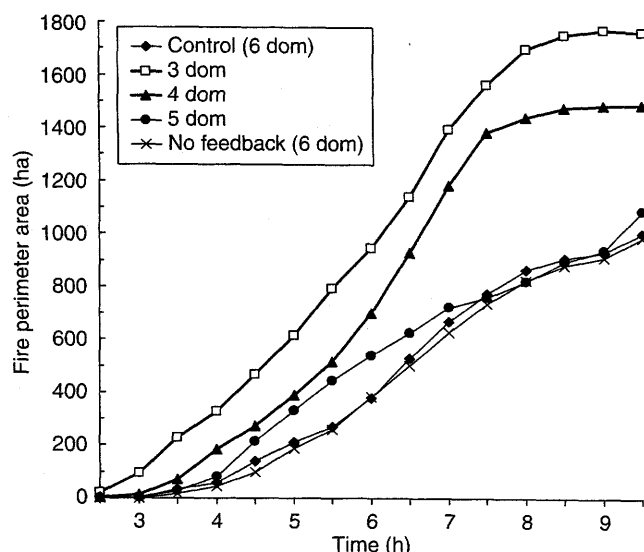


Fig. 8. Fire perimeter growth for five simulations: Control (six domains), three, four and five domains, and the Control experiment with No feedback from the fire to the atmosphere.

resolutions such as those on the scale of real fires would make the prospect of real-time simulations daunting on single-processor machines. The result that configurations with 100–500 m horizontal grid spacing capture the overall spread of the fire is significant because this model configuration covering a domain large enough to span an actual wildfire runs over six times faster than real time on a single Alpha 866 Hz processor. Thus, faster-than-real-time predictive simulations of wildfires are currently possible with this coupled atmosphere–fire model. The six-domain simulation runs eight times slower than real time; however, neither of these configurations was optimized for speed. Implementing the model in a distributed-memory parallel-computing environment, which is still in progress, could reduce the computing time by factors corresponding approximately to the number of processors used and the relative processor speed.

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Coherent vortical structures in numerical simulations of buoyant plumes from wildland fires

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Abstract. The structure and dynamics of buoyant plumes arising from surface-based heat sources in a vertically sheared ambient atmospheric flow are examined via simulations of a three-dimensional, compressible numerical model. Simple circular heat sources and asymmetric elliptical ring heat sources that are representative of wildland fires of moderate intensity are considered. Several different coherent vortical structures that dominate the plume structure and evolution are evident in the simulations, and these structures correspond well with those observed in plumes from wildland fires. For the circular source, these structures include: (i) a counter-rotating vortex pair aligned with the plume trajectory that is associated with a bifurcation of the plume, (ii) transverse shear-layer vortices on the upstream face of the plume, and (iii) vertically oriented wake vortices that form periodically with alternating sign on either side of the downstream edge of the plume base. For the elliptical ring source, a streamwise counter-rotating vortex pair is apparent on each flank, and a transverse horizontal vortex is observed above the head of the source. In all simulations the plume cross section is represented poorly by a self-similar Gaussian distribution.

Introduction

Plumes from wildfires and prescribed fires represent a critical aspect of smoke management, and as such an accurate prediction of plume rise and dispersion is critical for air quality assessment purposes. To develop such accurate predictions, it is important to unravel the basic mechanisms underlying the dynamics of buoyant plumes due to intense surface heat sources over a wide range of atmospheric conditions (i.e. wind and temperature profiles).

Many current models employed for the prediction of plume rise are so-called 'integral models', which assume self-similar cross-sectional profiles of physical quantities, typically of Gaussian structure, along the centre line of the plume, together with simple laws describing the entrainment of ambient air into the plume. Development of these models largely follows the analysis of Morton *et al.* (1956), extended to include the effects of neutral and stably stratified cross flows (e.g. Slawson and Csanady 1967; Briggs 1975; Weil 1988). Integral models may provide accurate predictions of

the plumes arising from low-intensity fires; however, they are frequently inadequate for describing the behaviour of plumes originating from fires of higher intensity, particularly those with complex spatial configurations. In such fires, the dynamical interaction between the buoyancy-generated vorticity in the plume and the vorticity in the ambient atmosphere may be significant, and the behaviour of the plume may differ substantially from the predictions of the integral models, possibly dramatically so in the region near the fire.

These plumes are typically dominated by coherent vortex structures of varying size and nature that control the evolution of the flow. In this regard, buoyant plumes in a cross flow appear to share many common properties with non-buoyant jets in a cross flow, which have been studied much more extensively (for a review, see Margason 1993). Several types of vortices have been identified in such flows, and these are illustrated schematically in Fig. 1. The dominant feature in these flows is typically the counter-rotating vortex pair, which is observed in the far field aligned with the jet or plume

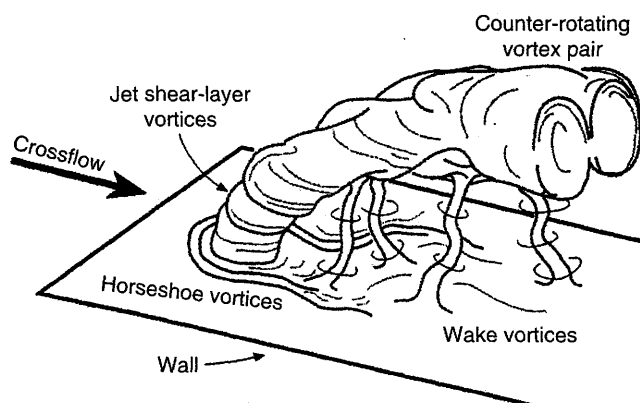


Fig. 1. Schematic illustration depicting four types of vortical structures associated with a non-buoyant jet in a cross flow: jet shear-layer vortices at the perimeter of the bending jet; the counter-rotating vortex pair; horseshoe vortices on the wall; and wake vortices extending from the wall to the jet. From Fric and Roshko (1994).

trajectory; this vortex pair is manifested as an approximately kidney-shaped cross section for scalar concentration, and in extreme cases can result in the split, or bifurcation, of the jet or plume some distance downstream of the source. Observations of plume bifurcation in association with intense ground fires are relatively common (Church *et al.* 1980; Haines and Smith 1987; Banta *et al.* 1992; McGrattan *et al.* 1996), although a visible bifurcation is often short lived, possibly due to mixing by ambient atmospheric turbulence.

While the counter-rotating vortex pair is apparently a ubiquitous feature of jets in a cross flow, and might thus be expected to be so for plumes also, there is still considerable debate about its generation and evolution (e.g. Coelho and Hunt 1989; Morton and Ibbetson 1996; Yuan and Street 1998; Yuan *et al.* 1999; Cortelezzi and Karagozian 2001; Lim *et al.* 2001). Nevertheless, there is a general consensus in the contemporary literature that in non-buoyant jets this feature arises as a consequence of the interaction between the cross-flow boundary layer and the jet shear layer. For plumes in a cross flow, the presence of buoyancy may further complicate the description. The existence of a kidney-shaped, or possibly bifurcated, structure for the plume cross section might also call into question the common practice of using Gaussian profiles for plume cross-section structure (Janseen *et al.* 1990), although it is possible that at large distances from the source, where buoyancy generated turbulence in the plume is comparable to ambient turbulence in the atmosphere, a Gaussian distribution may be appropriate. It is also possible that the time-averaged plume cross section structure might be close to Gaussian.

The wake vortices depicted in Fig. 1 and studied experimentally by Fric and Roshko (1994) for jets in a cross flow also have a counterpart in buoyant windblown plumes. Specifically, vertically oriented vortices bearing a strong

resemblance to dust devils, or even tornadoes, have frequently been observed downwind of intense fires (Church *et al.* 1980; McRae and Flannigan 1990; Soma and Saito 1991; Wood 1992), extending from the ground to the underside of the bent-over plume. Occasionally, such vortices pass over and entrain burning material, resulting in a so-called fire whirl with an internal flame that can reach heights of tens of meters.

Although jets in a cross flow have been the subject of many numerical investigations (see Yuan *et al.* (1999), Cortelezzi and Karagozian (2001), and references therein), with notable exceptions there have been relatively few numerical simulation studies of buoyant plumes in a cross flow. To date, the majority of numerical investigations in this regard have been either two-dimensional (Luti 1980, 1981; Heilman and Fast 1992; Porterie *et al.* 1999) or quasi-three-dimensional, in which the streamwise velocity is taken to be constant, and the governing equations are solved in planes perpendicular to the streamwise direction (Zhang and Ghoniem 1993; McGrattan *et al.* 1996; Trelles *et al.* 1999). Although these studies have yielded significant insight into the structure and dynamics of buoyant plumes, and the quasi-three-dimensional models are able to represent the counter-rotating vortex pair, none of these models can represent fully the evolving three-dimensional vorticity. More recently, Baum and McGrattan (2000) have performed three-dimensional large-eddy simulations of plumes from outdoor industrial fires; however, the three-dimensional vortical structures evident in their simulated plumes were not the primary focus of their study.

The present study represents our preliminary attempts to simulate buoyant plumes arising from wildland fires in the presence of a cross flow using a fully three-dimensional numerical model. Of particular interest in this regard is whether the model is capable of capturing the vortical structures described above, and the dependence of these structures on physical and numerical parameters. Also of interest is the extent to which the simulated plumes demonstrate similarities or differences with the predictions of the simple integral models. The numerical model is based on the three-dimensional, compressible Navier–Stokes equations including heat source and surface drag terms, and is solved using direct and large-eddy simulation techniques. These simulations are employed to examine the effect of vertical wind shear near the surface, and of heat-source shape and intensity, on buoyant plumes arising from heat sources on the ground.

In the next section, the numerical model to be employed in this study is described, along with the form of the heat sources that will be used. In the subsequent two sections, results are presented from simulations using respectively a simple circular heat source and a heat source with spatial structure more representative of a wildland fire. In the final section, a summary of the primary results of the present study is presented, along with a discussion of the relevance of the results to the behaviour of buoyant plumes in general.

Model formulation and solution technique

We consider the ambient atmospheric wind profile to be an initially laminar flow, $U(z)$, in the positive x -direction as follows:

$$U(z) = U_0 \tanh(z/z_0) \quad (1)$$

where $U_0 = 4.5 \text{ m s}^{-1}$, and z_0 is varied in an effort to understand the impact on plume dynamics of vertical wind shear near the surface. The specification of a laminar cross flow is consistent with the goal of this investigation of developing a basic understanding of the dynamics of the plume, but it also has a physical basis in the assumption that the turbulence generated by the plume is expected to be significantly larger in magnitude than the ambient atmospheric turbulence. For moderate to intense heat sources, this assumption would appear to be reasonable within 1 or 2 km from the fire. Nevertheless, the effects of ambient atmospheric turbulence on plume rise and dispersion are important for practical prediction, and will be addressed in future work.

Governing equations

The governing equations are those for three-dimensional, compressible flow in a density-stratified atmosphere forced by a prescribed volumetric heat source, Q :

$$\frac{\partial}{\partial t}(\rho u_i) + \frac{\partial}{\partial x_j}(\rho u_i u_j) + \frac{\partial p'}{\partial x_i} - g \rho' \delta_{i3} = \frac{\partial \tau_{ij}}{\partial x_j} - D_i, \quad (2)$$

$$\frac{\partial}{\partial t}(\rho \theta) + \frac{\partial}{\partial x_j}(\rho \theta u_j) = \frac{\partial}{\partial x_j} \left(\frac{k}{c_p} \frac{\partial \theta}{\partial x_j} \right) + \frac{Q \theta}{c_p T}, \quad (3)$$

$$\frac{\partial \rho}{\partial t} + \frac{\partial}{\partial x_j}(\rho u_j) = 0, \quad (4)$$

where ρ is the density, u_i is the wind velocity in the x_i direction with $i, j, k = 1, 2$, or 3 representing streamwise (x), spanwise (y), or vertical (z) directions, respectively, p' and ρ' are the departures of pressure and density, respectively, from a hydrostatic base state, D_i is a drag term, θ is the potential temperature, T is the temperature, c_p is the specific heat at constant pressure, and k is the thermal conductivity. The stress tensor, τ_{ij} , is given by

$$\tau_{ij} = \mu \left(\frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right) - \frac{2}{3} \mu \frac{\partial u_k}{\partial x_k} \delta_{ij}, \quad (5)$$

where μ is the viscosity. Finally, the drag term, D_i , is intended to represent the effects of a vegetation canopy, and is defined as

$$D_i = \rho C_d a V u_i, \quad (6)$$

where C_d is the canopy drag coefficient, a is the canopy density, and V is the magnitude of the wind speed parallel to the ground. More generally, this term can be used to represent a rough lower boundary in simulations of the atmospheric boundary layer; Brown *et al.* (2001) show that

the use of a canopy drag term identical to equation (6) in large-eddy simulations of a neutral boundary layer led to a significantly improved representation of logarithmic mean wind profiles in the near-surface region over those predicted by a traditional wall stress modelling approach. In this paper, we choose $C_d = 0.1$ and $a = 0.25 \text{ m}^{-1}$ on the lowest grid level and zero elsewhere. It should be noted, however, that no extensive testing has been performed regarding the choice of these parameters, and further work is required to assess the optimal values and distributions in this regard.

In this paper, we employ two numerical approaches to solving the governing equations: the first is based on a direct approach to simulating the equations, whereas the second is based on large-eddy simulation (LES) methods. In the direct approach, it is assumed that the important energy-containing features are resolved on the grid and that no subgrid-scale model is required. Although such simulations are restricted to Reynolds numbers that are much smaller than those typically found in the atmosphere, there is evidence to suggest that insight into many important aspects of high Reynolds number turbulence can indeed be obtained through direct simulations of low Reynolds number flows (e.g. Moeng and Rotunno 1990; Coleman 1999). Moreover, the absence of the uncertainty that may be introduced by the use of a subgrid-scale model lends clarity to the results of these simulations. Consequently, the primary goal of the direct approach to simulating the governing equations is to attempt to gain insight into the basic physical processes relevant to buoyant plumes in a cross flow, and how this behaviour depends qualitatively on the Reynolds number. In this approach, the viscosity and thermal conductivity assume constant 'molecular' values that are related via the Prandtl number,

$$\text{Pr} = \frac{\mu c_p}{k}, \quad (7)$$

which is taken to be equal to 0.7.

For the LES, τ_{ij} is interpreted as a subgrid-scale stress tensor, and the viscosity and thermal conductivity are interpreted as subgrid-scale eddy coefficients that must be modeled. Since no well accepted approach currently exists for LES when large subgrid-scale gradients of buoyancy and temperature may be present, we adopt a conventional method that has been shown to be successful for buoyancy-dominated boundary layers (e.g. Deardorff 1973; Moeng and Wyngaard 1988) and acknowledge at this point that this approach may not be entirely appropriate. In this scheme, the eddy viscosity depends on the subgrid-scale turbulent kinetic energy, e , as follows

$$\mu = \rho C_k e^{1/2} \Delta, \quad (8)$$

where C_k is a constant and Δ is the grid spacing. The eddy conductivity is then related to the eddy viscosity via a turbulent Prandtl number of the form given by equation (7), which is assumed to be equal to 1/3 herein. To close the model,

an additional equation for the prediction of subgrid-scale turbulent kinetic energy is required, and is given by

$$\frac{\partial}{\partial t}(\rho e) + \frac{\partial}{\partial x_j}(\rho e u_j) = -\tau_{ij} \frac{\partial u_i}{\partial x_j} + 2 \frac{\partial}{\partial x_j} \left(\mu \frac{\partial e}{\partial x_j} \right) - \frac{gk}{c_p \theta} \frac{\partial \theta}{\partial z} - \frac{\rho C_e e^{3/2}}{\Delta}, \quad (9)$$

where the last term on the right-hand side represents the dissipation of subgrid-scale turbulent kinetic energy. Following Moeng and Wyngaard (1988), values for the constants in equation (8) and equation (9) are chosen to be $C_k = 0.1$ and $C_e = 0.93$. It is important to note, however, that these numbers are empirically defined based on Kolmogorov inertial-subrange theory, and a variety of different values both for these constants and for the turbulent Prandtl number have been used by different investigators (e.g. Schumann 1991).

Finally, we note that the solution method, which is described below, exhibits numerical diffusion. Such diffusion is itself sometimes interpreted as an implicit turbulence model (e.g. Hussaini 1998; Margolin *et al.* 2002), although a clear theoretical foundation and detailed validation for such an approach is still lacking. Nevertheless, the diffusion provided by the numerical scheme is weak in comparison to the stress term for most of the simulations to be shown in this paper.

Form of the heat sources

In this initial attempt to simulate buoyant plumes from wildland fires, we assume for simplicity a form for the heat source in equation (3) that is of fixed spatial form, and constant in time after a brief (10 s) ramp-up period. A volumetric heat source is employed in an effort to represent heating due to a fire, with the total heat release rate chosen to be comparable to that expected in a wildland fire of moderate intensity. In future work, we will incorporate heat sources that evolve in a coupled manner with the atmosphere.

To explore the basic dynamics of plumes in a cross flow, a simple circular heat source is examined in the following section that has the form of a smoothed top-hat distribution:

$$Q(x, y, z, t) = \begin{cases} Q_0 \tanh(t/t_0) \exp(-z/h), & r < R_1, \\ Q_0 \tanh(t/t_0) \exp(-z/h) \times \left[\frac{1}{2} - \frac{1}{2} \tanh\left(\frac{r-R_c}{d}\right) \right], & R_1 \leq r < R_2, \\ 0, & r \geq R_2, \end{cases} \quad (10)$$

where Q_0 is a constant that controls the intensity of the heat source, t_0 is the ramp-up time scale, h is the scale height, $r = [(x - x_h)^2 + (y - y_h)^2]^{1/2}$ is the horizontal distance from the centre of the heat source at $(x_h, y_h, 0)$, R_1 is the radius of the uniform heating area, R_2 is the radius of the total heating

area, $R_c = (R_1 + R_2)/2$ and d is a scale width. For the simulations shown in this paper, we take $h = 25$ m, $x_h = 450$ m, $y_h = 600$ m, $R_1 = 75$ m, $R_2 = 150$ m, and $d = 12.5$ m.

Wildland fires rarely exhibit such simple configurations, however, and the maximum heat release is seldom located in the centre of the fire area. Typically, fires possess an asymmetric ring-like structure that is elliptical (with major axis in the streamwise direction), most intense at its forwards end (the head), less intense on the sides (the flanks), and weakest at its rear end (the back). Consequently, in this paper we also consider heat sources with shapes that are representative of this typical structure; specifically, the following functional form for the heat source is adopted:

$$Q(x, y, z, t) = Q_0 \tanh(t/t_0) \exp(-z/h) \times \left\{ \frac{(\cos[\operatorname{atan}(Y/X)] + 1)}{2[2 - (X/A)^2 - (Y/B)^2]^4 + 2} \right\} \quad (11)$$

where $X = x - x_h$, $Y = y - y_h$, and we choose $A = 200$ m and $B = 100$ m such that the heat source assumes the form of an ellipse with major axis in the streamwise direction. All other parameters are the same as in equation (10). This heat source is referred to as the 'realistic' heat source. Horizontal distributions of the heat sources given by equation (10) and equation (11) are depicted in Fig. 2.

The total heat release rate associated with these sources is obtained by integrating the expressions for Q given in equation (10) and equation (11) over the volume in which Q is non-zero. For the smoothed top-hat source in equation (10), choosing $Q_0 = 1$ kW m⁻³ gives a total heat release rate of 1000 MW, and for the realistic source in equation (11), choosing $Q_0 = 1$ kW m⁻³ gives a total heat release rate of 1700 MW. Such values for the total heat release rate and the heat release per unit volume are considered representative of those found in typical wildland fires of moderate intensity.

Solution technique

To solve the governing equations described in the preceding sections, we use the Weather Research and Forecasting (WRF) model code (Skamarock *et al.* 2001). For time integration, a split-explicit method is employed that combines a third-order Runge-Kutta scheme for advection with a forwards-backward scheme for the high-frequency acoustic modes (Wicker and Skamarock 2002). Spatial discretization of the advection terms is treated via a fifth-order upwind-biased scheme.

The computational domain is depicted schematically in Fig. 3. The domain is a rectangular box with spatial dimensions $L_x \times L_y \times L_z$, where $L_x = 1800$ m, $L_y = 1200$ m, and $L_z = 1500$ m. A uniform grid spacing of 10 m is employed in all directions. At the lateral and downstream boundaries a non-reflecting outflow condition is prescribed (Orlanski

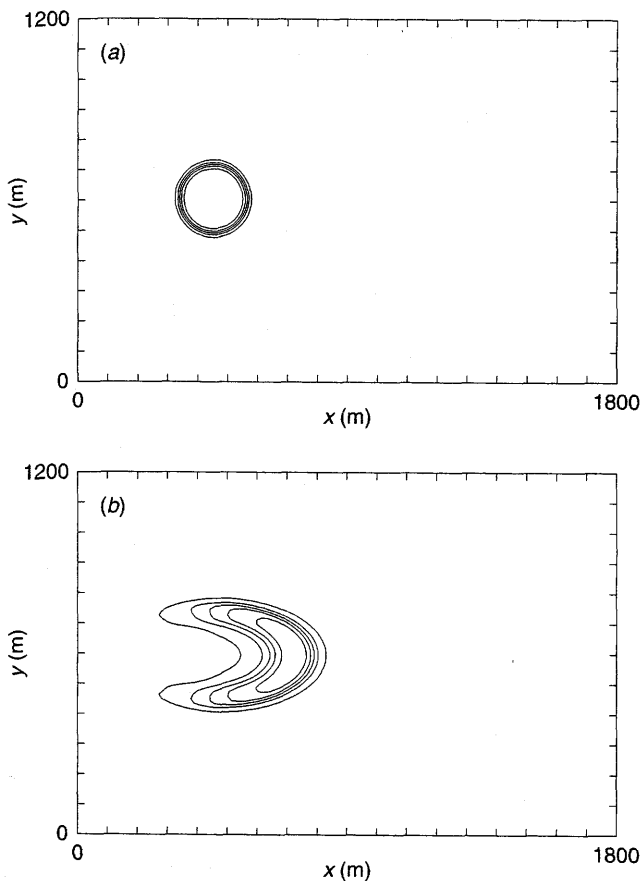


Fig. 2. Non-dimensional function for the spatial distribution of heating at the lowest model level for (a) the circular heat source and (b) the realistic heat source. Contours are for values between 0.1 and 0.9 with an interval of 0.2.

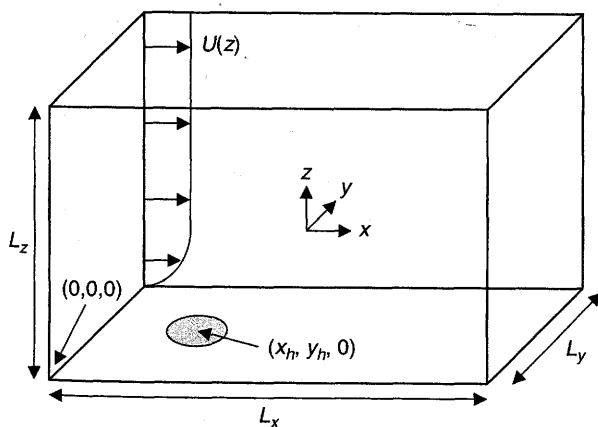


Fig. 3. Schematic illustration of the computational domain. The heat source, indicated by the shaded area, is centred at $(x_h, y_h, 0)$.

1976). At the upper and lower boundaries, solid-wall free-slip conditions are imposed; although such conditions are not optimal for models of the dimensions and spatial resolution used herein, it is noted that the upper boundary is

high enough such that the outflow occurs entirely on the lateral and downstream boundaries of the domain. We have performed sensitivity experiments to ensure that the upper boundary has been placed sufficiently high that this choice has negligible effect on plume behavior. In addition, a layer of enhanced damping is placed adjacent to the upper boundary to avoid the reflection of energy back into the domain. At the lower boundary, the effects of the rough surface are treated via the drag term (equation 6). In all simulations, the ambient potential temperature is taken to be uniform and equal to 300 K.

Buoyant plumes from an idealized circular heat source

In this section, we describe simulations of the buoyant plumes arising from the heat source (equation 10) in the presence of the wind profile (equation 1). In an effort to gain basic insight into the laminar and turbulent structure of such plumes, several simulations using the direct approach have been performed with a range of different values of the viscosity, μ , keeping the Prandtl number fixed at 0.7. This approach is essentially equivalent to changing the Reynolds number of the simulation; however, given that there are several equally plausible ways to choose velocity and length scales for the plume in a cross flow, and since the model is not dimensionless, we do not define a specific Reynolds number for the present situations.

Figure 4 and Figure 5 show respectively the potential temperature and vorticity magnitude, $|\nabla \times \mathbf{V}|$, at $t = 1100$ s for four particular simulations with successively decreasing viscosity (increasing Reynolds number) as follows: $\mu = 4 \text{ kg m}^{-1} \text{ s}^{-1}$ (Figs 4a, 5a), $\mu = 1 \text{ kg m}^{-1} \text{ s}^{-1}$ (Figs 4b, 5b), $\mu = 0.15 \text{ kg m}^{-1} \text{ s}^{-1}$ (Figs 4c, 5c), and $\mu = 0.0015 \text{ kg m}^{-1} \text{ s}^{-1}$ (Figs 4d, 5d). For the large μ case, the plume is essentially laminar, and exhibits a well defined symmetric bifurcation (Fig. 4a). The corresponding vorticity field (Fig. 5a) clearly shows the associated counter-rotating vortex pair; however, no other vortical structures are apparent in this simulation. Near the surface the vertical component of vorticity in the pair is significant, with positive vorticity on the right-hand side looking downstream, and negative vorticity on the left-hand side, consistent with descriptions of observed plumes (e.g. Church *et al.* 1980). As μ is decreased (equivalent to increasing the nominal Reynolds number) the simulated plumes become turbulent and asymmetric, and additional vortical structures become evident. Nevertheless, the counter-rotating vortex pair remains a central, albeit unsteady, feature in all of the simulations, and the plume cross sections do not exhibit a self-similar Gaussian structure.

The transition to turbulence in these plumes appears to arise via the shear (i.e. Kelvin–Helmholtz) instability of the upstream face of the plume that leads to shear-layer vortices analogous to those seen in Fig. 1. Such a transition is apparent in Fig. 4b and Fig. 5b in which transverse vortices can be seen at the top of the laminar base of the plume; above

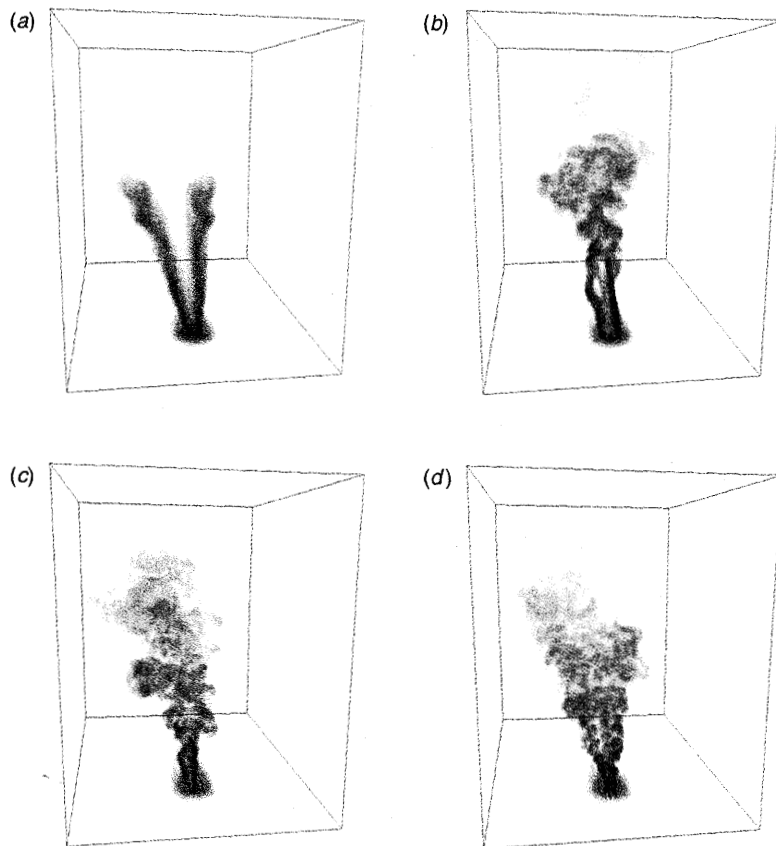


Fig. 4. Volume rendered potential temperature for the circular heat source simulations with: (a) $\mu = 4 \text{ kg m}^{-1} \text{ s}^{-1}$, (b) $\mu = 1 \text{ kg m}^{-1} \text{ s}^{-1}$, (c) $\mu = 0.15 \text{ kg m}^{-1} \text{ s}^{-1}$, and (d) $\mu = 0.0015 \text{ kg m}^{-1} \text{ s}^{-1}$. In all cases the view is looking downstream. Darker shades represent higher values of potential temperature.

these vortices the plume is turbulent. At the early stages of the lower Reynolds number simulations (not shown), the onset of this instability can be seen as a regularly spaced array of these transverse vortices, consistent with the low Reynolds number laboratory experiments of Perry and Lim (1978). Further decreases (increases) in the viscosity (Reynolds number) result first in the transition to turbulence occurring closer to the base of the plume (e.g. Figs 4c,d, 5c,d), and subsequently in the breakdown of the symmetric wake structure into an asymmetric one involving vertically oriented wake vortices. The view perspective in these figures does not provide an ideal illustration of the wake vortices, however, and these features are examined in further detail later in this section.

Before exploring the nature of vortical structures in the turbulent plumes, however, it is worthwhile to examine in more detail the low Reynolds number case in an attempt to gain additional insight into the dynamics of the counter-rotating vortex pair, which is a robust feature of all the simulations. Specifically, we explore the dependence of this feature on some of the parameters that define the present problem. To this end, six simulations were performed with $\mu = 4 \text{ kg m}^{-1} \text{ s}^{-1}$ for two different intensities for the heat source (i.e. $Q_0 = 1 \text{ kW m}^{-3}$ and 0.5 kW m^{-3}) and three different depths for the cross-flow shear layer (i.e. $z_0 = 50 \text{ m}$, 100 m , and 150 m). Cross sections of potential temperature in the y - z plane at $x = 1750 \text{ m}$ for these simulations are shown

in Fig. 6 for the time at which the flows are essentially steady (typically achieved after approximately 600 s). The principal result emerging from these simulations is that for larger z_0 , which corresponds to a deeper cross-flow shear layer, the bifurcation of the plume is wider. Additionally, for a given cross flow, bifurcation is wider for the weak heat source than the strong heat source (compare Fig. 6e with Fig. 6f).

Based on this limited set of results we conjecture that, since the magnitude of the heat source is a major factor in the magnitude of the vertical velocity, the width of the bifurcation is controlled by the amount of time it takes for a buoyant parcel to rise through the shear layer. Presumably the relevant process is the entrainment and organized rearrangement of ambient horizontal vorticity in the cross-flow shear layer by the rising buoyant elements, although further investigation of the details of this process is necessary. A similar result will be shown in the simulations of the realistic heat source in the next section. It is important to note, however, that the presence and degree of bifurcation does not appear to impact the plume rise for a given heat source (e.g. compare Fig. 6b,d,f).

As discussed in the Introduction, the origin of the counter-rotating vortex pair in jets or plumes in a cross flow is still an unresolved issue, although it is generally accepted that this feature arises through an interaction between the vorticity

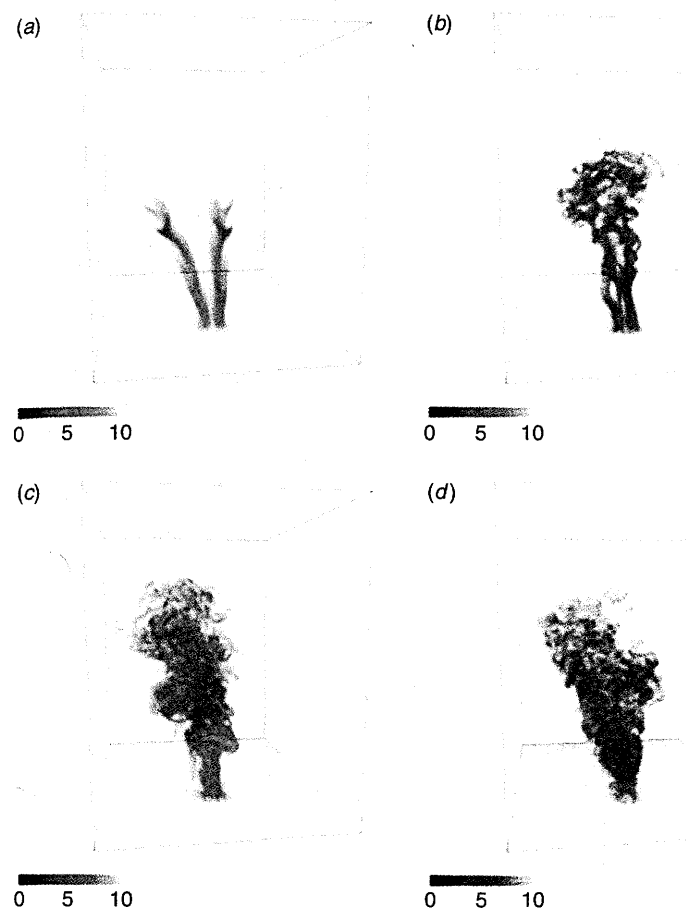


Fig. 5. Volume rendered vorticity magnitude (in s^{-1} , shaded as indicated) for the circular heat source simulations with: (a) $\mu = 4 \text{ kg m}^{-1} \text{ s}^{-1}$, (b) $\mu = 1 \text{ kg m}^{-1} \text{ s}^{-1}$, (c) $\mu = 0.15 \text{ kg m}^{-1} \text{ s}^{-1}$, and (d) $\mu = 0.0015 \text{ kg m}^{-1} \text{ s}^{-1}$. In all cases the view is looking downstream.

in the cross flow and the vorticity in the jet or plume. For the simulations shown here it is apparent that the counter-rotating vortex pair originates near the ground as a pair of vortices that are oriented primarily in the vertical direction (Fig. 5; see also Fig. 9 below). This result also appears to be true for non-buoyant jets in a cross flow (Yuan *et al.* 1999; Lim *et al.* 2001). It is important to note at this point that buoyancy can only directly generate horizontal vorticity, and that vertical vorticity can be generated only by tilting horizontal vorticity into the vertical direction; these are fundamental properties of the governing equations, not of the strength or distribution of the heat source. In this regard, in the absence of a cross flow the vorticity in a buoyant plume can be described in terms of horizontal vortex rings that propagate upwards. When a cross flow is present, the horizontal vorticity in the cross-flow shear layer is tilted upwards within these buoyant vortex rings, leading to a positive-negative couplet of vertical vorticity on the flanks of the plume. On the other hand, since the cross flow is horizontally uniform, there is no mechanism initially to tilt the horizontal vorticity of the plume, although once the cross-flow vorticity has been tilted into the vertical, the interaction becomes complex and three-dimensional.

In light of the foregoing discussion, we assert that the primary source of vorticity at the origin point of the counter-rotating vortex pair is the tilting of horizontal cross-flow vorticity into the vertical by the plume, since it is not obvious that the buoyancy-generated horizontal vorticity in the plume can be tilted into the vertical in such an organized fashion. Farther from the source in the bent-over region of the plume, however, the counter-rotating vortex pair is oriented primarily horizontally, and buoyancy may contribute directly to the vorticity in the counter-rotating vortex pair in the manner described by Zhang and Ghoniem (1993).

The foregoing simulations provide important insight into the basic dynamics of buoyant plumes in a cross flow, particularly with respect to the associated coherent vortical structures. As discussed previously, however, the direct approach to simulating the equations is restricted to low Reynolds number flows, and to simulate the high Reynolds numbers characteristic of the atmosphere, an explicit subgrid-scale model is required to represent small-scale turbulent mixing. Consequently, we continue our exploration into the dynamics of buoyant plumes in a cross flow using the LES version of the model, in which the turbulent mixing is parameterized as described in the previous section. For the remainder

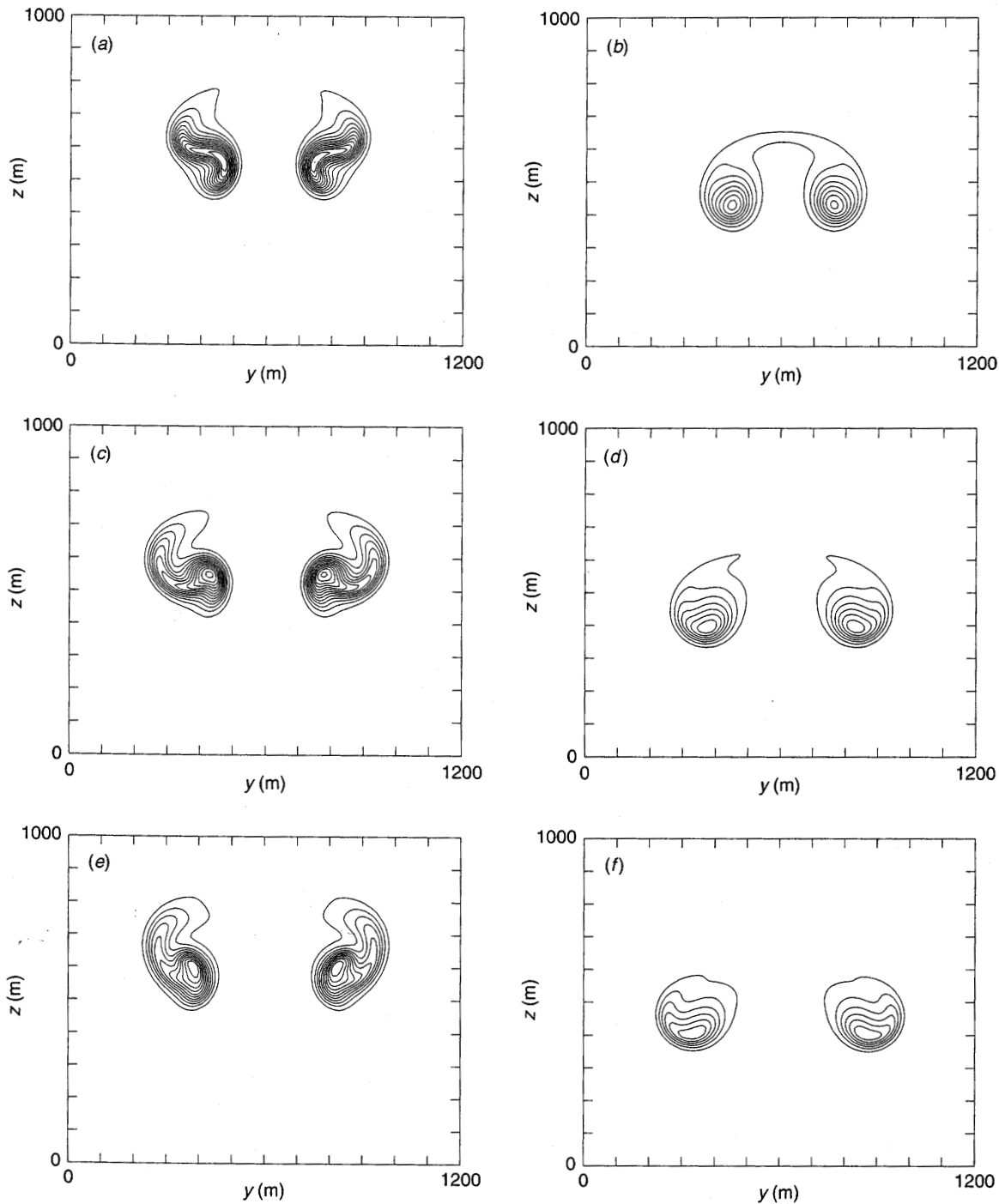


Fig. 6. Cross sections of potential temperature in the y - z plane at $x=1750$ m for the circular heat source simulations ($\mu = 4 \text{ kg m}^{-1} \text{ s}^{-1}$) with $Q_0 = 1 \text{ kW m}^{-3}$ (left column) for (a) $z_0 = 50$ m, (c) $z_0 = 100$ m, (e) $z_0 = 150$ m, and with $Q_0 = 0.5 \text{ kW m}^{-3}$ (right column) for (b) $z_0 = 50$ m, (d) $z_0 = 100$ m, (f) $z_0 = 150$ m. Contour interval is 0.25 K with the first plotted contour equal to 300.25 K.

of this section, we focus on one specific simulation, namely for $Q_0 = 1 \text{ kW m}^{-3}$ and $z_0 = 100$ m.

As with all other simulations, the counter-rotating vortex pair and the associated plume bifurcation are apparent throughout the evolution; however, in the present simulation

these features undergo a quasi-periodic oscillation in which each branch of the pair is successively dominant over the other. This oscillation is illustrated in Fig. 7, in which it can be seen that at $t = 850$ s (Fig. 7a,c) the right-hand branch of the pair dominates, while at $t = 950$ s (Fig. 7b,d) the

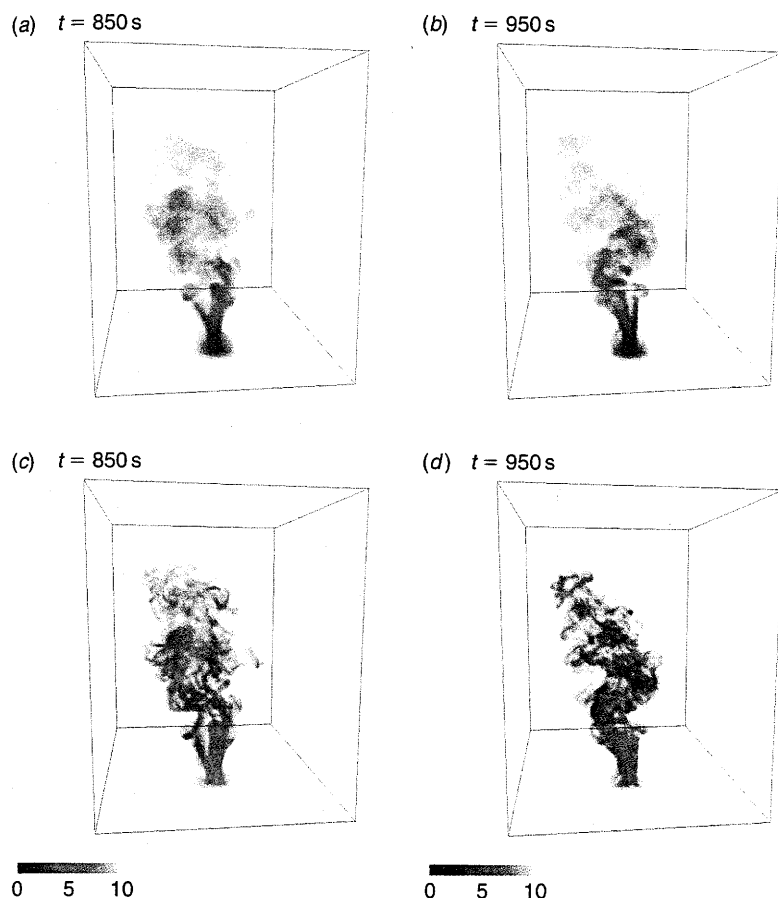


Fig. 7. Volume rendered potential temperature (darker shades represent higher values) at (a) $t = 850$ s and (b) $t = 950$ s, and vorticity magnitude (in s^{-1} , shaded as indicated) at (c) $t = 850$ s and (d) $t = 950$ s for the circular heat source LES.

left-hand branch dominates. To explore this oscillation further, we show in Fig. 8 the vertical component of vorticity within the cross-flow shear layer (at $z = 45$ m) for a range of times encompassing the oscillation of the plume seen in Fig. 7. It is readily apparent from this figure that there is a 'shedding' of localized regions of vertical vorticity with alternating sign from either side of the plume, with a period of about 200 s (note the similarity between Fig. 8a and Fig. 8d). This vertical vorticity corresponds to the wake vortices depicted in Fig. 1. The presence of a narrow turbulent wake downstream of the heat source near the surface is also apparent.

A non-dimensional number that has been shown to be relevant for the wake vortex system, both in non-buoyant jets in a cross flow (e.g. Fric and Roshko 1994) and in flow around a circular cylinder (e.g. Batchelor 1967), is the Strouhal number, defined by

$$\text{St} = \frac{nD}{U_0}, \quad (12)$$

where n is the frequency of the oscillation, $D = 2R_c$ is the diameter of the heat source, and U_0 is the maximum cross flow speed. Based on an estimated period of 200 s, the Strouhal number in this case is approximately $\text{St} = 0.25$. This value is comparable with that seen in experiments of the wake

system in non-buoyant jets ($\text{St} \approx 0.05\text{--}0.2$), and also for the wake vortices associated with flow around a solid cylinder ($\text{St} \approx 0.2$), although it is stressed that the mechanism for the creation of the wake vortices is different for jets and plumes than it is for solid bodies, and the analogy of the plume as an obstacle in the flow is only partially accurate.

A graphic illustration of the wake vortices in this simulation is provided in Fig. 9, which shows the vertical component of vorticity for the same times as in Fig. 7. The successive formation of positive and negative vertical vortices downstream of the plume is readily apparent, as is their continued advection downstream. The 'positive phase' of the oscillation at $t = 850$ s, characterized by the creation of a positive vortex on the right-hand side of the plume (Fig. 9a), corresponds to the stage at which the plume is skewed to the right-hand side (Fig. 7a), while in the 'negative phase' at $t = 950$ s, the creation of a negative vortex on the left-hand side (Fig. 9b) corresponds to the plume being skewed to the left-hand side (Fig. 7b). The vortices are subsequently swept downstream and tilted away from the vertical by a combination of the plume and the cross flow.

It is important to note that the ability of the model to resolve the wake vortex system is a direct consequence of its three-dimensional nature. The quasi-three-dimensional models, in which the streamwise velocity is fixed (Zhang and

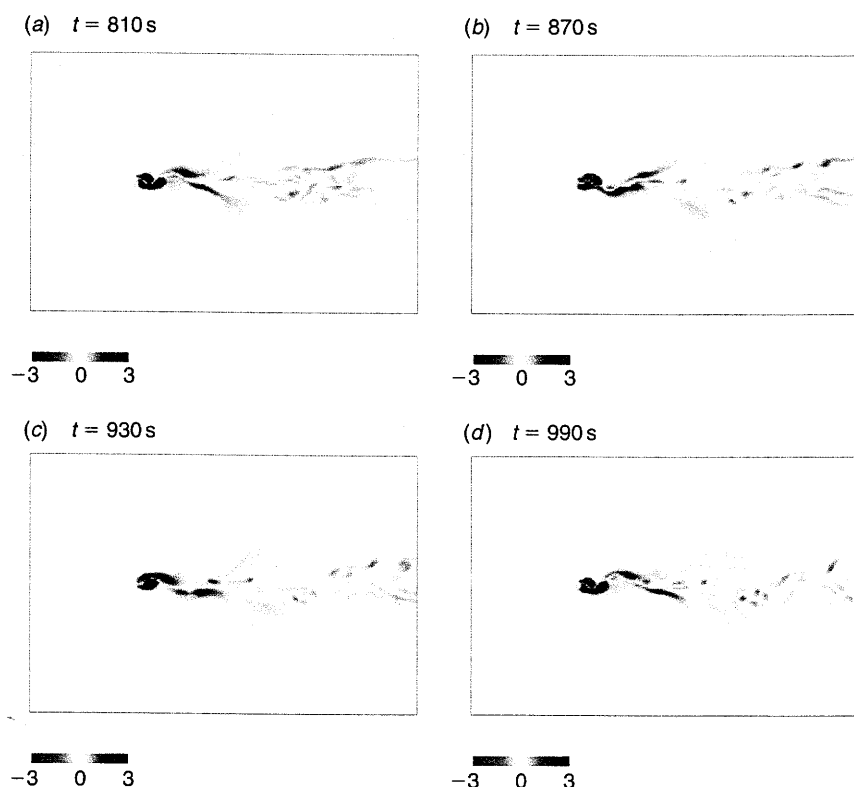


Fig. 8. Vertical component of vorticity (in s^{-1} , shaded as indicated) in the x - y plane at $z = 45$ m at (a) $t = 810$ s, (b) $t = 870$ s, (c) $t = 930$ s, and (d) $t = 990$ s for the circular heat source LES.

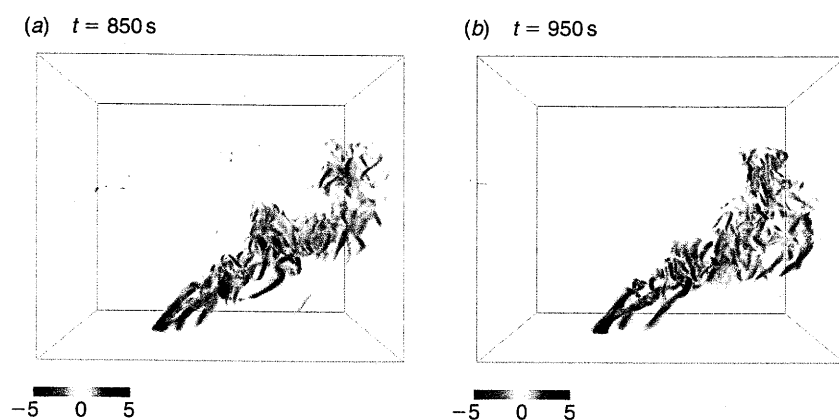


Fig. 9. Volume rendered vertical component of vorticity (in s^{-1} , shaded as indicated) at: (a) $t = 850$ s and (b) $t = 950$ s for the circular heat source LES.

Ghoniem 1993; McGrattan *et al.* 1996; Trelles *et al.* 1999), are able to treat only one of the three components of vorticity, namely the streamwise component. Consequently, these models are able to represent only the horizontal vorticity in the counter-rotating vortex pair. Nevertheless, since that is the dominant vortical structure in most buoyant plumes, these models are quite successful at predicting downwind plume concentrations, and they have the advantage over the present model that they can simulate larger downstream domains while using similar resolutions.

It is also noteworthy that the presence or absence of the wake vortex system depends on the effective Reynolds number of the simulations, and for low Reynolds numbers the

wake region is essentially laminar and symmetric. The wake vortices are observed only for the higher Reynolds number direct simulations and the LES. We are unaware of any previously published results describing simulations of the wake vortex system for buoyant plumes in a cross flow; indeed, despite the large number of numerical studies of jets in a cross flow, simulations of the wake vortex system in that case are apparently also rare (e.g. Rudman 1996).

As a final note in this section, there was no apparent evidence in any of the simulations for the fourth vortical structure depicted in Fig. 1, namely the horseshoe vortex. This is not entirely surprising, however, since these features are almost entirely located within the cross-flow boundary layer

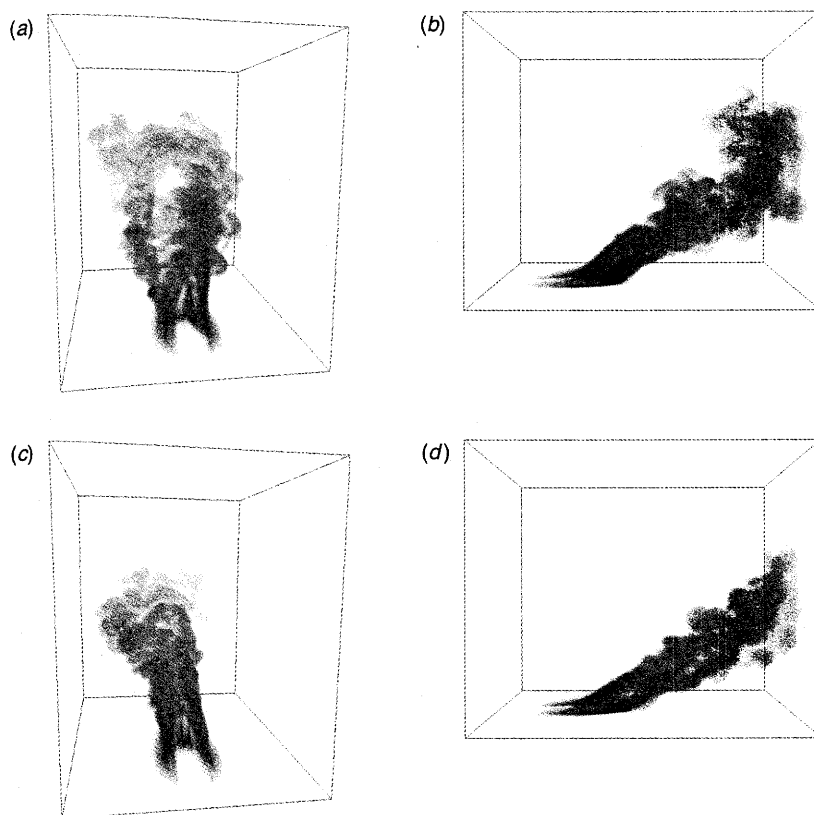


Fig. 10. Volume rendered potential temperature at $t = 1100$ s for the realistic heat source LES. The upper two images are for the deep shear layer (i.e. $z_0 = 150$ m) case with views from (a) the inflow boundary and (b) the lateral boundary. The lower two images are for the shallow shear layer (i.e. $z_0 = 50$ m) case with views in (c) and (d) identical to those in (a) and (b) respectively. Darker shades represent higher values of potential temperature.

and the resolution of the present simulations is inadequate to resolve the details of the near-surface flow.

Buoyant plumes from a heat source representative of a wildland fire

The preceding results demonstrate the ability of the model to reproduce the dominant features of buoyant plumes in a cross flow associated with circular heat sources. As noted previously, however, wildland fires represent a far more complex heat source with regions of heating that are essentially aligned in the streamwise direction (the fire flanks) and regions of heating that are aligned in the spanwise, or transverse, direction (the fire's head and back). The elliptical heat source depicted in Fig. 2b provides an idealized representation of a wildland fire with intensities increasing along the fire flanks towards the head. Using this heat source for cases with $z_0 = 50$ m and 150 m, referred to as shallow and deep shear layers, respectively, we employed the LES version of the model to examine its ability to simulate vortices associated with wildland fires as described by Haines and Smith (1987).

The first vortex type described by Haines and Smith was a horizontal vortex pair located along each fire flank. Figure 10 clearly shows the development of a dominant structure along each flank for both the deep (Fig. 10a) and the shallow (Fig. 10c) shear layers. While the transient nature of the flow renders the direct comparison of simultaneous instantaneous

images from the two simulations rather difficult, the plume in the deep shear layer case maintains a more pronounced horizontal split between the structures of each flank than does that in the shallow shear layer case, consistent with the results shown in Fig. 6. The vertical spread of the plume at the downwind edge of the model domain is also greater for the deep shear layer; however, as with the bifurcated plumes in the circular heat source cases, the mean plume height is nearly identical for the two shear-layer depths (compare Fig. 10b with Fig. 10d).

The presence of counter-rotating vortex pairs along each flank is more clearly evident by viewing the vorticity field (Fig. 11). Vorticity maxima are located within the flanks of the fire, and examination of the vertical component of vorticity shows the counter-rotating nature of these vortices (Fig. 11b,d). Near the heat source the plume is composed of the dominant counter-rotating vortex pairs along each flank as well as smaller vortex pairs located along the fire perimeter. For the deep shear layer case, wake vortices are also evident just downstream of the heat source and extend from the underside of the bent-over plume down to the surface. As in the circular heat source case, positive vortices are preferred on the right-hand side downwind of the heat source, while negative vortices are preferred on the left-hand side. Wake vortices are not as obvious in the shallow shear layer case due to their transient nature; however, small and relatively

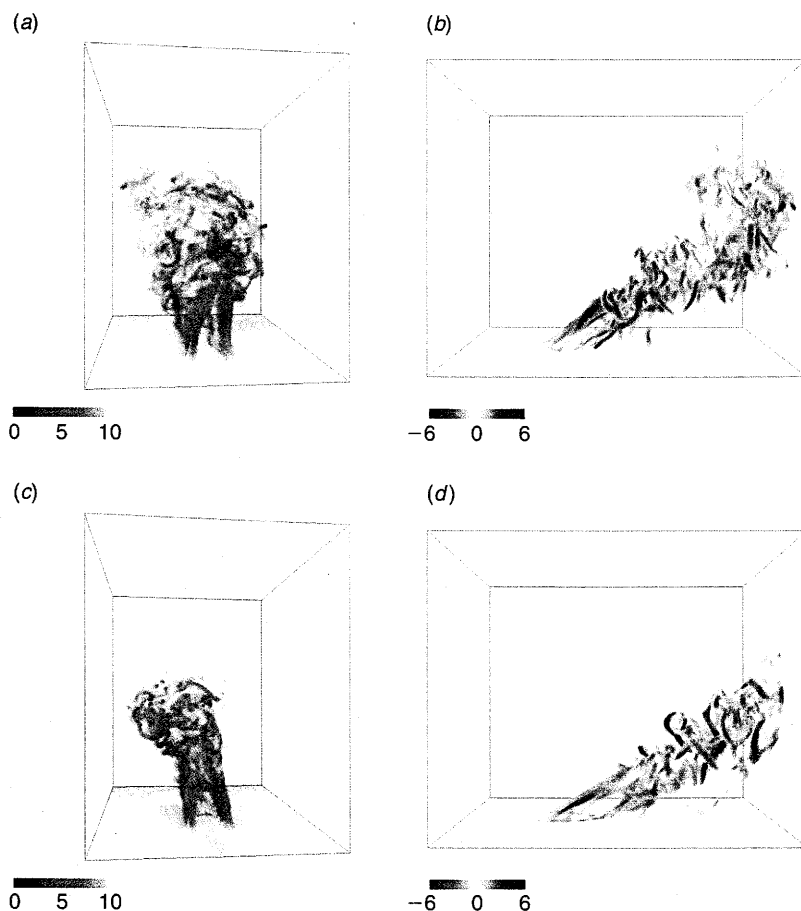


Fig. 11. Volume rendered vorticity magnitude (left column; in s^{-1} , shaded as indicated) and vertical component of vorticity (right column; in s^{-1} , shaded as indicated) at $t = 1100$ s for the realistic heat source LES for: (a) and (b) $z_0 = 150$ m; (c) and (d) $z_0 = 50$ m.

weak regions of vertical vorticity that are possible remnants of wake vortices are observed near the surface and downwind of the heat source.

The mechanism for producing the counter-rotating vortex pairs has been discussed in the previous section, and involves the interaction of the cross-flow shear-layer vorticity with the vorticity of the plume. However, it is to be expected that the details of this behavior may be significantly different for a spatially complex heat source. Examination of vortex lines at early stages of the shallow shear layer simulation (Fig. 12) reveal that the dominant counter-rotating vortex pairs along each fire flank are generated by the tilting of horizontal vorticity associated with the cross flow into the vertical to form loops in the vortex lines (Fig. 12b). As these loops are stretched in the vertical, the cross flow also acts to reorient them into streamwise vortices. As they near the head of the fire, however, the loop vortices from each flank begin to bend towards the region of most intense heating and become intertwined with the horizontal vorticity created by buoyancy at the head of the fire, creating an elevated horizontal vortex (Fig. 12c,d). This mechanism of concentrating horizontal vorticity at the head of fire may be analogous to the transverse vortices described by Haines and Smith (1987),

since there appears to be some degree of periodicity to their formation, and they propagate along the underside of the plume. Nevertheless, more investigation of this phenomenon is required.

Using an idealized representation of a wildland fire as the heat source we were able to reproduce two of the three vortices observed by Haines and Smith (1987) in association with wildfires (longitudinal counter-rotating vortex pairs and the transverse vortex). The third vortex discussed by Haines and Smith (1987)—a single longitudinal vortex—is thought to require directional shear in the ambient flow, which was not included in the current simulations. In addition, Haines and Smith (1987) described the bifurcation of the smoke column, characterized by a separation of the counter-rotating vortex pairs along each flank, for a crown fire in Wisconsin (the New Miner Fire). The simulations presented here suggest that the depth of the shear layer may play a significant role in the bifurcation, since the plume is noticeably wider for the deep shear layer than for the shallow shear layer. The split column extends well outside the fire flanks and displays an oscillatory nature, and we hypothesize that this behavior may result in the plume dropping to the ground as observed in the New Miner Fire if either the shear layer was deeper or the ambient

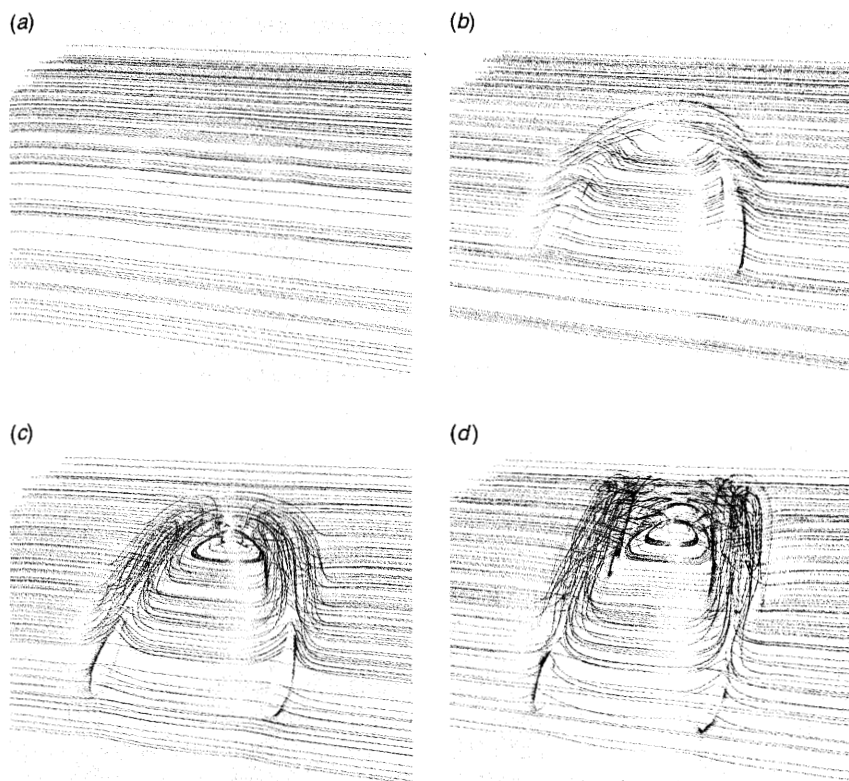


Fig. 12. Vortex lines and the 0.9 s^{-1} isosurface of vorticity magnitude at (a) $t = 10 \text{ s}$, (b) $t = 30 \text{ s}$, (c) $t = 50 \text{ s}$, and (d) $t = 70 \text{ s}$ for the realistic heat source LES with $z_0 = 50 \text{ m}$.

stratification was unstable instead of neutral, as is the case for the present simulations. Exploring the possibility will be a subject of future simulations.

Summary and discussion

In this paper, we have described several three-dimensional numerical simulations of buoyant plumes arising from isolated, finite-area heat sources in the presence of a vertically sheared cross flow. The numerical model is based on the compressible Navier–Stokes equations and is solved using both direct and large-eddy simulation techniques in the framework of the Weather Research and Forecasting (WRF) model code (Skamarock *et al.* 2001), modified to include a volumetric heat source and surface drag. Two types of heat source were examined: (i) a simple circular (smoothed top-hat) heat source; and (ii) an asymmetric elliptical ring heat source—referred to as the ‘realistic’ heat source—that represents an idealization of a typical wildland fire, with intensities increasing along the fire flanks towards the head. Of particular interest in the present study was the extent to which the simulations were able to capture the variety of coherent vortical structures commonly seen, not just in wildland fires but more generally in buoyant plumes in a cross flow arising from a non-specific buoyancy forcing. These vortical structures control the turbulent mixing in the plume and, in the case of wildland fires, can potentially lead to erratic fire behavior if they are strong enough to interact with the fire (e.g. Haines and

Smith (1987); see also the coupled atmosphere–fire simulations of Clark *et al.* (1996a, 1996b) and Linn *et al.* (2002)).

The plumes arising from circular heat sources in the presence of a cross flow exhibit many similarities with non-buoyant circular jets in a cross flow, particularly with respect to the types of coherent vortical structures present in these flows, although there are some differences related to the presence of buoyancy and the associated generation of horizontal vorticity. Specifically, the simulations presented herein were able to capture:

- (1) The counter-rotating vortex pair aligned with the plume trajectory that is responsible for the bifurcation of the plume seen in many observed cases of wildland fires and other sources of buoyancy;
- (2) The shear-layer (i.e. transverse) vortices that arise from a Kelvin–Helmholtz type instability on the upstream face of the plume; and
- (3) The vertically oriented wake vortices that form with alternating sign and regular period in the lee of and beneath the bent-over plume.

These simulations also suggest that the degree to which the plume bifurcates is related to the depth of the cross-flow shear layer (a property also of the plumes arising from the realistic heat source), and that the periodicity of the wake vortices is linked to a periodic oscillation of the plume bifurcation in

which each branch of the bifurcation is successively dominant; however, more analysis is required in this regard to further characterize these simulated features.

The simulated plumes arising from the realistic heat source exhibit vortical structures that, although related to those seen in association with the circular source, are configured in a dramatically different manner. In particular, counter-rotating vortex pairs oriented in the streamwise direction form along each flank of the heat source, and transverse horizontal vortices form above the head of the heat source. The results of these simulations compare favorably with observations of plumes in wildland fires (e.g. Haines and Smith 1987).

The results presented in this paper have broad relevance to the dynamics of buoyant plumes in a cross flow, and the primary purpose of this paper is to illustrate the potential utility of numerical models to explore these phenomena. A considerable amount of additional analysis is now required to be able to draw firm conclusions about the detailed dynamical nature of the structure and evolution in these cases, particularly with respect to the identification of parameters that can be used to describe and characterize the flow over a wide range of conditions. Nevertheless, the present results are highly encouraging, and highlight many avenues for future research.

Finally, it is important to note that none of the simulated plumes shown in this paper displays a Gaussian cross-section; nor indeed have we seen Gaussian cross-sections in any of our simulations. Moreover, initial attempts to calculate and characterize the time-mean properties of the plume (not shown) have not yielded Gaussian structure either. Although the presence of bifurcation in the plume does not appear to affect plume rise significantly, it can be expected to play a critical role in smoke transport and dispersion, as may the other types of vortices seen in the simulations. The manner in which these vortical structures affect the transport and dispersion of smoke is not immediately apparent and requires further study, but the absence of self-similar Gaussian structure in the simulations shown here would appear to call into question the common practice of using plume rise and dispersion models that are based on these assumptions. From the perspective of the transport of particulate matter, the updrafts found along the centerline of the plume associated with the counter-rotating vortex pair may be strong enough to keep fine particles with small terminal velocities airborne long enough to prevent them from being deposited with a maximum concentration along the centerline, but instead they may be transported downward by the counter-rotating vortex pair on the flanks of the plume. In cases where the plume bifurcates, the majority of particles may be channeled into one of the branches. Such internal plume circulations might possibly lead to different distributions between particulate matter of different sizes (e.g. between PM_{2.5} and PM₁₀ larger than 2.5 μm), and as such it is conceivable that present

models are providing inaccurate predictions of the transport of fine particulate matter. With respect to the wake vortex system, it is of interest to examine the possibility that pollutants can be entrained from the plume into the wake downstream and beneath it via these vortices, leading potentially to localized but large instantaneous concentrations of smoke at the surface.

In light of the foregoing discussion, we suggest that an important aspect of future research is to assess the applicability of traditional plume models that assume a self-similar Gaussian structure, and to use a numerical model such as that described herein to explore alternative strategies for simplified plume modeling.

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The role of released moisture in the atmospheric dynamics associated with wildland fires

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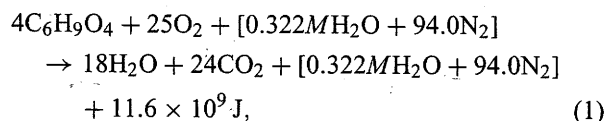
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Abstract. Combustion of woody material produces and releases water, but the effects of this water on the atmospheric circulation created by a wildfire are rarely recognized, let alone understood. This paper presents observational data and basic physical arguments to support the hypothesis that this moisture can constitute a large portion of the total water content in a fire plume. Calculations demonstrate the effects this moisture could have on fire-driven atmospheric circulations, specifically updrafts and downdrafts, on time and space scales important for fire behavior and fire-fighter safety. This study should be considered exploratory; it does not prove the presence or importance of this moisture, but seeks to show that further study is needed to determine how much moisture a fire adds to the air, and whether that amount is or is not important.

Additional keywords: combustion; convection; downdraft; plume.

Introduction

Wood is largely a hydrocarbon. Byram (1959) states that the 'average' chemical composition is approximately $C_6H_9O_4$. He then provides an equation for the complete combustion of wood, restated in SI units by Johnson and Miyanishi (2001) as



where M is the percentage moisture content of the wood. The bracketed quantity represents moisture in the fuel and nitrogen in the air, which as Byram states, 'should be indicated in the combustion equation even though [they are] chemically inert'. This equation is a simplification, as there are far more compounds released by burning wood than just water and carbon dioxide. Also, while moisture in the fuel may not change from water to another compound, it is important to note that, after the fuel burns, the water is no longer bound in the wood, but is free vapor. Note that water in the fuel is usually the lesser part of the total water on the right-hand side of the equation. Even if wood were to burn with a moisture content of 30%, this water would constitute only 35% of the water on the right-hand side. Actual fuel moistures are often in the 10% range.

The combustion equation is central to this paper's thesis: burning woody material puts water into the atmosphere. This moisture can be a large portion of the water present in

the fire's updraft plume, and it can have a substantial impact on the atmospheric processes and dynamics, which in turn affect fire behavior. In this paper, combustion moisture will henceforth refer to the water molecules produced during combustion of woody material. This is distinct from fuel moisture, which is water before the combustion process and is contained in the fuel tissue. Released moisture refers to the combination of fuel moisture and combustion moisture entering the atmosphere.

The production of moisture by burning woody material is generally accepted, even if it is rarely considered explicitly. What is less accepted is that this moisture can influence how the fire and atmosphere behave. In discussing forest fire combustion, Byram (1959) mentioned that the latent heat of vaporization, an energy sink near the fire, may indeed serve as a source of energy for the atmosphere at higher altitudes under certain conditions. He stated that, at the time of writing, it was impossible to determine how important this may be or what effect it might have on fire behavior. Since Byram's comment, there seems little acknowledgment of the presence of combustion moisture except when someone is computing the low heat of combustion and subtracts the latent heat of vaporization from the high heat (e.g. Wade and Ward 1973; Simard *et al.* 1983). In these instances, the latent heat is considered removed from the fire system and no longer relevant. A few researchers have noted increased atmospheric moisture in fire plumes (e.g. Stocks and Flannigan 1987; Goens and Andrews 1998) and that this

moisture may have influenced the atmospheric dynamics and fire behavior, but these statements are not based on direct observation of that moisture. While some individuals recognize that released moisture exists and can affect fire behavior, there is no general acknowledgement of this potential by the research community as a whole.

Coupled fire-atmosphere models like those of Clark *et al.* (1996) and Linn *et al.* (2002) are in many ways the state-of-the-science in fire-atmosphere interaction research. To date, these have included water only as a sink of energy in calculating the spread of fire, or as the water affects average air density. They do not track the water as a separate substance as it goes into the atmosphere, though they do have this capability. None of the commonly used fire weather indices (Haines, Fosberg, Canadian Fire Weather Index, for example) includes any consideration of released moisture. They include a representation of ambient atmospheric dryness, but have no explicit representation of how any added moisture might influence fire behavior.

This paper argues in support of two hypotheses. The first is that released moisture, most of which is combustion moisture, can constitute a large portion of the total water content in a fire's plume. Observational evidence provides support for this claim. The second hypothesis is that released moisture influences the updraft dynamics of fire plumes and other aspects of the fire-driven atmospheric circulation such as downdrafts. Released moisture is not only a contributing factor, but at times a controlling or critical factor in fire-atmosphere interactions on time and space scales important for understanding fire behavior and fire-fighter safety.

The intent in this paper is to show that moisture can be important, but not that it is always important. For clarity I will repeat Byram's statement that, in the case of small fires, the latent heat of vaporization is essentially lost from the fire. Only for a fire large enough that condensation occurs in the updraft, under conditions where the convective column maintains a significant core of air unaltered by entrainment, would many of the following arguments apply. This paper ignores the effects of wind and wind shear; both of these would increase entrainment rates, and strong winds can remove any potential latent heat release far enough from the fire region that the subsequent effects on fire behavior are negligible.

Observational support

Several types of observations can provide indirect or direct evidence of moisture being injected into the atmosphere by a wildfire. These include plume characteristics, atmospheric stability and moisture content, and the occurrence of precipitation. Taken alone or in combination, these factors can support the hypothesis that a fire adds significant amounts of moisture to the ambient atmosphere.

The strongest evidence stems from plume characteristics (specifically, estimated level of condensation in the plume) compared to ambient atmospheric surface moisture.

Comparing the plume height with ambient surface air equilibrium height is less conclusive because either heat or moisture added by a fire can determine plume height. The occurrence of precipitation seems, at first glance, as if it would be strong evidence of moisture added by the fire. However, moist surface air, raised high enough, will produce enough condensation to allow for the possibility of rain developing.

Observations from four fires follow. The arguments presented are similar in each case, but all are included to reinforce the underlying premise that released moisture can be comparable in magnitude to environmental moisture, and can have a significant impact on fire dynamics.

The Mack Lake Fire took place in Michigan on 5 May 1980. Simard *et al.* (1983) state that the fire's convective plume rose to 4600 m (15 000 feet) and that skies were otherwise clear. In their fig. 24 photograph, there is a clear pyrocumulus cloud atop the fire plume. Based on a visual estimate, the pyrocumulus begins at a height no higher than 2300 m—therefore 2300 m is an estimate of the lifting condensation level (LCL) for the air in this plume. From the 00Z 6 May 1980 atmospheric profile at Flint, MI (the nearest upper air observation site, 185 km away), and with surface temperature (28.3°C) and dew point (4.6°C) observed at the site and time of the fire, the LCL of the environmental air was 3160 m. To lower the LCL to 2300 m would require adding more than 4 g kg^{-1} of water to the air. This is only an estimate but, even if the LCL were 500 m higher, it would require about a 1.5 g kg^{-1} increase over the ambient mixing ratio.

As a second case, consider the Air Force Bomb Range fire in North Carolina (Wade and Ward 1973). This report lacks a photograph or any information indicating the LCL in the fire plume, but it does note a plume height of approximately 4600 m (15 000 feet). The sounding from Cape Hatteras, NC (56 km away) shows the ambient environment had an equilibrium height of 2760 m. Reaching 4600 m would have required the addition of over 3 g kg^{-1} of moisture or a temperature increase greater than 3°C, or some combination of moisture and temperature increase.

The Red Lake #7 Fire (Stocks and Flannigan 1987) took place in north-western Ontario, Canada, in May 1986. On the afternoon of 29 May, the authors note that the plume rose 13 000 m above the fire. It produced thunder and lightning and observers reported 5–10 mm of precipitation on the ground, falling from the convective column produced by the fire. The nearest upper air sounding (International Falls, Minnesota, 287 km away) indicates an environmental equilibrium height of 9900 m. For near-surface air to reach as high as just 12 000 m required adding at least 3 g kg^{-1} , or more than 3°C, or some combination of moisture and temperature increase.

Latham (1994) shows a photograph of the plume from the Sandpoint fire in Montana on 5 July 1985. The plume reportedly reached a height of 9100 m (30 000 feet). The environmental air, based on a sounding from Great Falls, MT

(77 km away) that afternoon, had an equilibrium height of 5100 m. Even if the surface air temperature had been raised by 3°C, the equilibrium height would only be 6100 m. However, adding 2 g kg⁻¹ of moisture would have changed the equilibrium height to 10 800 m; adding 1 g kg⁻¹ of moisture and increasing surface air temperature by 1°C would have raised the equilibrium height to 10 100 m.

These are specific cases, selected from the available literature where the needed information is provided. They are anecdotal, and in no way 'prove' that released moisture was a large portion of the total water content in the plumes, or that released moisture plays a significant role in the atmospheric dynamics of a fire plume. They do, however, suggest this possibility. In fairness, there are reports that contain information on plume heights or LCLs but do not support the argument that released moisture can be important—though they do not refute such a role, either. Specifically, the Robie Creek Fire (Small 1957) and the Basin Fire (Chandler 1961) reports both include plume height information. The Robie Creek Fire report also includes an estimated LCL. However, in each case, the estimated plume height is substantially lower than the equilibrium height one obtains from the nearest upper air sounding. If the ambient air could have risen higher than the plume was estimated to rise, there is little insight to be gained by arguments based on adding energy. One could also consider the estimated plume heights to be inaccurate, in which case the estimated LCL is of little or no use for the present arguments.

There is one additional source of observational evidence clearly showing that smoke contains much more moisture than the ambient air. Achtemeier (2003) describes very simple measurements of dew point temperature taken near and within the small smoke plumes from smoldering patches after a prescribed burn. He reports dew point temperature increases of 6°C to 20°C in the plume compared to the environment. Using Achtemeier's reported ambient dew point of 13.9°C and assuming atmospheric pressure of 100 kPa, the ambient mixing ratio in this case was 10.1 g kg⁻¹. Air with an 11°C dew point increase would have a mixing ratio of 20.3 g kg⁻¹, and air at the highest dew point temperature observed, 38°C, would have a mixing ratio of 44.2 g kg⁻¹. These observations are from smoldering combustion, and it is likely that complete combustion would have released still more moisture.

In summary, the observations described above all indicate that moisture could have had an impact on the height and strength of the convective column over the fires examined. For the large fires discussed, estimated values of the amount of moisture required to produce the observed plumes range from approximately 1 to 4 g kg⁻¹. For the smaller fires studied by Achtemeier (2003), the moisture amounted to 10–30 g kg⁻¹. For the remainder of this paper, I will assume that any moisture release from a large fire has a net impact of 1–3 g kg⁻¹.

Implications for atmospheric dynamics and energetics

People often compare fire convection with free atmosphere storm convection. This is based largely on the strong updrafts, the winds these updrafts generate, and the towering clouds often seen in both cases. The range of parameters known to correlate with storm cell intensity (Weisman and Klemp 1986), and likely to matter for fire convection, includes atmospheric instability, wind speed, directional wind shear, total convective available potential energy (CAPE), the vertical distribution of that energy, the downdraft convective available potential energy (DCAPE, as defined by Emanuel 1994), and more.

The present discussion will examine only CAPE and DCAPE as they depend strongly on the moisture content of the air and that is the focus of this paper. Many studies over the past 50 years have examined or discussed the possible importance of lower-atmosphere instability (e.g. Byram and Nelson 1951; Small 1957; Davis 1969; Brotak and Reifsnnyder 1977; Haines 1988). Ambient instability appears important, but the reader should also consider the stability implications of released moisture as he or she reads the following sections of this paper.

Implications for CAPE

While the basic concept of CAPE is straightforward, several minor variations can appear in its calculation. Often, the average properties in a layer of air extending from the ground to some height, such as 500 m or 1000 m, represent the potential temperature of the rising air. For fires, where inflow air is likely to come from near the ground, these deep layers may be inappropriate, so the following discussion uses only surface air temperature in computations of CAPE.

Potential temperature frequently represents the environmental temperatures in computing parcel buoyancy. Because this study focuses strongly on moisture, the most relevant way to calculate CAPE is by using virtual potential temperature to describe the densities of both the environment and the parcel. This results in the following equation for CAPE:

$$\text{CAPE} = g \int_0^{z_{\text{EL}}} \frac{\theta_v(z) - \bar{\theta}_v(z)}{\bar{\theta}_v(z)} dz. \quad (2)$$

Here g is the acceleration due to gravity; z_{EL} is the equilibrium height (level of neutral buoyancy) for surface air; $\theta_v(z)$ is the virtual potential temperature of surface air after it has ascended to height z (with condensation if appropriate); and $\bar{\theta}_v(z)$ is the virtual potential temperature of the environment.

The earlier discussion noted that a fire modifies air entering its plume, increasing its temperature and mixing ratio by several degrees or grams per kilogram, respectively. Increasing either of these will increase the CAPE and z_{EL} , but the magnitude of the increase depends strongly on the initial atmospheric profile. If surface air in the initial sounding does not reach saturation as it rises to z_{EL} , and air with an increased

Table 1. Fires used in CAPE and DCAPE calculations

Dates listed do not necessarily reflect the duration of the fire, but the days examined in this study. Atmospheric soundings used for calculations of convective available potential energy (CAPE) and downdraft convective available potential energy (DCAPE) are taken for the time closest to the time of the fire's noted plume development or behavior, and representative of the airmass in which the plume developed as discussed in the reference

Fire	Day	Nearest upper air station (distance from fire)	Reference
Robie Creek	5–9 Sept. 1955	Boise, ID (19 km)	Small (1957)
Gaston	1 Apr. 1966	Charleston, SC (143 km)	DeCoste <i>et al.</i> (1968)
Sundance	1 Sept. 1967	Spokane, WA (112 km)	Anderson (1968)
Air Force Bomb Range	22 Mar. 1971	Hatteras, NC (56 km)	Wade and Ward (1973)
Bass River	22 July 1977	John F. Kennedy Airport, NY (138 km)	Smith (1978)
Mack Lake	5 May 1980	Flint, MI (185 km)	Simard <i>et al.</i> (1983)
Sandpoint	5 July 1985	Great Falls, MT (77 km)	Latham (1994)
Garibaldi	22 Aug. 1986	Sault Ste. Marie, MI (241 km)	McRae and Stocks (1987)
Lowman	26 July–6 Aug. 1989	Boise, ID (82 km)	Werth and Ochoa (1993)
Dude	26 June 1990	Tucson, AZ (256 km)	Goens and Andrews (1998)
Red Lake #7	28–30 May 1986	International Falls, MN (287 km)	Stocks and Flannigan (1987)

mixing ratio but unchanged surface temperature does not reach saturation before it reaches its z_{EL} , then there will only be minor changes in CAPE and z_{EL} due to the change in virtual temperature. If, however, the added moisture leads to the parcel saturating as it rises, while it did not saturate in the initial profile, then there can be a substantial increase in the parcel's CAPE and z_{EL} . If the fire increased only surface air temperature and not moisture, then CAPE and z_{EL} would increase. The largest changes in CAPE and z_{EL} will occur when the temperature increase leads to saturation and moist ascent, where the original temperature profile did not saturate.

I examined several large, well documented fires, listed in Table 1. Some fires were single-day and some were multiple-day fires. In the latter case, I sometimes considered more than one of the days and sometimes only one day. This depended on the nature of the information available from the documentation available, in that I sought days with stated plume heights or other detailed fire behavior observations. For each fire and day considered, I computed the atmospheric CAPE using equation (2), as well as the CAPE for a parcel of surface air warmed by 2°C or 3°C, and/or moistened by 2 g kg⁻¹ or 3 g kg⁻¹. These computations relied on the nearest upper air site, and the nearest surface observation for the time of the fire replaced the surface observations from the sounding. In cases where a front passed during the fire, I chose the previous or following sounding that represented conditions in the frontal sector where extreme fire behavior was noted. For example, if extreme behavior was reported at 2000 UTC and a front passed the area at 1600 UTC, I used the following 0000 UTC sounding so that the sounding would indicate post-frontal atmospheric conditions.

Table 2 shows the results these changes in temperature or moisture had on CAPE, expressed in the form of differences from the basic profile's CAPE. There is a wide range in the base state CAPE values for these fires, from 1 J kg⁻¹ for the Garibaldi fire (essentially a stable atmosphere near the ground) to 2547 J kg⁻¹ for the Dude fire; the average for

the days considered is ~420 J kg⁻¹. A CAPE of 800 J kg⁻¹ is capable of producing severe weather, while 2000 J kg⁻¹ can produce a supercell storm (McCaul and Weisman 2001). In all cases, the effect of an increase in temperature or moisture is to increase the CAPE. Adding 2°C increases CAPE by ~310 J kg⁻¹, on average, and adding 3°C raises CAPE by an average 570 J kg⁻¹. Raising surface air moisture by 2 g kg⁻¹ or 3 g kg⁻¹ increases the average CAPE by 340 J kg⁻¹ or 620 J kg⁻¹ respectively. The effect of combined increases in moisture and temperature is often greater than the sum of the two individual increases. The CAPE increases produced by these temperature or moisture alterations generally yield total CAPE values typical of severe weather, and sometimes above the highest values even considered for non-fire storm convection (e.g. the CAPE of a parcel warmed 3°C and moistened 3 g kg⁻¹ using the Dude fire sounding would be 5084 J kg⁻¹). The Garibaldi fire stands out as the one fire where even 3°C and 3 g kg⁻¹ failed to produce storm-like CAPE.

Implications for descending air (DCAPE)

In storms and fires, the updraft drives the development of a downdraft somewhere. If winds tilt the convective column substantially, the downdraft may occur far downwind from the fire itself. For a plume that is more or less vertical, the downdraft air may reach the ground close enough to influence the fire's behavior. Goens and Andrews (1998) found that a downburst—an unusually intense downdraft—was largely responsible for the entrapment and fatalities on the Dude fire.

While CAPE is a rather well-known quantity in the atmospheric sciences, downdraft convective available potential energy, DCAPE, is less well known. Emanuel (1994) proposed DCAPE as an indicator of the amount of energy available for a parcel of air starting at some height z_0 and descending to the surface, remaining saturated at all times. The concept assumes that the water necessary to maintain saturation is available in the form of cloud and/or rain falling into the parcel as it descends. The continuous evaporation

Table 2. Computed values of CAPE for observed atmosphere and moistened and/or warmed surface air for fires listed in Table 1
All convective available potential energy (CAPE) and Δ CAPE values are in units of J kg^{-1}

Fire	Day	CAPE	Δ CAPE					
			+2°C	+2 g kg ⁻¹	+2/+2	+3°C	+3 g kg ⁻¹	+3/+3
Robie Creek	5 Sept. 1955	1353	+471	+686	+1197	+716	+1096	+1900
	6 Sept. 1955	1147	+453	+641	+1098	+684	+997	+1694
	7 Sept. 1955	522	+450	+662	+1137	+684	+1047	+1787
Gaston	1 Apr. 1966	304	+461	+832	+1357	+718	+1310	+2129
Sundance	1 Sept. 1967	74	+94	+15	+114	+154	+23	+1520
Air Force Bomb Range	22 Mar. 1971	171	+151	+100	+266	+232	+199	+783
Bass River	22 July 1977	564	+196	+89	+755	+301	+683	+1412
Mack Lake	5 May 1980	466	+488	+866	+1413	+749	+1366	+2193
Sandpoint	5 July 1985	231	+286	+375	+922	+469	+795	+1668
Garibaldi	22 Aug. 1986	1	+13	+1	+19	+32	+1	+44
Lowman	26 July 1989	80	+159	+19	+200	+277	+31	+1754
	27 July 1989	113	+176	+28	+1568	+1005	+43	+2324
	28 July 1989	48	+175	+28	+2144	+1517	+42	+2893
	29 July 1989	836	+581	+949	+1556	+883	+1460	+2371
	30 July 1989	147	+225	+35	+1137	+655	+1028	+1854
	31 July 1989	118	+200	+32	+742	+322	+49	+1421
	1 Aug. 1989	118	+164	+28	+198	+258	+42	+563
	2 Aug. 1989	115	+189	+228	+586	+295	+508	+1090
	3 Aug. 1989	0.1	+2.5	+0.4	+3.2	+4.4	+0.6	+1127
Dude	26 June 1990	2547	+613	+926	+1619	+939	+1474	+2537
Red Lake #7	28 May 1986	65	+157	+23	+238	+253	+36	+920
	29 May 1986	438	+481	+905	+1465	+737	+1438	+2307
	30 May 1986	5	+834	+6	+1850	+1114	+10	+2663

to maintain saturation cools the parcel as it descends, and holds the parcel at a constant wetbulb potential temperature, θ_w . The same arguments about buoyancy and air density that suggested using θ_v in equation (2) for CAPE lead to using it to compute DCAPE:

$$\text{DCAPE} = g \int_{z_{\text{sfc}}}^{z_0} \frac{\bar{\theta}_v(z) - \theta_v(z)}{\bar{\theta}_v(z)} dz. \quad (3)$$

Computation of DCAPE requires specifying a starting level for the descending air. It is unclear what serves as the source region of any descending air in a storm or fire situation. For this study, I have computed DCAPE for parcels up to 4 km above ground level (AGL) for each fire at 500 m intervals, and noted the greatest value. This method yields a maximum DCAPE possible for air between the ground and 4 km AGL. The choice of 4 km here is based on comments in Gilmore and Wicker (1998), and is meant as only an approximation.

Because DCAPE is less well known than CAPE, several points bear mention here. First, while θ_v appears in equation (3), the computation could just as easily be done with θ . Whether this yields greater or smaller values of DCAPE depends strongly on the moisture profile of the environment. Air descending from a very dry level through a moist environment will have greater DCAPE if computations use θ_v and, conversely, moist air descending through a dry environment will have a smaller DCAPE if computations use θ_v . Frequently, atmospheric profiles during severe fires exhibit

extreme dryness near the ground, thus falling into the latter category (i.e. DCAPE using θ_v is smaller than DCAPE using θ). Another way of saying this is that consideration of the buoyancy effects of water vapor on parcel energy in a typical fire profile reduces DCAPE. (This is the opposite of moisture's effect on CAPE through buoyancy.) Srivastava (1985) saw this influence of profile moisture on downdraft strength in simulations using a numerical model.

Now consider the effect of entrainment on DCAPE, particularly in a situation where θ_w increases with height. If a parcel of air descends into and entrains air with lower θ_w , maintaining saturation will require evaporation of additional moisture, and result in a parcel-average θ_w lower than the original parcel θ_w . This increases the numerator in equation (3) and therefore leads to a higher value of DCAPE.

Finally, DCAPE is a theoretical quantity and rarely, if ever, fully realized. The main limit on this is the availability of liquid water for the evaporation necessary to maintain saturation during descent. Insufficient water will keep the descending air from cooling through evaporation, decreasing the parcel's negative buoyancy. In a fire situation, it is even more probable that water will be in short supply than it is in a storm situation. Just where and when a limited supply of water evaporates during a parcel's descent determines the actual energy of the parcel when it reaches the ground. The higher above ground a given amount of water evaporates in a parcel, the farther that air will descend with a lower temperature, and the greater its DCAPE.

Table 3. Computed values of maximum DCAPE between the ground and 4 km AGL for observed atmosphere for fires listed in Table 1

All downdraft convective available potential energy (DCAPE) values are in units of J kg^{-1} , Δq_v values are in g kg^{-1} , and $\text{DCAPE}/\Delta q_v$ values are in J g^{-1}

Fire	Day	DCAPE	Δq_v	$\text{DCAPE}/\Delta q_v$
Robie Creek	5 Sept. 1955	983	7.1	138
	6 Sept. 1955	938	7.6	123
	7 Sept. 1955	1003	7.2	139
Gaston	1 Apr. 1966	708	8.0	89
Sundance	1 Sept. 1967	870	8.0	109
Air Force Bomb Range	22 Mar. 1971	385	6.3	61
Bass River	22 July 1977	646	7.2	90
Mack Lake	5 May 1980	867	7.5	116
Sandpoint	5 July 1985	1460	9.5	154
Garibaldi	22 Aug. 1986	165	5.5	30
Lowman	26 July 1989	1094	9.3	118
	27 July 1989	1385	9.9	140
	28 July 1989	1440	10.0	144
	29 July 1989	1513	10.0	151
	30 July 1989	1402	9.9	142
	31 July 1989	1371	10.1	136
	1 Aug. 1990	856	9.1	94
	2 Aug. 1990	535	6.7	80
	3 Aug. 1990	876	8.9	99
	26 June 1986	1508	7.4	204
Dude	26 June 1986	1508	7.4	204
	28 May 1986	822	9.7	85
	29 May 1986	1018	9.3	109
Red Lake #7	28 May 1986	822	9.7	85
	30 May 1986	1057	9.9	107

Because of the importance of the amount of water evaporated to produce a given amount of DCAPE, I computed not only the maximum DCAPE for air below 4 km AGL in each sounding, but also the change in water vapor mixing ratio necessary to achieve that DCAPE. The amount of DCAPE generated by evaporation of a given mass is, as noted above, not constant. However, the average mixing ratio change ($\text{DCAPE}/\Delta q_v$) for a descending parcel of air gives some indication of how the average DCAPE changes for each g kg^{-1} of change in the mixing ratio.

Table 3 shows the results of the DCAPE computations for the cases considered. The DCAPE values for the fire days range from 165 J kg^{-1} for the Garibaldi fire to 1513 J kg^{-1} for 29 July during the Lowman fire, with mixing ratio changes necessary to produce these DCAPE values between 5.5 g kg^{-1} (Garibaldi fire) and 10.1 g kg^{-1} (31 July during the Lowman fire). The average DCAPE produced by each g kg^{-1} change in mixing ratio for a parcel ranged from 30 J g^{-1} for the Garibaldi fire to 204 J g^{-1} for the Dude fire. Given the behavior documented for the Garibaldi fire, one would expect the atmosphere at that time to show extreme values of both CAPE and DCAPE. The prevailing conditions at the nearest radiosonde site do not exhibit these properties, and I have no substantial proof or information that would explain whether or why there may have been different conditions at Garibaldi than at Sault Ste Marie, Michigan.

In general, many of the DCAPE values in Table 3 are comparable to the CAPE values associated with mid-latitude

convective storms. This similarity suggests the potential for comparable (but opposite) vertical velocities in the two situations. The realization of these different types of potentials, however, depends on so many other factors that drawing any conclusion other than the similarity of the values is extremely tenuous.

The DCAPE values in Table 3 are interesting in their own right, but even more so considering the implications for released moisture. Perhaps the biggest difference between fire convection and storm convection is the moisture in the atmosphere for the two situations. The type of profile conducive to an air-mass thunderstorm or supercell storm is very moist near the ground, and derives much of its energy from the continuous ingestion of that moist air as it propagates across the landscape. The profiles related to many severe fire convection situations are extremely dry at low levels, and would be unlikely to foster a storm of any duration—scattered cumulus clouds and dry lightning are far more common with these profiles. A strong downdraft can kill a supercell storm simply by pushing the gust front too far ahead of the leading updraft (Gilmore and Wicker 1998). But with a severe fire, such a downdraft can lead to erratic, sudden changes in winds and fire behavior and can result in human fatalities. A downdraft requires some moisture to initiate and further moisture may be needed to sustain it all the way to the ground. Though that moisture is not commonly available from the ambient atmosphere, it may be available from released moisture. Sensible heat release at the

fire cannot by itself create a downdraft that would behave this way.

There are several patterns or associations that appear (and may or may not be meaningful) in the data shown in Tables 2 and 3. One association is that CAPE, DCAPE, and $\text{DCAPE}/\Delta q_v$, are all high for the Dude fire and the Lowman fire on 29 July. These are both days when extreme behavior occurred, and as noted earlier in the case of the Dude fire, Goens and Andrews (1998) suggest that a downburst was responsible for fatalities.

Generally speaking, the eastern fires in the tables (Gaston, Air Force Bomb Range, Bass River, Mack Lake, Garibaldi, and Red Lake) seem to have lower values of both CAPE and DCAPE than the western fires. Here I have arbitrarily chosen the Mississippi River to separate east and west, though the dataset is such that almost any longitude passing through the Dakotas would yield the same subgroupings. (Statistical analyses of these data are meaningless because of the obviously biased selection of the data and the small sample sizes.) The average eastern (unmodified) CAPE is 250 J kg^{-1} , while western CAPE is 510 J kg^{-1} ; average eastern DCAPE is 710 J kg^{-1} and average western DCAPE is 1150 J kg^{-1} . The higher western CAPE values result from higher surface sensible heating, which produces large differences between surface-layer air θ and the average θ of the mixed layer above it. In the east, higher available moisture and higher evapotranspiration rates reduce sensible heating and the surface-layer lapse rate, bringing surface θ closer to mixed-layer θ . Similarly, eastern profiles tend to be moister so there is less opportunity for evaporation to occur and DCAPE values would be smaller.

This raises a potentially important point of distinction between the western and eastern fire-atmosphere interactions and the conditions needed in each region to produce extreme fire behavior. Eastern fires may require more added moisture to yield CAPE or DCAPE values typical of extreme fire situations. Eastern forests and ecosystems, however, may be more able to provide that moisture. Consider that the moisture content cited for the Air Force Bomb Range fire (Wade and Ward 1973) is 70%, while that for the Sundance fire (Anderson 1968) is closer to 10%. The total moisture potentially released for the former would then be 1.26 kg of water per kg of fuel burned, while for the latter the potential release would be 0.66 kg of water per kg of fuel burned. The near doubling of available moisture from the eastern fire would allow greater realization of the calculated DCAPE value and may thus compensate for the lower theoretical DCAPE in the east. Similarly, CAPE for the Air Force Bomb Range fire is more likely to be close to the 3 g kg^{-1} value in Table 2, while CAPE for the Sundance fire would be closer to the 2 g kg^{-1} value or lower.

While the computations described above examine all levels of air below 4 km for DCAPE, and the discussion focuses on profiles that are generally dry through deep layers near the

ground, such a layer is not necessary for downdraft formation. All that is required is some layer of dry air above the ground and below the portion of the updraft containing precipitation. In fact, as Srivastava (1985) noted, DCAPE is greater in cases where a single dry layer exists above a humid layer with a dry adiabatic lapse rate. This situation produces the most potentially negatively buoyant air and hence the largest value of DCAPE.

Conclusions

I have shown that there is observational and theoretical support for the hypothesis that released moisture can be sufficiently plentiful to be an important portion of total fire-plume water mixing ratios. I have also shown that the addition of moisture to the air in updrafts for a selection of documented severe fires across the United States could substantially increase the CAPE of that air. Finally, I showed that extreme fire atmospheric profiles are typically high in DCAPE and that, if precipitating moisture is available for these situations, the fire and atmosphere could produce strong downdrafts or downbursts.

These results suggest that released moisture is sufficiently important, or has the potential to be, to merit closer examination. The next logical steps would be performing field observations to determine how much water is in the air in the plume of a large fire, and to examine the impact of added moisture on numerical simulations of fire convection. The former is the only way to prove that this moisture is indeed present, and is the most important step if research is to proceed in this area.

In a general sense, a fire provides energy and organization to maintain a coherent updraft structure, where ambient conditions may have produced only small updrafts that entrained large amounts of air. If the fire organizes the updraft so that a core of plume air maintains its properties and can rise to its equilibrium level, this alone can subsequently influence fire behavior. The factors working against the organization are turbulence, strong winds and wind shear (a source of turbulence). Any of these can dilute the air in the plume and prevent it from reaching its full potential. How do wind and wind shear affect the properties of the air in the plume? Is there a clear point where the wind tilts the plume so much that any return circulation affects the fire behavior, and how does moisture affect that circulation? Fire researchers already recognize these as important questions but, as we seek answers, we should also consider released moisture.

The traditional definition of fire behavior describes the controlling factors as fuels, atmosphere, and topography. If released moisture is indeed an important factor controlling fire behavior, then it presents an area of fire behavior research that requires strong knowledge and understanding of both fuel conditions and the atmospheric conditions. The link between these two becomes a strong two-way interaction that cannot

be studied or understood in separate fuel and atmospheric pieces.

There are also implications of this work for management, though practical application is far down the road. If a manager knew that a certain rate of moisture release was a threshold for extreme fire behavior on a given fire and day, the manager may attempt to control rate of spread during a specific time period in the hope that the moisture release rate would stay below the threshold, thus preventing possible erratic behavior. Fuel managers could also begin considering fuel loads that would hold the possible released moisture down below a climatologically determined level that divided blow-up from well behaved fire probabilities.

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Planned Burn-Piedmont. A local operational numerical meteorological model for tracking smoke on the ground at night: model development and sensitivity tests

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Abstract. Smoke from both prescribed fires and wildfires can, under certain meteorological conditions, become entrapped within shallow layers of air near the ground at night and get carried to unexpected destinations as a combination of weather systems push air through interlocking ridge–valley terrain typical of the Piedmont of the Southern United States. Entrapped smoke confined within valleys is often slow to disperse. When moist conditions are present, hygroscopic particles within smoke may initiate or augment fog formation. With or without fog, smoke transported across roadways can create visibility hazards. Planned Burn (PB)-Piedmont is a fine scale, time-dependent, smoke tracking model designed to run on a PC computer as an easy-to-use aid for land managers. PB-Piedmont gives high-resolution in space and time predictions of smoke movement within shallow layers at the ground over terrain typical of that of the Piedmont. PB-Piedmont applies only for weather conditions when smoke entrapment is most likely to occur—at night during clear skies and light winds. This paper presents the model description and gives examples of model performance in comparison with observations of entrapped smoke collected during two nights of a field project. The results show that PB-Piedmont is capable of describing the movement of whole smoke plumes within the constraints for which the model was designed.

Additional keywords: drainage winds; nocturnal smoke; smoke entrapment; visibility.

Introduction

Slope winds are increasingly recognized as important in the local transport of air pollution and airborne biota. One class of slope wind, the drainage wind, occurs frequently at night when synoptic scale weather associated with high and low pressure systems that produce large-scale wind are relatively weak. At some time during the day (usually near sunset), long-wave radiation from sloping land surfaces exceeds incoming short-wave solar radiation. The land surface and a shallow layer of air just above the surface cool. Minuscule pressure forces combined with negative buoyancy drive cooled air downslope toward valleys. The cool air accumulates in the valleys and either ponds or flows down-valley depending on the slope of the valley floor.

Much research on drainage winds has been conducted in mountainous areas of the world—Australia, Europe, United States (USA), Canada and Mexico. For a list of international local wind studies, see Papadopoulos and Helmis (1999). Wind speeds usually range between 2 and 6 m s⁻¹ (Post and Neff 1986; Clements *et al.* 1989; King 1989; Neff and King 1989; Sakiyama 1990). Drainage layer depths grow from the valley floor to 200–400 m, usually the depth of

steep-sided valleys (Neff and King 1989; Sakiyama 1990). Top-to-bottom temperature differences within these drainage flows are usually 4–5 K (Post and Neff 1986; Neff and King 1989; Sakiyama 1990) giving inversion strengths of 0.01–0.02 K m⁻¹, although stronger inversions have been recorded (McKee and O'Neal 1989).

Drainage flows also occur in the shallower stream valleys of the Piedmont (Garrett and Smith 1984). Winstead and Young (2000) observed 48 drainage flows exiting from valleys in Maryland onto Chesapeake Bay. Piedmont drainage flows are typically shallow and weak. They form predominately when larger scale circulations are weak (light winds) and when radiational cooling maximizes pressure forces and negative buoyancies within drainage airmasses (clear skies). These concurrent events occur within anticyclones (Lavdas and Achtemeier 1995).

The Piedmont is a low plateau roughly 100–300 km wide and 1500 km long extending down the south-east USA from Maryland to Alabama. It separates the Atlantic Coastal Plain and Gulf Coastal Plain from the Appalachian Mountains. The Piedmont rises from ~200 m along the coast-facing side to ~400 m on the mountain-facing side. The Piedmont is highly

eroded into a network of interlocking ridges and valleys. Ridge/valley elevation differences typically range from 20 to 50 m.

Piedmont drainage flows are subject to disruption by larger-scale wind systems (Achtemeier 1993). Synoptic scale pressure fields can exceed weak pressure fields within drainage flows. When this happens, drainage flows can be reversed to blow up-drainage and smoke and other pollutants transported into side valleys and/or into adjacent valleys through gaps in ridges. Transport of smoke into adjacent valleys presents a challenge for numerical modeling if the goal is an operational model. It is no longer practicable to define a model grid to enclose a single valley or even the valley and its tributaries (O'Steen 2000), or to compress a grid domain for the valley of choice (Varvayanni *et al.* 1997). The model domain must be extended several tens of kilometers to enclose adjacent drainage systems with the same resolution as that for the target valley.

Knowledge of air movement near the ground at night is important to land managers who use prescribed fire in the management of forest and agricultural lands. A prescribed fire is planned and set subject to fire safety and air quality regulations to accomplish a predefined management objective. Although the vast majority of prescribed burns are completed without incident, there are occasions when smoke from smoldering fuels persists after sunset and becomes entrapped within slow-moving drainage flows. Entrapped smoke can drift into populated areas and impact residents, particularly those with respiratory problems. Smoke-laden air masses can drift across roadways and contribute to poor visibility. Smoke and associated fog have been implicated in multiple-car pile-ups that have caused numerous physical injuries, heavy property damage and fatalities (Mobley 1989).

Planned Burn (PB)-Piedmont is an operational numerical weather and smoke model designed to simulate near-ground smoke transport at night. The sections to follow present brief descriptions of the constraints placed on an operational numerical drainage flow/smoke model, model assumptions, model equations, a method to solve the model equations, and examples of smoke transport simulations over the complex interlocking terrain of the Piedmont of the South.

PB-Piedmont: assumptions and model theory

PB-Piedmont is designed to predict the transport of smoke trapped near the ground at night over complex interlocking ridge/valley systems typical of the Piedmont. The operational requirements are that PB-Piedmont must be installed on desktop computers operated by land managers. The model must be capable of being run by those with little computer experience and no experience in meteorological modeling. The model must generate timely predictions if it is to be a factor in decision-making.

PB-Piedmont must be able to model smoke on the terrain scales through which local circulations transport smoke. The

grid scale must be fine enough to resolve shallow gaps in ridges. The grid must be large enough to enclose adjacent drainages when necessary. Therefore, to satisfy the spatial requirements and to produce timely predictions, the number of calculations must be minimized. It is required that the mathematics be simple and the physical terms describing complex processes be simplified or replaced with empirical terms.

Basic model assumptions

The model design depends on several assumptions regarding the meteorology of the Piedmont. First, Achtemeier (1993) found that Piedmont drainage flows were typically 15 m deep and generated down-valley winds as slow as 0.2 m s^{-1} . Shallow drainage layers can be 'vertically integrated', meaning that air movements within the layer are approximated by the mean wind for the layer. The vertically integrated approach has been used successfully by Garrett and Smith (1984) and O'Steen (2000). Entrapped smoke is a passive tracer and is assumed to be uniformly distributed through the depth of the drainage layer. The model therefore is not appropriate for simulating nocturnal smoke movement in the mountains because drainage flows there may be much deeper and internal circulations may become significant in transporting smoke.

Second, it is assumed that the meteorology of the Piedmont during clear sky and light wind episodes (prevailing weather when most smoke entrapment occurs) consists of two scales of motion—the synoptic scale pattern of low- and high-pressure centers and fronts, which can be described by the National Weather Service network of surface weather stations, and the drainage scale (meso-beta), which is characterized by the terrain of the Piedmont. Weather disturbances that are too small to be resolved by the existing network of surface stations (meso-alpha) are assumed to be negligible or non-existent. The model therefore is not appropriate for simulating nocturnal smoke movement near the coasts or large lakes where there are land breezes.

Third, because the domain size is limited, it is assumed that regional scale temperatures, winds and pressure data needed by the model can be represented by single values interpolated from surrounding weather stations to the location of the prescribed burn.

Fourth, it is assumed that the drainage layer airmass is effectively 'decoupled' from overlying airmasses. Near sunset, under conditions of light winds and clear skies, the ground and the air immediately above it cool rapidly. Mixing with the overlying airmasses ceases as the stable drainage airmass deepens. Decoupling makes it possible to model just the meteorology of the drainage layer.

Governing equations

Figure 1 shows the schematic structure of the PB-Piedmont model. The drainage layer is defined as a sublayer within

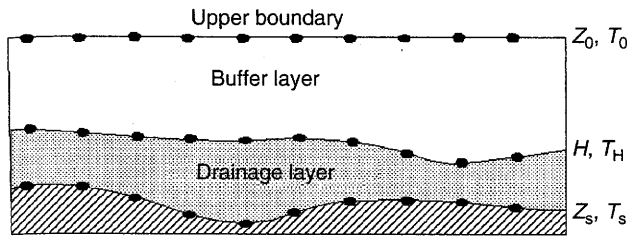


Fig. 1. Vertical structure of the PB-Piedmont numerical wind model. Dots represent grid points.

the nocturnal boundary layer that initially overlays the whole domain—ridge tops and valley bottoms. This is done to satisfy constraints of spatial continuity throughout the model domain. The drainage layer extends from the surface, Z_s , to a height, H , defined as the top of the drainage layer. Above it a buffer layer (also within the nocturnal boundary layer) extends from H to Z_0 , a level of constant height 100 m above the highest ground as defined from the elevation data within the model domain. It is assumed that Z_0 is a level of constant pressure. This assumption is not severe when PB-Piedmont is run under weak anticyclonic conditions. The assumption of a vertically integrated drainage layer constrains the definition of the vertical coordinate for PB-Piedmont as one bounded by H – Z_s . The lower coordinate surface follows the terrain, Z_s . The upper coordinate follows the top of the drainage layer, H . The vertical coordinate is defined by,

$$S = (H - z)/(H - Z_s). \quad (1)$$

The Lagrangian difference forms for the drainage layer prognostic equation for the layer mean wind components, u and v , per unit mass, are,

$$\begin{aligned} \frac{\Delta u}{\Delta t} = & - \left\{ R\bar{T} \frac{\Delta \ln(p)}{\Delta x} + g \left[(1 - S) \frac{\Delta H}{\Delta x} + S \frac{\Delta Z_s}{\Delta x} \right] \right\} \\ & + f_v + m_x(U - u) - \frac{1}{(H - Z_s)} C_{Du} \sqrt{u^2 + v^2} \\ \frac{\Delta v}{\Delta t} = & - \left\{ R\bar{T} \frac{\Delta \ln(p)}{\Delta y} + g \left[(1 - S) \frac{\Delta H}{\Delta y} + S \frac{\Delta Z_s}{\Delta y} \right] \right\} \\ & - f_u + m_y(V - v) - \frac{1}{(H - Z_s)} C_{Dv} \sqrt{u^2 + v^2}, \end{aligned} \quad (2)$$

where R is the gas constant for dry air, $\bar{T}$ is the mean temperature over the difference interval, and g is the acceleration of gravity. See Table 1 for a list of definitions of the variables that appear in the mathematical equations.

The forcing functions are described as follows:

- (1) The pressure gradient terms of the hydrodynamic equations transformed into the 'S'-coordinate system are enclosed in braces. Over sloping surfaces, the pressure gradient in the sigma coordinate system is expressed by

Table 1. Definitions of the variables that appear in the mathematical derivations

Variable	Definition
C_D	Drag coefficient
c_r	Coefficient of scatter for the smoke particle generator
C_T	Drainage layer cooling rate ($2.78 \times 10^{-3} \text{ K s}^{-1}$)
D_r	Drainage layer replenishment rate ($5 \times 10^{-4} \text{ m s}^{-1}$)
f	Coriolis parameter
g	Acceleration of gravity (9.8 m s^{-2})
h	Depth of the drainage layer (m)
H	Height of the top of the drainage layer (m)
k	Shape factor for the Gaussian weight function (m^2)
m_x, m_y	x, y components of the mixing coefficient (5×10^{-4})
p	Pressure
R	Universal gas constant ($287.04 \text{ m}^2 \text{ s}^{-2} \text{ K}^{-1}$)
R_e	Frequency response for the A-Function
s	Coordinate oriented along valley-axis
S	Hybrid terrain—and drainage layer height—following vertical coordinate system
t	Time (s)
T	Temperature (K)
T_H	Temperature (K) at the top of the drainage layer
T_0	Temperature (K) at the top of the model domain
T_R	Reference synoptic temperature (K) taken 3 h before sunset and interpolated to the fire location from National Weather Service hourly weather data
T_S	Synoptic scale temperature (K) interpolated to the fire location from National Weather Service hourly weather data
$\bar{T}$	Average temperature as specified
u, v	x, y -component of the model drainage layer horizontal mean wind (m s^{-1})
U, V	x, y -component of the observed surface wind interpolated to the burn site from surrounding weather stations
$U_{\max}$	Maximum wind speed found in the model domain at a given time
u_g	Geostrophic wind
x	x (east direction)
y	y (north direction)
z	Up as defined in the modified sigma coordinate
Z_s	Ground surface elevation (m)
Z_0	Height (m) at the top of the model domain
δ_1, δ_2	Weights for the cooling function
γ	Drainage layer lapse rate of temperature (K m^{-1})
Γ_d	Dry adiabatic lapse rate of temperature (0.01 K m^{-1})
λ	Wavelength (m)

two terms, in this formulation, the $\ln(p)$ term and the bracketed term multiplied by gravity. These terms can be large and are of opposite sign. Thus, the pressure gradient over sloping surfaces can be the small difference between large values. Achtemeier (1991) showed that the taking of differences of pressure over sloping coordinate surfaces can generate hydrostatic truncation errors of magnitude sufficient to overwhelm the minuscule pressure forces that drive Piedmont drainage flows. The problem can be eliminated if the pressure at any grid point is calculated downward from the top of the model domain by stepping down in thin layers of thickness no greater than 10 m.

- (2) The Coriolis terms (second terms of equation 2) represent the impact of the earth's rotation on the existing wind. The Coriolis force turns the wind to the right of its existing motion in the northern hemisphere. The parameter f is a function of the latitude and the rotation rate of the earth. The terms are small but they can impact the movement of slow drainage flows over flat basins.
- (3) The third terms couple/decouple the modeled drainage layer with the synoptic scale wind (U , V) supplied as an upper boundary condition. For the heated part of the day the bulk mixing coefficients (m_x , m_y) are set to $5 \times 10^{-4} \text{ s}^{-1}$ to drive the solution to the synoptic scale wind. Then, using the synoptic scale temperature for 3 h before sunset as a reference temperature, T_R , the coefficients are allowed to decrease to 0.0 when the temperature has cooled 2 K from T_R . This step simulates the nocturnal transition period just before sunset (Achtemeier 1993). The exact functional form for the coefficients is not critical to PB-Piedmont as the important drainage layer forcing begins after sunset.
- (4) The friction terms (last terms of equation 2) are weighted inversely by the depth of the drainage layer. The deeper the drainage layer, the less is the frictional drag on the vertically integrated airmass. The bulk drag coefficient for a southern pine forest is $C_D = 10^{-2}$ (Garrett and Smith 1984). The Piedmont land cover is typically southern pine forest interspersed with open fields—clearings for dwellings, pasture lands, thinned plots or cut-over areas. Thus part of the Piedmont land cover could be represented as grassland with $C_D = 10^{-3}$. Sensitivity tests revealed that drainage winds generated by the terrain of the Piedmont overwhelm forcing generated by differences in land use. Land use forcing is constrained to modulating the rate of down-valley flow and causing minor within-valley wind deflection and channeling. Thus, it is not unreasonable to set $C_D = 10^{-2}$ for the whole domain.

The Lagrangian difference forms for the drainage layer prognostic equations for height and temperature are:

$$\begin{aligned} \frac{\Delta H}{\Delta t} &= \frac{\Delta Z_s}{\Delta t} + (H - Z_s) \left(\frac{\Delta u}{\Delta x} + \frac{\Delta v}{\Delta y} \right) + D_r \\ \frac{\Delta T}{\Delta t} &= \frac{\Delta H}{\Delta t} \Gamma_d + \frac{\Delta T_s}{\Delta t_h} + \delta_1 \delta_2 C_T. \end{aligned} \quad (3)$$

Drainage layer height is adjusted for air movement over unlevel terrain (first term) and deepening within a divergent wind field (second term). The third term (D_r) represents the drainage layer replenishment rate during the course of the night. The drainage layer (not to be confused with the nocturnal boundary layer: Achtemeier 1993) is assumed not to exceed two-thirds the depth of Piedmont valleys (approximately 50 m). Thus the drainage layer is allowed to

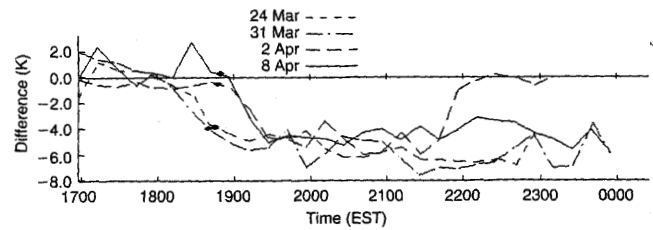


Fig. 2. Time series of the difference in temperature between a ridge and a basin site for the following dates in 1992: 24 March, 31 March, 2 April and 8 April. The elevation difference is 30 m.

grow 24 m over a typical 12 h nocturnal period in addition to the initial 10 m depth of the layer. Therefore, cooled air that drains from the ridges into the valleys is constantly replenished at a rate of 2 m h^{-1} ($5 \times 10^{-4} \text{ m s}^{-1}$). Furthermore, in the absence of confirming data, the drainage layer replenishment rate is assumed to be constant in time and uniform in space.

Temperature is adjusted as follows. The first term is an adiabatic adjustment for changes in the height of the drainage layer. These include adjustments for movement over changing elevation, deepening caused by a convergent wind field, and drainage layer replenishment. The second term represents synoptic scale cooling (heating) defined as the difference between two consecutive hourly observations of temperature (T_s) interpolated to the center of the model grid from surrounding National Weather Service surface weather stations and thence converted into cooling (heating) rate in K/s by dividing by 1 h (Δt_h).

The third term represents cooling within the drainage layer after sunset. The local cooling rate is generated within PB-Piedmont and applies to the temperature at the ground. It was derived from observations of cooling within a basin during clear skies at night (Achtemeier 1993) and confirmed by other observations of local inversion formation in the Piedmont areas of the South. Figure 2 shows a time series of the temperature difference between a valley site and a ridge-top site (elevation difference of 30 m) on four nights during March–April 1992. Negative differences occur when the basin is colder than the ridge. Figure 2 shows that, on all four nights, temperature differences were approximately zero before sunset (identified by the shaded circles between 1845 and 1900 EST). Differences dropped to approximately -5 K within 0.5 h of sunset as the valley cooled much more rapidly than did the ridge. Thereafter cooling in the valley was approximately the same as cooling on the ridge. Temperature differences averaged approximately -5 K after 1930 EST. An exception was the night of 2 April when gusty winds after 2145 EST mixed out the valley, a frequent occurrence for shallow drainage flows (Gudiksen *et al.* 1992).

Based on Fig. 2, the maximum within-drainage-layer cooling rate is $C_T = 2.78 \times 10^{-3} \text{ K s}^{-1}$. However, the actual cooling is determined by the coefficients that take on the

following values:

$$\delta_1 = \begin{cases} 0 & \text{if before sunset} \\ 1 & \text{if after sundown} \end{cases}$$

$$\delta_2 = \begin{cases} 0 & \text{if } \gamma \geq \gamma_R \\ 1 & \text{if } \gamma < \gamma_R \end{cases},$$

where

$$\gamma = \frac{(T_H - T)}{H - Z_s}.$$

The first coefficient allows for cooling within the drainage layer only after sunset. The second coefficient is designed to prevent the development of excessive lapse rates of temperature within shallow layers. For each time step, the drainage layer lapse rate of temperature, γ , is calculated as the difference between the temperature at H (T_H) and the surface temperature, divided by the depth of the drainage layer. Based on Fig. 2, under dry conditions, cooling is stopped when γ exceeds a reference lapse rate, $\gamma_R = 0.1667 \text{ K m}^{-1}$.

Additional corrections are made for moist conditions. As the air temperature falls to the dew point temperature, dew formation commences beginning first in the cooler basins. Latent heat release slows the rate of cooling but the removal of water dries the air thus lowering the dew point temperature slightly. PB-Piedmont allows the dew point temperature in valleys to fall no more than 1.75 K below the dew point temperature interpolated to the center of the model grid from surrounding National Weather Service surface weather stations.

The use of empirical cooling functions is preferred over complicated time-dependent cooling equations for three reasons. First, simple empirical relations require little computational power and seem justified with respect to other simplifying assumptions in the design of PB-Piedmont such as the vertically integrated drainage layer. Second, land surface conditions such as soil moisture, ground cover and soil type will not be available to most users of PB-Piedmont. Third, average cooling rates should apply to a first approximation for weather conditions of clear skies and light winds for which PB-Piedmont is applicable.

Integration methods for PB-Piedmont

PB-Piedmont uses a downstream-stepping semi-Lagrangian scheme—a departure from most semi-Lagrangian methods, which are upstream-stepping schemes. Staniforth and Cote (1991) provide a general review of upstream-stepping schemes. The downstream-stepping approach is used because terrain is not smoothed in PB-Piedmont. For highly divergent drainage flows over the complex terrain of the Piedmont, it may not be possible to accurately define back-trajectories needed for upstream-stepping methods. The time step, valid for the whole domain, is calculated internally subject to the Courant, Friedrichs, Levy (CFL) linear stability criteria ($U_{\max} \Delta t / \Delta x < 1$) (Thompson 1952). The time step is defined

by half the grid spacing divided by the maximum velocity found in the domain. Velocity component changes at each grid point are found by multiplying equation (2) by the time step. Then each point is moved forward a distance equal to the product of the total velocity (existing velocity plus velocity change) with the time step through the predictor–corrector method developed for streamlines by Achtemeier (1979) and expanded for trajectories (Scott and Achtemeier 1987). The state variables, T and H , are carried to the new locations along with the velocity. Temperatures are adjusted adiabatically for changes in elevation.

The interpolation step: the A-Function

Upon completion of the Lagrangian downstream step, the new field of points is no longer located at grid square intersections but is quasi-regularly distributed over the grid. The next step is to recalculate the properties of the field at the grid square intersections via interpolation of properties from the surrounding field of points. Various interpolation methods have been used for upstream-stepping semi-Lagrangian methods; the most widely used being cubic spline interpolation (Staniforth and Cote 1991). Ritchie (1986) devised a scheme to eliminate interpolation and the associated smoothing. However, these schemes are specific to the upstream-stepping method.

An objective analysis or surface fitting scheme is appropriate for gridding quasi-regularly distributed data. The behavior of the Gaussian function as a weight function for successive corrections interpolation of meteorological data has been studied by Barnes (1964, 1973, 1994), Achtemeier (1987, 1989, 1994), Pauley and Wu (1990) and others. Barnes (1964) produced a theory of how the Gaussian interpolation scheme smoothes data as a function of wavelength and reinterpolation. Knowledge of the response characteristics for an interpolation scheme is useful when deriving a local mass balance restoration equation.

Successive corrections interpolation requires:

- (1) An estimation of the value of a function at points on a grid by interpolation from surrounding data points;
- (2) An estimation of the value of the function at the data points by reinterpolation from the surrounding grid points; and
- (3) An improvement in the estimation of the function at the grid points by adding an additional interpolation of the discrepancies found in step (2) to the original estimation.

Conceptually, steps (2) and (3) can be repeated indefinitely at great computational cost to cause the estimation to converge toward, but not always to, the original function. The data generated by the Lagrangian step in PB-Piedmont are quasi-regularly distributed, are error-free, and correctly resolve the motion scales carried in the model. Thus interpolation smoothing should be minimized or eliminated, if possible.

Table 2. A-Function theoretical response and measured response for analysis of equally spaced data for wavelengths ranging from $48\Delta x$ to $2\Delta x$, with Δx being the separation between grid points $4k = 1.40; 0.70$

Wavelength	Theoretical response	Measured response
48	1.0000	1.0002
24	0.9999	1.0006
16	0.9993	1.0009
12	0.9978	1.0006
8	0.9896	0.9964
6	0.9697	0.9829
4	0.8776	0.9154
3	0.6143	0.7424
2	0.3242	0.6474

In the derivation found in Appendix 1, the restorative properties of the Gaussian function are investigated and the results used to derive an algorithm—the A-Function—that reproduces the accuracy of the successive corrections methods but without the need for steps (2) and (3). Achtemeier (1989) found that the Gaussian weight applied to data distributed on a regular mesh restores more of the short wavelengths than predicted by theory. Table 2 compares the theoretical responses calculated from equation (A8) with the measured response (defined as the ratio between the amplitude restored by the A-Function applied to regularly spaced data and the amplitude of the original function) for the range of wavelengths from $48\Delta x$ to $2\Delta x$. Here Δx is the spacing between grid points.

The overall impact of the departure from theory is to increase the robustness of the A-Function. The A-Function amplifies all wavelengths, with the greatest amplification being for the shorter waves. With the choices for $4k$ in Table 2, the measured response is slightly greater than 1.0 for the longer wavelengths, meaning that the A-Function produces a slight amplification of these waves. Thus the wind components, temperatures and depths of drainage flows are not smoothed for valleys for which ridge-to-ridge distances are defined by eight or more grid points. Shorter wave features are smoothed. For example, the A-function smooths $\sim 9\%$ of drainage flows within ravines defined by the $4\Delta x$ wave, and $2\Delta x$ features are smoothed by 36% per time step.

Drainage layer mass adjustment

For each time step, the field being interpolated is smoothed non-uniformly through application of the A-Function. In PB-Piedmont, interpolation is done on u , v , T and the thickness of the drainage layer, $h = H - Z_s$. Table 2 shows that amplitudes of wavelengths smaller than $8\Delta x$ suffer smoothing. Thus drainage flows within small-scale features such as deep ravines will become weaker, warmer and shallower. Drainage flows within larger valleys and basins will be largely unaffected by the A-Function.

PB-Piedmont replaces some of the mass lost by interpolation through a mass redistribution function. The integrated or total mass loss over the whole model domain can be found by taking the difference between the total mass at successive time steps and correcting for the mass lost or gained through the lateral boundaries of the domain and drainage layer replenishment. The model equations are normalized by unit mass. Therefore the mass change is represented by the change in the volume of the drainage layer domain or, in reference to the S-coordinate system, the volume of the layer contained below H . The volumes of the drainage layer domain at the current time step and at the previous time step, corrected for drainage layer replenishment, D_r , are given by

$$V^t = (\Delta x)^2 \sum_{i=1}^m \sum_{j=1}^n h_{i,j} \quad (4)$$

$$V^{t-1} = (\Delta x)^2 \sum_{i=1}^m \sum_{j=1}^n (h_{i,j}^{t-1} + D_r).$$

The volume of the drainage layer passed through the boundaries of the model domain is

$$V_b = \Delta x \Delta t \left[\sum_{i=1}^m (h_{i,1} v_{i,1} - h_{i,n} v_{i,n}) + \sum_{j=1}^n (h_{1,j} u_{1,j} - h_{m,j} u_{m,j}) \right]. \quad (5)$$

The volume replacement to be added to the total volume so that total volume is conserved is then

$$V_d = (V^{t-1} + V_b) - V^t. \quad (6)$$

However, because the horizontal dimensions of the domain are fixed, the volume replacement is equivalent to adjusting the layer depth for the layer depth discrepancy. The adjusted layer depth is

$$h_{i,j}^{\text{adj}} = h_{i,j} + V_d \left(\frac{h_{i,j}}{V^t} \right). \quad (7)$$

Total drainage layer volume (mass) is conserved by equation (7); however, the replacement of mass is not uniform. The greatest mass replacement occurs where the drainage layer is deepest, namely in the valleys. Therefore, for the interpolation/mass correction cycle of a given time step, long wavelength ($> 8\Delta x$) valleys, which were unsmoothed by interpolation, receive mass through equation (7) and thus gain mass. Short wavelength valleys receive mass through equation (7) but not at sufficient magnitude to compensate for mass loss through interpolation.

Initial and boundary conditions and synoptic time stepping for PB-Piedmont

In keeping with the assumptions described at the beginning of this section, the vertical structure of the model atmosphere

is in two layers, the shallow drainage layer near the ground and a buffer that extends from it to an elevation of 100 m above the highest terrain. Figure 1 summarizes the layer division along with model details. Initial temperature, dew point temperature and wind components are interpolated from surrounding National Weather Service surface weather stations to the location of the burn site by using a Gaussian scheme developed by Barnes (1964, 1973) and optimized by Achtemeier (1987, 1989) for objective analysis of quasi-regularly spaced data. Horizontal pressure gradients are interpolated with a variant of the direct derivative method developed by Caracena (1987).

These interpolated synoptic scale weather data are used to develop the initial upper boundary as follows. The procedure should be started at least 3 h before sunset to assign the reference temperature for the mixing coefficients in equation (2). The lapse rate of temperature for the buffer layer can vary with time but is constant in space. The initial buffer layer lapse rate is dry adiabatic. The initial drainage layer depth (H) is 10 m and the initial drainage layer lapse rate of temperature is also dry adiabatic. The temperature (T_0) at the top of the model domain (Z_0) is calculated by adjusting the synoptic temperature dry adiabatically for the difference in elevation between the elevation of the burn site and the top of the model domain. Then the temperatures at H and Z_s at each grid point are calculated adiabatically from Z_0 to the top of the drainage layer and the ground.

The wind and pressure fields for the drainage layer are initialized with the synoptic wind components and horizontal pressure gradients.

After the initial conditions are specified, PB-Piedmont is run out to 30 min past the hour. The model pauses for new data to update the synoptic weather for the next hour. Then the model is run from 30 min before the hour to 30 min past the hour. The above procedure is repeated for each hour so that the model runs with centered time differences for synoptic weather.

In its current design, PB-Piedmont is predictive but must wait for the availability of key boundary data. When numerical model prediction data become routinely available for forestry users, PB-Piedmont will be made fully predictive.

Elevation data

Critical for accurate modeling of the smoke movement near the ground at night is specification of the lower boundary. Smoke can be entrapped in small basins, flow along road and stream cuts or through small gaps in ridges. Therefore, the model requires very fine scale spatial resolution and accurate elevation data in order to model smoke on the terrain scales through which local circulations transport smoke. The US Geological Survey's 30 m digital elevation model (DEM) data provide the lower boundary for PB-Piedmont. The horizontal grid spacing is set by the user within the range 30 m (the minimum spacing of USGS DEM elevation data) to 150 m (larger

spacing degrades ground features in terrain characteristic of the Piedmont).

Smoke particle generator

Residual smoke is assumed to occur anywhere within an area burned. The area burned, supplied by the user, is converted to a square centered at the coordinates given for the burn. A 5×5 grid is laid over the square and 25 particles per time step are released at the grid intersections. These particles serve as tracers carried by the local winds. The 25-point grid requires that all parts of the burn area contain smoke sources.

Particle locations are moved with the local winds generated within PB-Piedmont. The new locations are found through the predictor/corrector method for generating streamlines (Achtemeier 1979). However, because all particle transport is by mean flow within a vertically integrated domain, the particles will eventually converge to a line at the centers of valleys. To account for re-circulations and small-scale turbulence within the drainage layer wind field, the particle locations are subjected to additional displacements through

$$\begin{aligned}x_{\text{new}} &= x + c_r(0.5 - x_r) \\y_{\text{new}} &= y + c_r(0.5 - y_r),\end{aligned}\quad (8)$$

where $c_r = 0.25$ and $0 < x_r, y_r < 1$ are random numbers. This choice for coefficients for equation (8) allows maximum horizontal displacements of up to 12% of a grid separation per time step and these are independent of the mean air flow within drainage layers. However, given that the displacements can be of either sign, cumulative displacements are relatively small. For a large number of particles in a uniform wind field, equation (8) yields a Gaussian distribution.

PB-Piedmont does not calculate smoke concentrations. The amount of material engaged in smoldering may vary greatly within a burn area and among burn sites. Thus the smoldering emission rates for a particular burn are poorly known. Therefore the primary contribution of PB-Piedmont to the smoke management problem is in providing an answer to the question: 'Where does the smoke go?'

PB-Piedmont validation through aerial photography and surface temperature observations

Tests with PB-Piedmont show that the combination of synoptic scale wind systems with weak drainage winds that form over terrain typical of the Piedmont of the south-east can distribute smoke in complex plumes and patches. Valley orientation with respect to the winds and with respect to adjacent valleys, steepness of slopes, locations of gaps in ridges, and valley depth can make the difference between entrapment and ventilation. Modeled smoke can be transported among valleys through small gaps in ridges. Whether these complex smoke plume patterns exist is a subject for model validation.

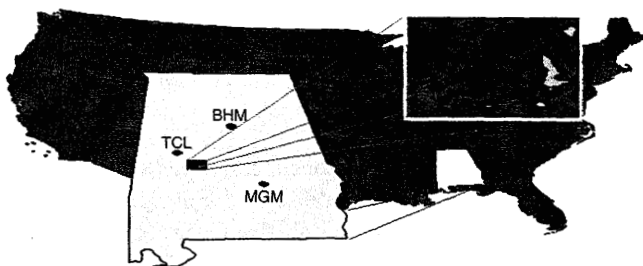


Fig. 3. Location of the site of experimental burns conducted at the Oakmulgee Wildlife Management Area in Alabama during March 1997 in relation to three nearby weather stations.

Existing observations on smoke movement at night are few. Data exist as locations of highway accidents where smoke was implicated as a causal factor and recorded reports of smoke by personnel who drive on highways surrounding burn sites searching for smoke incursions. Data on the movement of entire smoke plumes near the ground at night have been non-existent. Therefore, a key factor in the validation of PB-Piedmont is obtaining observational data on the movement of a whole smoke plume near the ground at night.

A video camera equipped with a light-enhancing device and an infrared cut-off filter was mounted in a Beechcraft King Air aircraft and flown over an experimental burn site at the Oakmulgee Wildlife Management Area located in the Talladega National Forest in western Alabama (Achtemeier *et al.* 1998) (Fig. 3). The site was selected for terrain typical of the Piedmont, safety and the absence of light sources. The site was located along a service road at the bottom of a stream basin that flowed to the north-east. Smoke was observed successfully and recorded during the evenings of 20 and 21 March 1997. Raw video images, methods of image analysis and resulting smoke distribution relative to surrounding landforms were described for 20 March 1997 by Achtemeier (1998) and Achtemeier *et al.* (1998). These results, along with results from 21 March 1997 are compared with simulations from PB-Piedmont in the following paragraphs.

20 March 1997

Modeling smoke movement for the night of 20 March 1997 was a difficult test for PB-Piedmont. Beginning at 2145 local standard time (LST), Forest Service ground personnel ignited 50 bales of hay soaked in diesel fuel along a road next to a stream basin that flowed to the north-east. Once the hay bales were flaming vigorously, the fire was extinguished with water. The bales then smoked profusely. Ground crews also detonated 60 smoke bombs that had burn lifetimes of ~2 min each. Aircraft overflights at ~1500 m altitude commenced at 2148 LST and continued at 7-min intervals for 2 h.

The project forecast called for winds to decrease to near-calm with rapid cooling in the basin to entrap smoke there. Drainage and valley flows favored slow movement of smoke down-valley to the north-east.

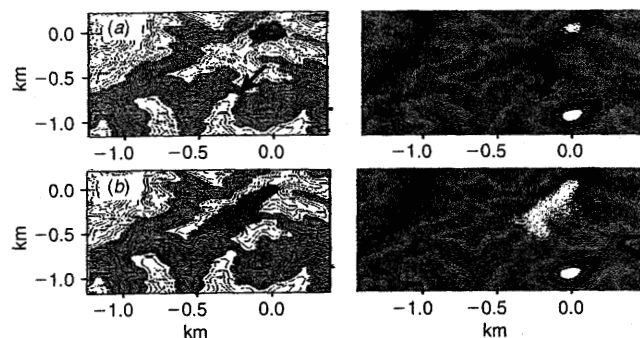


Fig. 4. Left panels: Image analysis of smoke plume on night of 20 March 1997 overlain on 30 m DEM elevation data. Right panels: PB-Piedmont simulation of smoke for the corresponding time—(a) 2150 LST and (b) 2215 LST. Arrows identify gap in ridge enclosing stream valley.

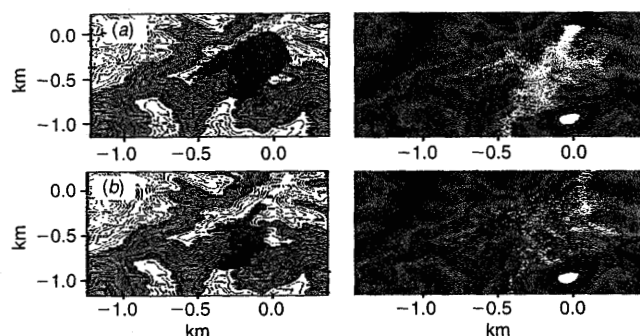


Fig. 5. Same as Fig. 4 except for (a) 2255 LST and (b) 2354 LST.

Figure 4 compares smoke movement relative to the surrounding elevation as extracted from video imagery with PB-Piedmont simulations for the first half of the experiment. All elevations are measured above mean sea level (msl). Elevations range from 100 m in the bottom lands to around 150 m along the ridge tops with a few high points near 170 m. Elevations greater than 130 m are shaded to better identify the drainage basin. Elevations above 135 m are shaded differently to highlight a 10 m deep gap in the ridge enclosing the southern end of the valley (arrows). Smoke generated at the burn site (Fig. 4a) did not move down-valley as expected. Instead, the smoke moved south-westward up-valley along the natural extension of the stream (Fig. 4b). Then the plume shifted to impinge directly upon a protruding ridge.

PB-Piedmont was initialized and updated with hourly observations of temperature, humidity, pressure and wind speed and direction as interpolated from the surrounding National Weather Service surface weather observing network. The right panels of Fig. 4 show smoke movement generated by PB-Piedmont for the corresponding times as observed in the left panels. Model smoke moves up-valley and divides around the protruding ridge.

During the last half of the experiment the smoke plume split around the ridge (Fig. 5a), flowed up a side valley and crossed the ridge through the shallow gap at the southern

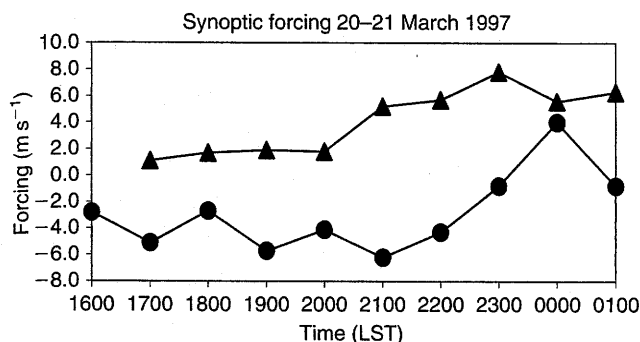


Fig. 6. Synoptic forcing for the 20 March 1997 experimental burn (circles) and for the 21 March 1997 experimental burn (triangles).

end of the valley. Smoke diversion through the side valley continued throughout the remainder of the burn (Fig. 5b). Model smoke also turns up the side valley and crosses the ridge through the gap at the southern end of the valley. PB-Piedmont results were nearly identical to the observed smoke movement with the exception that PB-Piedmont later allowed some smoke to drift down the valley. This reversal of the flow was partly caused when a key weather station ceased reporting hourly weather for the night. No smoke was observed down-valley from the burn site (Fig. 5b).

An explanation for the smoke movement observed on the evening of 20 March 1997 draws on the opposition between synoptic scale winds and drainage layer winds. As expected, drainage layer winds entrapped smoke within the stream valley. A weak low pressure center over the northern Gulf of Mexico turned the synoptic scale winds to blow from the north. These winds diminished after sunset and became light and variable during the course of the experimental burn.

However, the pressure forces that drove the synoptic scale winds did not diminish. A measure of the strength of the synoptic scale pressure forces is the 'geostrophic' wind calculated from the pressure gradients directed along the axis of a valley. The equation for synoptic forcing is

$$u_g = -\frac{RT}{f} \frac{\Delta \ln(p)}{\Delta s}, \quad (9)$$

where 's' is oriented parallel to the valley axis. Figure 6 shows the synoptic forcing for the night of 20 March 1997 (circles). Negative values identify when synoptic forcing favored winds blowing from the north or up-valley. Thus the synoptic forcing opposed the minuscule pressure forces that drove the drainage flows down-valley and dammed up the cooled airmasses at the south-western end of the valley. Synoptic forcing weakened after 2200 LST and became positive at midnight (when a key weather station stopped reporting) meaning that synoptic forcing favored winds blowing from the south or down-valley. Therefore, PB-Piedmont simulated smoke moving down-valley after 2300 LST.

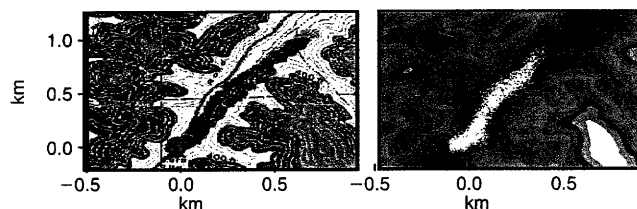


Fig. 7. Left panel: Image analysis of smoke plume at 2121 LST 21 March 1997 overlain on 30 m DEM elevation data. Right panel: PB-Piedmont simulation of smoke for the same time.

The deepened, smoke-filled drainage layer was easy to observe from the air. (An observer reported zero visibility in smoke and fog along the road leading up-valley from the burn.) However, once the smoke passed through the gap in the ridge at the southern end of the valley (Fig. 5) it was no longer visible from the air, presumably because the smoke layer became too shallow as the air accelerated downslope into the neighboring valley. A large tract of land located just south of the gap had been clear cut the previous year. Therefore, there were few if any large trees to obstruct smoke observation from the air.

21 March 1997

The procedures for conducting the experimental burn during the night of 21 March 1997 were nearly identical to those for 20 March 1997. The fire was started at the same location. Beginning at 2045 LST, Forest Service ground personnel ignited 50 bales of hay soaked in diesel fuel and detonated 60 smoke bombs. Aircraft overflights at approximately 1500 m altitude commenced at 2153 LST and continued at 7-min intervals for approximately 2 h.

The project forecast called for winds to decrease to near-calm with rapid cooling in the basin to entrap smoke there. Drainage and valley flows favored slow movement of smoke down-valley to the north-east. Figure 6 shows down-valley synoptic forcing for 21 March 1997 (triangles). Positive values mean that forcing was directed down-valley. Synoptic scale pressure forces reinforced drainage flows throughout the period of the experiment. Synoptic forcing was relatively weak (less than 2.0 m s^{-1}) until 2100 LST when forcing increased in response to an approaching weather system and remained above 5.0 m s^{-1} for the remainder of the experiment.

Figure 7 (left panel) shows the location of the smoke plume relative to the burn site and surrounding landforms at 2121 LST. The plume was defined as the locus of points taken from video imagery by image analysis methods described by Achtemeier (1998) and Achtemeier *et al.* (1998). Plume structure and orientation remained unchanged during the period of overflights. The smoke plume simulated by PB-Piedmont at 2121 LST is shown in the right panel of Fig. 7. Drainage winds confined the plume within the stream basin. The faint smoke plume was observable for a distance of $\sim 1 \text{ km}$. Smoke

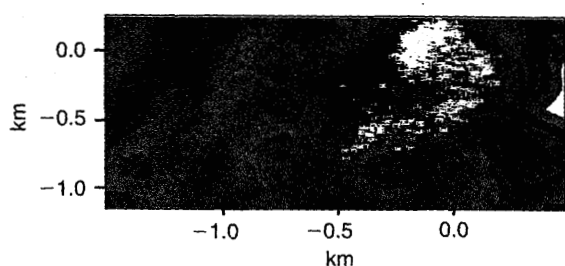


Fig. 8. Modeled smoke for 2255 LST 20 March 1997. Grid spacing is 150 m.

may have been partially obscured by overarching trees or the smoke layer thinned by spreading of the plume to the extent that smoke was no longer observable from the aircraft.

Internal validation

Critical to accurate simulations of smoke movement on the ground at night are the accuracies of underlying terrain, gradients of the atmospheric pressure at the surface used for calculating the synoptic scale forcing, and empirical cooling functions used for calculating drainage forcing. Model runs with bulk drag coefficient $C_D = 10^{-3}$ (grassland) in comparison with model runs using $C_D = 10^{-2}$ (southern pine forest) did not significantly alter either the location of the smoke plume or ridge-valley temperature lapse rates, which were critical for calculating drainage forcing.

Terrain forms the lower boundary of PB-Piedmont. If the terrain has been smoothed so that it no longer represents the terrain scales that channel smoke, then degraded performance of PB-Piedmont should be expected. Figure 8 shows model simulation of smoke for 2255 LST 20 March 1997 using a 150-m elevation grid. Figure 8 should be compared with the right panel of Fig. 5a. The 150-m analysis degraded the ridge located west of the valley containing entrapped smoke, effectively creating a shallower valley with slopes not as steep as those in the 30-m analysis. Smoke still was transported up-valley but the smoke was pushed to the east side of the valley and was late reaching the gap in the ridge. Additional runs with grid spacing in multiples of 30 m from 30 m to 150 m show that, for terrain characteristic of that of the Piedmont of the South, best performance is found for terrain at 30 m. Model performance degrades slowly through 150 m. For grid spacing greater than 150 m, terrain features that are important factors in the channeling of smoke may not be represented in the simulations.

Where smoke goes is determined by the sum of synoptic and drainage forcing. Synoptic scale forcing is calculated from horizontal gradients of atmospheric pressure at the surface. Horizontal pressure gradients are interpolated to the location of the burn from surface pressure data collected at surrounding weather stations. Several of these stations stopped reporting during the course of the night. Interpolated pressure data may therefore be degraded below the

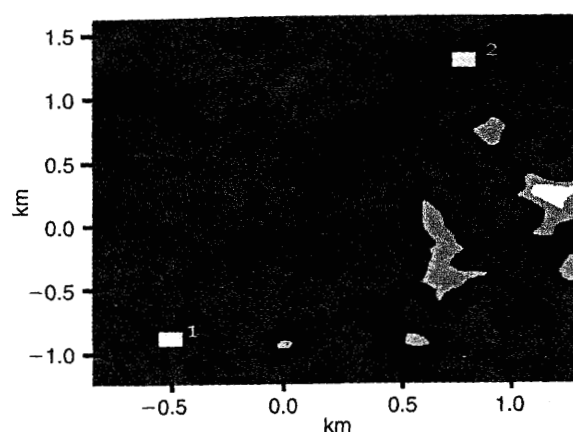


Fig. 9. Sites of two thermocouples for collecting temperature data on 20–21 March 1997 for validating PB-Piedmont. The red square identifies the burn site.

precision required for modeling smoke movement under light wind conditions. Down-valley synoptic forcing (Fig. 6) on 20 March 1997 (circles) averaged approximately -5.0 m s^{-1} for the period 1900–2200 LST. Forcing went to zero at 2300 LST and to $+4.0 \text{ m s}^{-1}$ at midnight after a key weather station suspended observations. These changes first left drainage forcing unopposed then reinforced down-valley forcing. PB-Piedmont moved smoke down-valley after 2300 LST (see top of right panel Fig. 5b). Neither did members of the burn crew nor personnel on board the aircraft observe smoke moving down-valley from the burn site. The inference from the continued transport of smoke up-valley after 2200 LST is that the synoptic scale forcing continued at approximately the same sign and magnitude for the next several hours.

As regards drainage forcing, cooling functions set forth in equation (3) determine the maximum rate of cooling and maximum lapse rate of temperature allowed for the drainage layer, $2.78 \times 10^{-3} \text{ K s}^{-1}$ and 0.1667 K m^{-1} respectively. However, the maximum lapse rate is not always attained. If the model lapse rate is, for example, less than the ambient lapse rate, then the model surface temperature is too warm relative to the temperature of the air just above the drainage layer. This means that the drainage flow is too weak and more likely to be dominated by synoptic forcing. Thus comparisons between the model-generated lapse rates and observed lapse rates provide an independent validation for PB-Piedmont regarding the strength of drainage flows.

Two temperature sensors were placed at different elevations along a road that ran from a ridge at the south end of the valley through the burn site (Fig. 9). The measurements were taken with thermocouples positioned $\sim 1 \text{ m}$ above ground. One sensor was located at the top of the ridge (146 m) south of the burn site and the other in the valley (91 m). The lapse rate of temperature was calculated as the difference between the observed temperature at the ridge top and the observed temperature in the valley divided by the difference in sensor

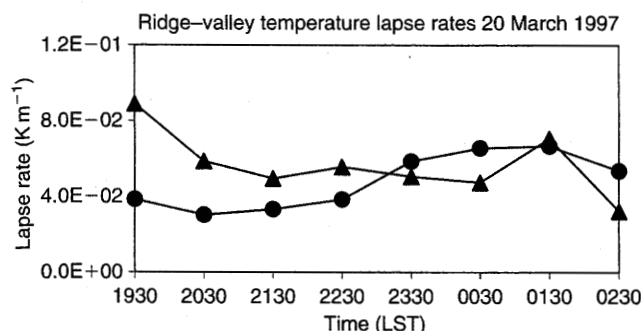


Fig. 10. Ridge-valley lapse rates of temperature (K m^{-1}) modeled (triangles) and observed (circles) for 20 March 1997. Lapse rates calculated from temperatures and elevations at sites 1 and 2 (see Fig. 9).

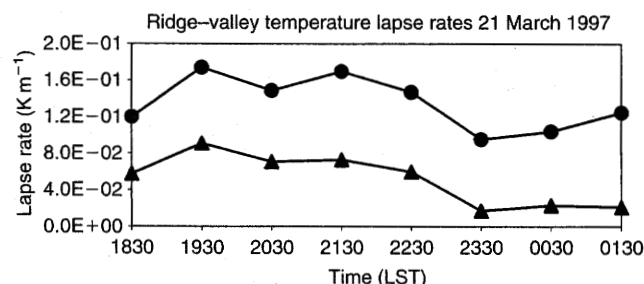


Fig. 11. Ridge-valley lapse rates of temperature (K m^{-1}) modeled (triangles) and observed (circles) for 21 March 1997. Lapse rates calculated from temperatures and elevations at sites 1 and 2 (see Fig. 9).

elevation. The PB-Piedmont lapse rate of temperature was taken as the difference between model-simulated surface temperatures at the ridge top and at the same location in the valley divided by the difference in elevation.

Figure 10 compares modeled with observed lapse rates of temperatures for 20 March 1997 for the period including the project overflights. All modeled and observed lapse rates were inversions. The model lapse rate of temperature at 1930 LST was approximately half of the maximum allowable yet was twice as great as the observed lapse rate of temperature. This means that the model was over-predicting the strength of the drainage forcing. The model lapse rate converged toward the observed lapse rate so that from 2130 LST on, correspondence was fairly good. Corrections for saturation were in effect from 2130 LST. Valley fogs were observed during the night of March 20.

Modeled and observed lapse rates of temperature were in excellent agreement for the first 3 h after sunset on 21 March (Fig. 11). Thereafter, model lapse rates of temperature decreased relative to the observed lapse rates of temperature. From Fig. 6, down-valley synoptic forcing increased beginning at 2100 LST. The impact of the larger synoptic forcing was to push the drainage air down-valley. Analysis of the drainage layer depth from 2100 LST to 2330 LST showed that the depth decreased by ~ 10 m. The drainage layer was being pushed down-valley faster than the drainage layer was

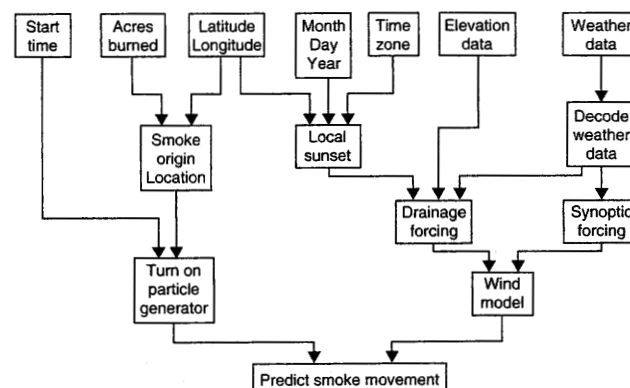


Fig. 12. Flow diagram showing inputs and key steps in PB-Piedmont for short-range predictions of smoke movement.

being replenished. Thus the model was under-predicting the depth of the drainage layer.

Comparisons with model-predicted smoke and observations of smoke in Figs 4, 5 and 7 suggest that the differences between the modeled and observed lapse rates of temperature in the drainage layers did not decrease the accuracy of the model predictions.

Discussion

PB-Piedmont is a time-dependent wind and smoke simulation model designed to aid the land manager in tracking the movement of residual smoke at night given that residual smoke is in fact present. Land managers experienced with tracking smoke on the ground at night know that wind reports from distant weather stations are often not representative of on-site air movement. PB-Piedmont provides critical numerical 'eyes' to 'see' where smoke is moving at night. Thus the model gives guidance on where smoke might be going.

Smoke observations and results from model simulations show that smoke movement at night over the Piedmont is a complex interaction between terrain and meteorology that cannot be easily anticipated by land managers. Ridge-valley elevation differences, small gaps in ridges and valley orientation with respect to other valleys and prevailing winds are factors in smoke transport. Synoptic forcing must reinforce drainage flows in some valleys and oppose drainage forcing in other valleys. Both synoptic forcing and drainage forcing are time dependent. Synoptic forcing changes as weather systems pass by during typical 10–12 h night-time periods. Drainage forcing depends on the depth of the drainage layer, cooling rates and moisture.

PB-Piedmont does not predict smoke concentrations. Emissions from residual smoke over a burn area are usually unknown. However, an advantage to this deficiency is that the inputs to the model are simple. User-defined inputs (Fig. 12) are limited to burn start time, acres burned, latitude and longitude of the burn site, the month, day, and year, and the time zone. The inputs are used to calculate local sunset,

locate the smoke origin grid, and turn on the particle generator. Elevation data are provided with the model. Weather data are available from select Internet sites.

This paper has presented the modeling philosophy for PB-Piedmont, the mathematical development of the model, and tests of the model using for validation aircraft video imaging of plumes from two experimental burns at night. These tests showed that the model performs well for drainage flows late in the evening.

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Appendix 1

The Gaussian function weights each datum inversely proportional to the square of its distance (r) from a grid point by

$$wt = \exp\left[\frac{-r^2}{4k}\right], \quad (A1)$$

where k is a shape factor for the weighting curve and therefore determines how smoothing will cause the final analysis to depart from the original data field. If $f(x, y)$ describes a field of data, then application of equation (A1) will generate a new function $g(x, y)$ such that,

$$g(x, y) = D_0(a, k)f(x, y), \quad (A2)$$

where $D_0(a, k)$ is a response function. Barnes (1964) showed that

$$D_0(a, k) = \exp(-a^2k), \quad (A3)$$

if the original data field is described by the function

$$f(x, y) = A \sin(ax),$$

where

$$a = 2\pi/\lambda \quad (A4)$$

and λ refers to wavelength. The optimal response for the 2-pass successive corrections Gaussian objective analysis is

$$B = D_0(2 - D_0) = 2D_0 - D_0^2 \quad (A5)$$

(Achtemeier 1987). To create the A-Function, add a constant to the original function so that

$$F(x, y) = C_0 + f(x, y). \quad (A6)$$

Then, using equation (A1), perform two separate single-pass interpolations with different k to yield two new functions:

$$\begin{aligned} G_1(x, y) &= D_1(a, k_1)F(x, y) \\ G_2(x, y) &= D_2(a, k_2)F(x, y). \end{aligned} \quad (A7)$$

The A-Function is defined as the ratio of the square of G_2 to G_1 and subtracting the constant:

$$F = \frac{G_2^2}{G_1} - C_0. \quad (A8)$$

The response equation for the A-Function is

$$R = \frac{(C_0 + D_2f(x, y))^2}{C_0 + D_1f(x, y)} - C_0, \quad (A9)$$

where the response function for $C_0 = 1$. Expanding equation (A9) yields

$$R = \frac{C_0(2D_2f(x, y) - D_1f(x, y)) + (D_2f(x, y))^2}{C_0 + D_1f(x, y)}. \quad (A10)$$

The response equation for the A-Function is a complex summation of products of functions with response functions. In addition, the presence of the second term in the denominator of equation (A10) acts to increase (decrease) the estimate for R depending on whether the sign of the term is negative (positive). This problem can be solved and equation (A10) simplified by choosing for C_0 a value that is much greater than the amplitude of $f(x, y)$. Thus terms not multiplied by C_0 can be neglected in equation (A10). Furthermore, from equation (A7) and the definition in equation (A3), let $k_1 = 2k_2$. Then

$$R = (2D_2 - D_2^2), \quad (A11)$$

which is equivalent to equation (A5). Thus the A-Function has the same response as the optimal 2-pass Gaussian method but without the additional interpolation.

Fire spread in chaparral—‘go or no-go?’

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Abstract. Current fire models are designed to model the spread of a linear fire front in dead, small-diameter fuels. Fires in predominantly living vegetation account for a large proportion of annual burned area in the United States. Prescribed burning is used to manage living fuels; however, prescribed burning is currently conducted under conditions that result in marginal burning. We do not understand quantitatively the relative importance of the fuel and environmental variables that determine spread in live vegetation. To address these weaknesses, laboratory fires have been burned to determine the effects of wind, slope, moisture content and fuel characteristics on fire spread in fuel beds of common chaparral species. Four species (*Adenostoma fasciculatum*, *Ceanothus crassifolius*, *Quercus berberidifolia*, *Arctostaphylos parryana*), two wind velocities (0 and 2 m s⁻¹) and two fuel bed depths (20 and 40 cm) were used. Oven-dry moisture content of fine fuels (<0.63 cm diameter) ranged from 0.09 to 1.06. Seventy of 125 fires successfully propagated the length (2.0 m) of the elevated fuel bed. A logistic model to predict the probability of successful fire spread was developed using stepwise logistic regression. The variables selected to predict propagation were wind velocity, slope percent, moisture content, fuel loading, species and air temperature. Air temperature and species terms were removed from the model for parsimony. The final model correctly classified 94% of the observations. Comparison of results with an empirical decision matrix for prescribed burning in chaparral suggested some agreement between the laboratory data and the empirical tool.

Introduction

Fire burns in living fuels such as chaparral in California, sagebrush and pinyon-juniper woodlands in the interior West, palmetto-gallberry in the south-eastern coastal plain, and coniferous forests in the USA annually. While these fires can be significant events, our ability to predict when fire will spread in these fuels is limited by two factors: (1) current fire spread models were not designed primarily for live fuels and (2) only a limited set of experimental data to develop and test models exists. Recent work in Europe and Australia has measured and modeled fire spread in various live fuels (Fernandes 2001; Marsden-Smedley *et al.* 2001; Bilgili and Saglam 2003; De Luis *et al.* 2004; Morvan and Dupuy 2004). In the USA, limited modeling of fire spread in live fuels has occurred (e.g. Albini 1967; Rothermel and Philpot 1973; Albini and Anderson 1982; Albini and Stocks 1986; Cohen 1986; Butler *et al.* 2004). While these various models exist and may be used by fire managers, empirical approaches to predict fire spread in live fuels are also used (e.g. Green 1981; Raybould and Roberts 1983; Campbell 1995).

In California, several different tools are used by fire managers to aid in the use of prescribed fire in chaparral. The key to successful use of prescribed burning is the development of a ‘prescription’ that defines the fuel, weather and fire conditions necessary to accomplish the objectives of the prescribed burn. BEHAVE (Andrews 1986), FIRECAST (Cohen 1986) and BRNPLN (Hilbruner 1988), all computer implementations of the Rothermel spread model (Rothermel 1972), can be used to estimate fire behavior under various weather settings. The matrix approach (Raybould and Roberts 1983) links a quasi-quantitative description of fire behavior and effects to a score computed from severity points assigned to various values of fuel and weather variables. Both the matrix approach and the Campbell Prediction System (Campbell 1995) utilize basic understanding of the variables that influence fire behavior to arrive at predictions.

Prescribed burning in chaparral is typically attempted in the spring to early summer when fuel moistures are higher in most cases than when wildfires occur (Green 1981). Burning conditions are often marginal and there seems to be a

threshold between no fire spread and successful propagation. Others have reported thresholds in fire behavior as influenced by various fuel and environmental variables (Bruner and Klebenow 1979). McCaw (1997) summarized many of these studies. While the factors influencing the spread thresholds in chaparral may be understood by successful prescribed fire managers, a systematic experimental investigation designed specifically to document these threshold conditions is currently lacking.

The basic formulation of the Rothermel model assumed that a fire would spread in the absence of wind and slope (R_0) (Rothermel 1972). Wind and slope functioned as multipliers of rate of spread:

$$R = R_0(1 + \phi_w + \phi_s),$$

where ϕ_w and ϕ_s are wind and slope multipliers respectively. The model does not predict a non-zero rate of spread in situations when wind or slope is required. For $R_0 = 0$, $R > 0$ if and only if $\phi_w > 0$, or $\phi_s > 0$ (Weise and Biging 1997). This phenomenon requires a different formulation for the effects of wind and slope on rate of spread. The Rothermel model was derived based on several simplifying assumptions. Fuels were assumed to be uniform, dominated by dead material and in close proximity to the ground. Environmental conditions were assumed constant. The various heat transfer mechanisms of radiation, convection and conduction were not explicitly described; a 'lumped capacity' approach was used. Fire spreads successfully in chaparral fuels at higher fuel moistures than most of the experimental data used to develop the Rothermel model. Under the influence of strong Santa Ana winds, nearly 304 000 ha were burned in southern California from 21 October to 4 November 2003 (Campbell 2004). Fuel moisture content in chamise chaparral in Los Angeles County was near 60% at this time (http://www.lacofd.org/Forestry_folder/LFM/Malibu.pdf). This level of live fuel moisture (60% or 0.6) is viewed by some southern California fire managers as a critical level indicative of high fire danger.

Wilson (1982, 1985, 1990) examined fire spread in moist fuel and proposed changes to the moisture content formulation in the Rothermel model. As part of this work, Wilson developed a 'predictive rule of thumb' to determine if fires would burn in wooden fuel beds. The rule of thumb predicted that a fire would not burn if $M \geq (\ln(2\sigma\beta\delta))/4$, where M , σ , δ and β are moisture content (expressed as proportion of oven-dry mass), fuel particle surface area-to-volume ratio (cm^{-1}), fuel bed depth (cm) and packing ratio (dimensionless, equation 1) respectively. Wilson (1990) also suggested that a term in the spread model involving wind and fuel moisture might be appropriate:

$$\beta = \frac{w_m}{(M+1)A\delta\rho_p} = \frac{w_o}{A\delta\rho_p} = \frac{\rho_b}{\rho_p}, \quad (1)$$

where w_m and w_o are total fuel bed wet and dry mass (kg), respectively, M is the live fuel moisture content, A is the surface area of the fuel bed, and ρ_p and ρ_b are fuel particle density and fuel bed bulk density (kg m^{-3}) respectively.

Equation (1) can be simplified to Rothermel's definition of packing ratio (Rothermel 1972); we used this form because our fuels were weighed when moist, not oven-dry as in other experiments. Wilson's rule of thumb was developed using data from fuel beds constructed of *Populus tremuloides* excelsior (shredded wood), and 0.64 and 1.27 cm diameter *Pinus ponderosa* sticks. In Wilson's experiments, fuel depth ranged from 2.5 to 20 cm, σ was 3.2, 6.3 or 81.3 cm^{-1} , and β ranged from 0.005 to 0.32. Fuel depths observed in chaparral crowns range from 30 to >120 cm, σ varies between various species from 20 to 60 cm^{-1} , and the crowns tend to be fairly porous (low packing ratio).

Countryman and Philpot (1970) reported packing ratios ranging from 0.00068 to 0.00374 for 16 chamise plants sampled near the location where fuels were collected for the current study. Reported fine fuel (leaves and stems <0.64 cm diameter) loading in various chaparral species mixes ranged from 0.71 to 3.33 kg m^{-2} (e.g. Countryman 1964; Ottmar et al. 2000). The application of Wilson's rule of thumb to shrub fuel beds such as chaparral is currently unknown. Given that current operational models do not adequately model fire spread in chaparral fuels and that data describing marginal burning conditions in chaparral do not currently exist, we have embarked on an experimental effort to determine the important fuel and environmental variables that determine propagation success in laboratory-scale fires in chaparral fuels.

Methods

The effects of wind, fuel moisture content, fuel bed height and slope on flame propagation in live fuels were investigated in a series of 125 experimental fires. Fuel beds (2 m long $\times$ 1.0 m wide $\times$ various depths) were constructed of live branch and foliage material collected from living chaparral growing at an elevation of 1160 m at an experimental area 50 km east of Riverside, CA, USA (Fig. 1a). Branches <0.64 cm diameter from manzanita (*Arctostaphylos parryana*), chamise (*Adenostoma fasciculatum*), hoaryleaf ceanothus (*Ceanothus crassifolius*) and scrub oak (*Quercus berberidifolia*) plants made up the fuels. The fuel beds were elevated above the surface of a tilting platform by 40 cm to simulate an aerial fuel. Air could be entrained from the ends of fuel beds; metal sheeting prevented air entrainment from the sides to reduce the curvature of the flame front and simulate a line fire. Plant material was generally collected in the morning so as to minimize moisture loss through transpiration. Dead fuel was removed to the extent possible. The fuels were then bagged and transported to the burn facility at the Forest Fire Laboratory. Moisture content was determined immediately before

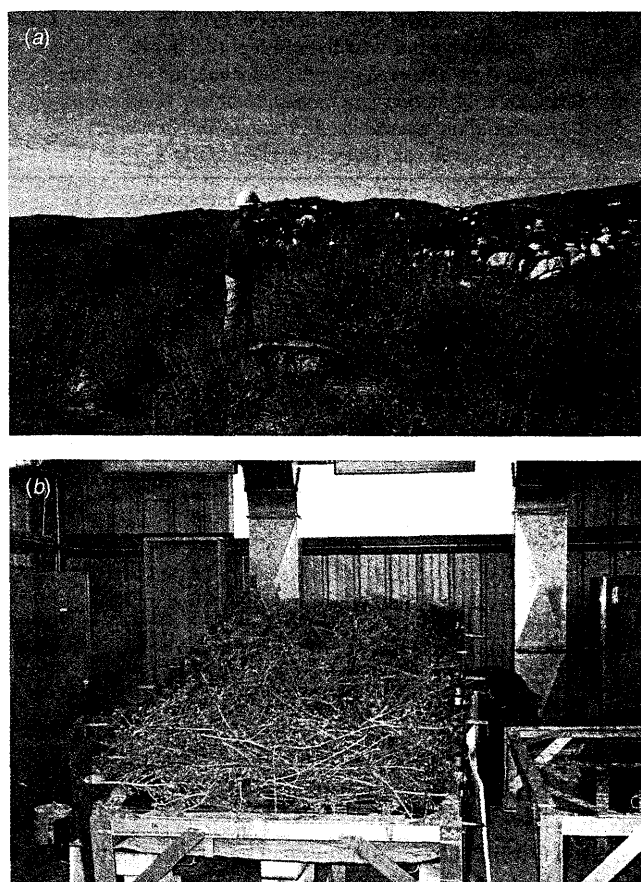


Fig. 1. (a) Fuel collection site located 50 km east of Riverside, CA, USA at an elevation of 1300 m in a mixed chaparral stand. Shrubs are young *Ceanothus crassifolius*. (b) Fuel bed constructed of live foliage and branches <0.64 cm diameter of *C. crassifolius*.

ignition. Fuel temperature was assumed in equilibrium with air temperature because air temperature at the collection site was generally only a few degrees different from the laboratory temperature. The five fuel beds constructed 24 h after collection using refrigerated fuels were allowed a few hours to come to equilibrium with ambient air temperature before ignition. The fuel was distributed uniformly in the fuel bed (Fig. 1b) to the greatest extent possible. Fires were ignited along the 1 m side with a flame zone depth of 50 cm section along the length of the live fuel bed. Between 300 and 400 g of excelsior and a small amount of isopropyl alcohol were added uniformly in the ignition zone to initiate and sustain the ignition.

While several variables potentially influence fire spread in live fuels, we selected three variables to control directly—wind velocity (0 or 2 m s^{-1}), species and fuel bed depth (20 or 40 cm). While we have the ability to impose slope, we have not completed a full set of experiments in which slope was a design variable. To date, 125 fires have been ignited (Table 1)—a fraction of the full factorial experiment, which will take several years to complete. The data are not

Table 1. Summary of experimental chaparral laboratory fires
Ceanothus, *Ceanothus crassifolius*; chamise, *Adenostoma fasciculatum*; manzanita, *Arctostaphylos parryana*; scrub oak, *Quercus berberidifolia*. Moisture content = (wet mass – dry mass)/dry mass

Species	Wind speed (m s^{-1})	Fuel bed depth (cm)	<i>N</i>	Moisture content range	Slope (%)
<i>Ceanothus</i>	0	20	6	0.66–0.93	0–70
	0	40	10	0.66–0.93	0–70
	2	20	5	0.66–0.88	0
	2	40	4	0.68–0.88	0
Chamise	0	20	19	0.09–0.91	0–70
	0	40	27	0.30–0.91	0–70
	2	20	9	0.39–0.91	0–40
	2	40	14	0.26–0.91	0–27
Manzanita	0	20	5	0.84–1.06	0–70
	0	40	6	0.84–1.06	0–70
	2	20	6	0.84–1.06	0
	2	40	4	0.84–1.06	0
Scrub oak	0	20	4	0.86–1.00	0–70
	0	40	2	0.86–1.00	0–40
	2	20	1	0.86–1.00	0
	2	40	3	0.86–1.00	0

balanced. As a result, individual species effects were not replicated evenly, were possibly confounded with other effects (for scrub oak), and may not have been captured in the analyses presented here.

Three rotary box fans (Model 9700*, 50.8 cm box fan; Air King Ventilation Products, West Chester, PA, USA) induced air flow to simulate wind. No attempt was made to 'smooth' out the vorticity in the flow. The fans produced an average velocity of 2 m s^{-1} above the fuel bed. We determined mean wind velocity by collecting 1-min point samples in vertical and horizontal transects above the fuel bed with an air flow mass velocity transducer (FMA-903; Omega Engineering, Stamford, CT, USA). The slope effect was generated by raising the down-wind end of the tilting platform. Ambient air temperature and relative humidity in the burn facility were measured at the beginning of each experiment using a Kestrel 3000 Pocket Weather Meter (Nielsen-Kellerman, Boothwyn, PA, USA). An experiment was described as successful if the live brush ignited from the burning zone and then propagated the length of the 2 m fuel bed (Fig. 2). The experiment was unsuccessful if the fire did not propagate. In the case of successful fire spread, the resulting rate of spread (ROS) was measured using data from 0.51 mm diameter type K thermocouples buried in the fuel bed and/or direct video image analysis from digital imagery collected at 30 Hz (Canon ZR10; Canon USA, Lake Success, NY, USA).

Because the response variable (spread success) was binary, logistic regression was used to develop a model to predict spread success (equation 2). The logit, $\log(p/(1 - p))$, of the

* The use of trade names and model numbers is for information purposes only and does not constitute endorsement by the U.S. Department of Agriculture.

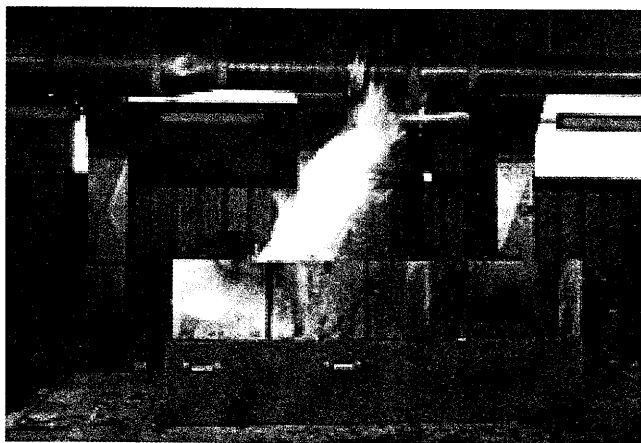


Fig. 2. A successful fire spread case in an *Adenostoma fasciculatum* fuel bed 40 cm deep with wind velocity of 2 m s^{-1} and no slope.

probability of spread success (p) was set equal to a linear function $X\theta$ of the predictor variables and the parameter estimates were determined by maximum likelihood estimation. Stepwise selection was used to choose the fuel bed and environmental variables to predict fire spread success. In order to compare the potential utility of the logistic model, several fire spread scenarios were developed and scored using the matrix approach of Raybould and Roberts (1983). Scenarios of air temperature, fuel moisture, wind velocity and fuel loading that were close to the laboratory conditions were developed:

$$\log\left(\frac{p}{1-p}\right) = X\theta \quad (2)$$

$$\Pr(\text{spread} = Y) = \frac{e^{X\theta}}{1 + e^{X\theta}}.$$

Results

In total, 125 tests of fire spread success were conducted (Table 2). Over half of the experiments used chamise as the fuel: ceanothus, 25 tests; chamise, 69; manzanita, 21; scrub oak, 10. Of the 125 tests, 79 were 'no wind' tests and 55 had a fuel bed depth of 20 cm. Seventy (or 56%) of the 125 tests resulted in successful fire spread. Chamise has been the dominant fuel tested because it is the dominant fuel at the collection site and it is the easiest to handle and produce uniform fuel beds (Table 2). A majority of the tests had no slope (93 of 125) or no wind (79 of 125).

The range in fuel moisture content for chamise was greatest (0.09–0.91). Most fuel beds were burned on the day of fuel collection to minimize moisture loss. Fuel beds burned 1–2 days after collection lost nearly half of their moisture, with resulting moisture content of 0.26–0.38. A single chamise fuel bed was allowed to dry for several weeks resulting in a moisture content of 0.09. For chamise fuel beds burned on the day of collection, moisture content exhibited a smaller range (0.64–0.91). This range is similar to that for the ceanothus

Table 2. Spread success of laboratory fires by species, slope percent, wind velocity and fuel bed depth

ROS, mean rate of spread (m min^{-1}) for success = 'Y' fires.

Ceanothus, *Ceanothus crassifolius*; chamise, *Adenostoma fasciculatum*; manzanita, *Arctostaphylos parryana*; scrub oak, *Quercus berberidifolia*

Variable	Value	Firespread success			
		No	Yes	Total	ROS
Species	Ceanothus	10	15	25	0.22
	Chamise	25	44	69	0.34 ^A
	Manzanita	13	8	21	0.20
	Scrub oak	7	3	10	0.74
Slope (%)	0	45	48	93	0.20
	27	3	6	9	0.38
	40	5	7	12	0.48
	70	2	9	11	0.72
Fuel depth (cm)	20	31	24	55	0.32
	40	24	46	70	0.31
Wind velocity (m s^{-1})	0	47	32	79	0.39
	2	8	38	46	0.26
Total		55	70	125	

^AMissing ROS observation for two tests of chamise with experimental settings of 0% slope, 40 cm depth and 0 m s^{-1} wind velocity.

fuel beds. The moisture content of the manzanita and scrub oak fuel beds was slightly higher. These fuel types do not typically have moisture contents that drop as low as chamise during the long summer droughts. The moisture content of the chamise fuel beds followed the typical annual trend in live fuel moisture in chamise, which is typical for many species in Mediterranean-type climates with mild, moist winters with rain followed by hot, dry summers with little rain (Fig. 3). Manzanita fuel moisture content increased from 0.95 in May 2003 to 1.06 in early July 2003, during which time new foliage was produced. Ceanothus and scrub oak exhibited increases in fuel moisture content similar to the manzanita increase.

Of the tests performed to date, wind was present in 30% (38/125) of the fires, aiding successful spread (Table 2). Because several experimental variables were examined in conjunction, care should be taken in identifying data trends from Table 2. For example, the 25 chamise fires that did not spread had different slope percents, fuel bed depths and wind velocities.

In the experiments reported here, physical fuel properties varied between species, but were assumed constant within a species. The exceptions to this were moisture content and packing ratio (Table 3). Manzanita, scrub oak, and ceanothus are all species with leaves that are generally ovoid in shape (Fig. 4). The leaf thickness varied between species with manzanita having the thickest leaves and scrub oak having the thinnest. In contrast, chamise leaves are linear in shape.

Several predictor variables were considered in the logistic equation: species, slope, moisture content, wind velocity, fuel bed depth, packing ratio, relative humidity, absolute temperature and fuel loading. Of the predictor variables considered,

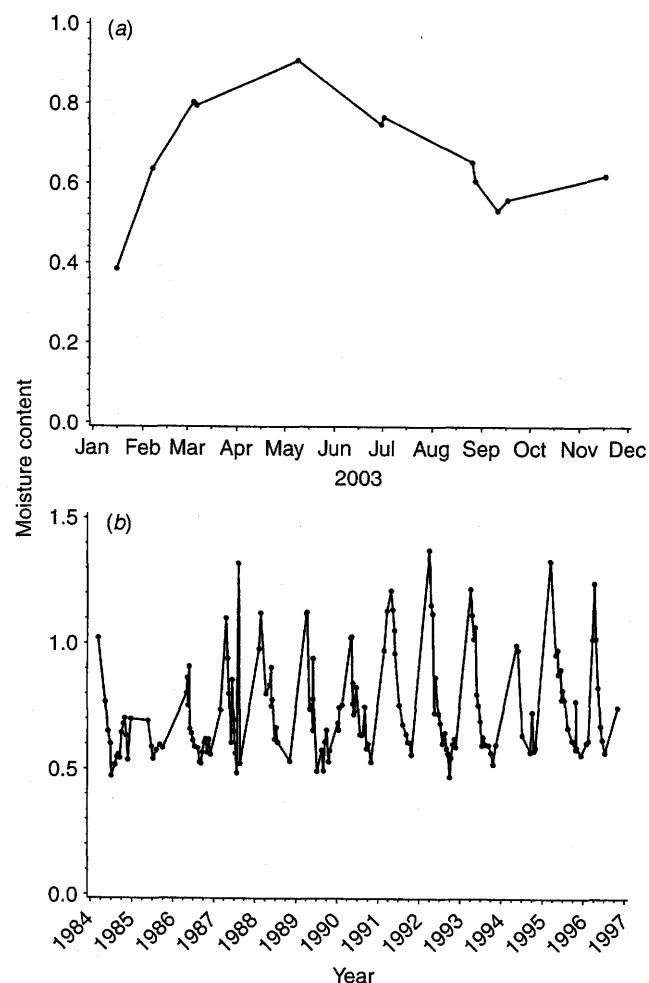


Fig. 3. Annual trend in chamise (*Adenostoma fasciculatum*) fuel moisture content for (a) laboratory fuel beds and (b) Strawberry fuel moisture sample site, San Bernardino National Forest.

Table 3. Physical properties of several chaparral species and fuel beds

Ceanothus, *Ceanothus crassifolius*; chamise, *Adenostoma fasciculatum*; manzanita, *Arctostaphylos parryana*; scrub oak, *Quercus berberidifolia*. Surface area-to-volume ratio: assumed 2000 ft⁻¹ for chamise (Cohen 1986), used 1771 ft⁻¹ for ceanothus (*Ceanothus velutinus*, Countryman 1982), measured sample for manzanita, unknown for scrub oak. Packing ratio, see equation (1)

Species	Surface area-to-volume ratio (σ) (cm ⁻¹)	Packing ratio (β)
Chamise	66	0.009–0.013
Ceanothus	58	0.012, 0.013
Manzanita	41	0.014, 0.016, 0.019, 0.020
Scrub oak		0.009, 0.013, 0.017

wind velocity was selected as the best single predictor of fire spread success (Table 4). Of nine variables considered, five were selected to predict the probability of fire spread success. Three of the variables were environmental (wind



Fig. 4. Foliage and fine branch samples of four chaparral species used in marginal burning experiment: (1) scrub oak (*Quercus berberidifolia*); (2) manzanita (*Arctostaphylos parryana*); (3) chamise (*Adenostoma fasciculatum*) and (4) hoaryleaf ceanothus (*Ceanothus crassifolius*). Coin = 1.9 cm diameter.

velocity, slope %, air temperature) and two were fuel (fuel loading, moisture content) variables. Plant species was also selected as a significant variable; however, for parsimony, species was not included in the final model because it was the last variable selected and inclusion of species in the model increased the classification accuracy only from 96 to 98% (Table 4). Correlation between the parameter estimates was examined and air temperature was also removed from the equation as it was highly correlated with the intercept term (correlation = -0.99). The equation that resulted from the stepwise logistic regression is:

$$X\hat{\theta} = 1.65 + 2.99WS + 0.10SI + 2.03L - 12.55M \quad (3)$$

$$\Pr(\text{success} = Y) = \frac{\exp(X\hat{\theta})}{1 + \exp(X\hat{\theta})},$$

where WS is wind velocity (m s⁻¹), SI is slope (%), L is fuel loading (kg m⁻²) and M is live fuel moisture content (proportion). Correlation of parameter estimates for moisture content with wind velocity and slope was between 0.7 and 0.8; correlation between wind and slope was also in this range. The correlation between wind velocity and slope is most likely due to the limited number of tests with slope. This simplified model (equation 3) correctly classified 94% of the 125 tests. The Hosmer and Lemeshow (2000) lack of fit test indicated no significant lack of fit of this model ($P > 0.1$).

The following discussion was condensed from Hosmer and Lemeshow (2000) to explain the odds ratio. For a dichotomous discrete variable $z = 0, 1$, the odds ratio approximates

Table 4. Variables selected to predict fire spread successd.f., degrees of freedom. Ceanothus, *Ceanothus crassifolius*; chamise, *Adenostoma fasciculatum*; manzanita, *Arctostaphylos parryana*; scrub oak, *Quercus berberidifolia*

Variable	Selection order ^A	d.f.	$\hat{\theta}$		Odds ratio	
			With species	Without species	With species	Without species
Intercept		1	-118.70	1.65		
Wind velocity	1	1	7.29	2.99	>0.999	19.85
Slope	3	1	0.19	0.10		1.10
Moisture content	4	1	-21.25	-12.55	<0.001	<0.001
Dry fuel loading (kg m ⁻²)	2	1	3.65	2.03	38.3	7.65
Air temperature (K)	5	1	0.39		1.5	
Species						
Ceanothus	6	1	1.54		10.4 ^B	
Chamise	6	1	2.39		24.3	
Manzanita	6	1	-3.13		<0.01	
Fires classified correctly (%)			98	93		

^AEffect of intercept (overall mean) is removed first by algorithm, then variables are selected according to criteria.^BOdds ratio for individual species in comparison with Scrub oak.

how much more likely it is for a fire to spread when $z = 1$ than when $z = 0$. The odds ratio is

$$\frac{\pi(1)/[1 - \pi(1)]}{\pi(0)/[1 - \pi(0)]},$$

where $\pi(z)$ is the conditional probability that the outcome is present given z . The odds ratio approximates relative risk and is estimated with $\exp(\hat{\theta}_i)$. For polytomous discrete variables, the calculation of the odds ratio is slightly more involved and depends on how the polytomous variable is coded. In our study, we used effects coding for the species variable:

- $\hat{\theta}_1$ estimates the effect of species = 'ceanothus';
- $\hat{\theta}_2$ estimates the effect of species = 'chamise';
- $\hat{\theta}_3$ estimates the effect of species = 'manzanita'; and
- $-\hat{\theta}_1 - \hat{\theta}_2 - \hat{\theta}_3$ estimates the effect of species = 'scrub oak'.

The odds ratio of Ceanothus relative to Scrub oak is

$$\exp(\hat{\theta}_1 - (-\hat{\theta}_1 - \hat{\theta}_2 - \hat{\theta}_3)) = \exp(2\hat{\theta}_1 + \hat{\theta}_2 + \hat{\theta}_3).$$

Substituting in the estimates from Table 4 yielded $\exp(2(1.54) + 2.39 - 3.13) = 10.38$. For the continuous variables wind velocity, dry loading, etc., the coefficient ($\hat{\theta}_i$) 'gives the change in the log odds for an increase of '1' unit in' the continuous variable. The odds ratio for an arbitrary change c in a continuous variable is $\exp(c\hat{\theta}_i)$ (Hosmer and Lemeshow 2000: 63). In Table 4, the odds ratios presented are for $c = 1$. The odds ratio for wind velocity in the 'without species' model is $\exp(2.99) = 19.89$. The difference is due to rounding. The odds ratio describes the change in risk (or likelihood of occurrence) of fire spread that is associated with a change in the predictor variable. The higher the odds ratio is for a variable, the higher the likelihood that the outcome will occur if the predictor variable changes. This analysis indicated that fire spread was more likely to occur with a

Table 5. Selected prescribed fire scenarios and severity scores

SCAL fuel model: B, mixed brush; C, chamise (Cohen 1986).

Assumed a fuel bed depth of 30 cm, air temperature of 294 K, slope = 0 and relative humidity = 60% for all predictions. Loading in kg m⁻². Severity score: see Raybould and Roberts (1983). Pr, estimated probability of successful fire spread (equation 3)

Fuel model/ loading	Scenario	Live fuel moisture content (%)	Wind velocity (m s ⁻¹)	Severity score	Pr
B: 1.92	1	0.9	0.0	40	0.003
	2	0.9	2.2	43	0.718
	3	0.9	4.5	47	1.000
	4	0.7	0.0	41	0.038
	5	0.7	2.2	44	0.969
	6	0.7	4.5	48	1.000
C: 0.74	7	0.9	0.0	43	0.000
	8	0.9	2.2	46	0.189
	9	0.9	4.5	50	0.995
	10	0.7	0.0	44	0.004
	11	0.7	2.2	47	0.741
	12	0.7	4.5	51	1.000

wind velocity of 2 m s⁻¹ than with no wind. As moisture content increased, the likelihood that fire spread would occur decreased (indicated by a small odds ratio <1). For the 'with species' model, fire spread was more likely to occur (relative to scrub oak) in fuel beds of ceanothus and chamise. Fire spread in manzanita fuel beds was less likely than fire spread in scrub oak fuel beds. Increased fuel loading increased the likelihood of fire spread success.

In an attempt to determine if laboratory results were similar to guidelines developed for prescribed fire use in chaparral, we selected 12 combinations of fire prescription variables and estimated a severity score following Raybould and Roberts (1983) (Table 5). The scenarios were created to mimic conditions as close to what was observed in the

laboratory as could be accomplished. The matrix has discrete classes for each fire prescription variable, which are assigned severity points. We calculated the probability of fire spread using equation (3) for both SCAL fuel model B (mixed brush) and fuel model C (chamise) (Cohen 1986). We assumed fuel continuity of 100% (severity points = 9), fuel bed percent dead = 15% (lowest possible value, 1 point), aspect = north (1), season = July (3), 10 h stick fuel moisture = 0.15 (1), time of day = 1300 (7) and age = 20 years. Loading for foliage and live wood <0.64 cm diameter was calculated using FIRECAST fuel load equations (Cohen 1986). For the two SCAL fuel models, only the foliage and 1 h fuel loading were used (equation 4) because the fuel beds were constructed from this size of fuel:

SCAL B

$$L = 0.224 \frac{\text{Age}}{0.4849 + 0.0170\text{Age}} (0.4373 - 0.561\text{FD}) \quad (4)$$

SCAL C

$$L = 0.224 \frac{\text{Age}}{1.4459 + 0.0347\text{Age}} (0.4373 - 0.561\text{FD}),$$

where L = fuel loading (kg m^{-2}), Age is chaparral stand age (years) and FD is the fraction of the total fuel bed that is dead.

The scenarios in Table 5 illustrate the relative sensitivity of the fitted logistic model to changes in wind velocity and fuel loading. If fuel model was held constant, the predicted probability of spread success increased appreciably when wind velocity increased from 0 to 2.2 m s^{-1} (scenarios 1,2; 4,5; 10,11). For both fuel models, predicted spread success was unity for wind velocity of 4.5 m s^{-1} . The combined effect of loading and species on predicted probability can be seen when comparing predicted probabilities for scenarios 2 and 8. It is not possible to separate the effect of loading from the effect of species. However, the effect of species differences in physical properties such as surface area-to-volume ratio is reflected in the score for these same scenarios. We anticipate that species will have a significant effect on prediction of fire spread success as the dataset becomes more balanced.

Raybould and Roberts (1983) related the severity score to a general description of fire behavior and effects in chaparral. The severity scores for the 12 scenarios in Table 5 would be considered 'moderate burns' in which 50% of the area might be burned, the fuel burned easily, and flame lengths of 1.2–1.8 m would be observed. It is not possible to directly relate the severity score and associated fire behavior to our data. However, burning half of the area implies that fire might spread successfully in half of the area. Stated differently, fire spread success = 0 for the unburned half. The severity scores in Table 5 are in the prescription window where marginal burning in chaparral has been observed to occur. While these results do not indicate that equation (3) can be used to predict prescribed fire success, they suggest that the experimental

work we are conducting may be similar to the field scale prescribed burns that we hope to better model. These results also suggest that the transition from no spread to successful spread in live fuels is a step function influenced by wind velocity.

With these data, it was not possible to determine if live fuels burn like wet dead fuels. Several authors have reported that rate of spread is reduced as fuel moisture content of dead fuels increases (e.g. Wilson 1990; Catchpole *et al.* 1998). For woody fuels, moisture content has been manipulated experimentally to produce a range of 0.02–0.33 in work at the Forest Service Fire Laboratory in Missoula, MT, USA. This range of moisture content is representative of the range reported by numerous authors over several decades for studies of fine, dead woody fuels. Catchpole *et al.* (1998) reported that fuel bed depth had little effect on spread rate in woody fuels. The stepwise logistic regression procedure did not select fuel bed depth as a predictor of fire spread success even though the data suggested otherwise (Table 2). Additional experiments may clarify this relationship. These data illustrate that wind is sometimes required for successful ignition and propagation in live fuels. This suggests that convective heating may play a more significant role in flame spread in live fuels than in dead fuels. The recent radiation-based crown fire model may need an additional heat transfer term for convection (Butler *et al.* 2004). Other concurrent work is examining combustion aspects of live fuels (Engstrom *et al.* 2004) and physical modeling of marginal burning (Zhou *et al.* 2005). With additional laboratory experimentation and field validation, we hope to provide land managers with a tool to improve their ability to predict prescribed burn success in live fuels.

Summary

Current fire spread models do not adequately model the transition between no spread and spread in live fuels. We conducted 125 experiments to determine the importance of fuel and environmental variables on fire spread success for four different species of chaparral in the laboratory. Analysis of these 125 fires indicated the importance of wind velocity and dry loading on fire spread success in live fuels. Fuel bed depth did not affect fire spread success. Increasing moisture content decreased the probability of successful spread. This has also been observed in dead woody fuels. Sustained wind-aided fire spread in live fuels has been demonstrated at moisture contents higher than previous dead woody fuel experiments. A logistic model to predict fire spread success using wind velocity, slope, moisture content and fuel loading has been developed. Comparison of results with a tool designed to determine fire behavior and effects in chaparral suggested that the laboratory data for marginal fire spread in chaparral are in general agreement with the empirical tool. Additional laboratory experimentation and comparison with field scale results are necessary before equation (3) can be used by land

managers to predict the success of planned prescribed fire operations.

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Smoke measurements during Gestosa-2002 experimental field fires

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Abstract. Currently, there is a growing awareness that smoke produced during forest fires can expose individuals and populations to hazardous concentrations of air pollutants. Aiming to contribute to a better understanding of the air pollution phenomenon associated with forest fires, this paper presents and analyses the atmospheric emissions and air quality concentration measurements performed in the 2002 fire experiments at Gestosa, Central Portugal. Two vehicles were equipped with a meteorological station and air quality analysers that were turned on continuously to acquire concentrations of particulate matter, nitrogen oxides and carbon monoxide. Nitrogen and sulfur dioxides were measured using a grid of fixed passive samplers. Also, firefighters and research-team members used passive samplers during the experiments in order to estimate the human exposure to these pollutants. Measurements of volatile organic compound emissions, using Tedlar bags, were carried out. Results were analysed taking into account not only the concentration values but also the variables involved, such as the combustion phase and the meteorology, and identifying possible relationships between them. Despite the small size of the burning plots when compared to wildfires, the measured levels of pollutants were however considerable, indicating the effect of these experiments on the local air quality and stressing the serious levels of air pollution that can be expected during wildfires.

Additional keywords: air quality; forest fire emissions.

Introduction

Concern associated with smoke from forest fires has been increasing during recent years. The severe air pollution episodes caused by fires in Amazonia (Brazil), Indonesia and the Philippines in 1997–1998, and more recently in Australia and Russia, has drawn worldwide attention to this problem. In Portugal, the summer of 2003 was considered the most devastating of the last decade, with fewer fires but ~4 times the average area burned annually. Also a clear effect on the air quality values measured by the national monitoring network was found (Martins *et al.* 2004). Currently, there is a growing awareness that smoke from forest fires can expose individuals and populations to hazardous air pollutants, stressing the importance of knowing in what conditions serious health effects can be expected and what procedures should be followed to prevent its occurrence or to diminish its effects. This concern also is associated with the use of prescribed fires, namely in Australia and North America where this fire management technique is used frequently. The World Health Organization (WHO) has developed policies and guidelines to reduce the health impacts of smoke generated during vegetation burning (WHO/UNEP/WMO 1999) and the legislative requirements regarding air quality should be met when performing prescribed burns in North America (Riebau and Fox 2001).

The type and amount of products emitted into the atmosphere from biomass burning are extremely dependent on the type of fuel, fire line intensity, fuel moisture, wind and fire temperature. The chemistry, phase (flaming and smouldering) and efficiency of the combustion process are of fundamental importance when estimating forest fire emissions. The effects of smoke on air quality cannot be determined through the estimation of emissions alone. Air pollution assessment requires estimates of the pollutants concentration some distance from sources with known emission characteristics and, in this respect, the atmospheric flow has a fundamental role in smoke transport and dispersion.

A significant body of research results can be found in the literature about the quantification of gaseous and particulate emissions from fires and the calculation of emission factors for both wildfires and prescribed burns. Andreae and Merlet (2001) presented and reviewed emission factors for biomass burning. Also, the WHO has produced a complete document on this subject (WHO/UNEP/WMO 1999), and Ward and Radke (1993) reviewed the methods to measure emissions from vegetation fires, ranging from very small controlled combustion experiments to satellite techniques. The USDA Forest Service has developed intensive measurements of smoke exposure among firefighters during both prescribed burns (Reinhardt *et al.* 2000) and wildfires (Reinhardt and

Ottmar 2000). In the scope of the European Commission (EC) ERAS Project ('Extension Retardant Application System'), a portable device for smoke analysis is being tested for the measurement of air pollutant concentrations during forest fires (NTUA 2004). Forest fire detection by smoke sensing with light detection and ranging (LIDAR) technology has led to some interesting results (Utkin *et al.* 2003). However, air pollutant concentration data acquired during forest fires and accompanied by other important data, like meteorology, visibility, fire and combustible characteristics, are still limited and it is not possible to completely understand and evaluate the effects of forest fires on air quality. The main purpose of this work is to contribute to a better understanding of forest fires as a source of pollutants to the atmosphere.

This paper presents the concentration values of a set of air pollutants measured during experimental fires in Central Portugal, Gestosa-2002, and analyses the acquired values in an integrated way, taking into account other variables involved, such as fire stage and meteorology, and identifying possible correlations in order to contribute to a better understanding of the air pollution phenomenon associated with forest fires. This work was performed within the scope of the SPREAD Project ('Forest Fire Spread Prevention and Mitigation'), a research project funded by the EC, and several other teams were involved. Part of the data used in the paper results from these teams' activities. The particular location of the study area, southern Europe, should be stressed because despite the intense occurrence of forest fires in this region, the information concerning air quality impacts is mainly available from other parts of the world.

Methodology

The burning experiments performed since 1998 at Central Portugal, Gestosa, aim to collect a large range of different but complementary experimental data, which should be used to support the development of new concepts and models and to validate existing methods or models in various fields of fire management (Viegas *et al.* 2002). These experiments, involving several research teams and covering a very extensive characterisation of variables related to fire behaviour, constitute a valuable source of fire-related data within the European territory and a particularly important opportunity to measure and analyse air pollutants concentrations during experimental field fires.

From the first field campaign (1998) until the last one (2004) it is possible to discern a clear evolution of the measurement approaches and techniques applied, as a result of the knowledge acquired during the previous burnings and the optimisation of the experimental procedures (Miranda and Borrego 2002). During 2002 main burns, conducted on 30 and 31 May, measurements of atmospheric emissions and air quality concentrations were taken as described below. Selection of the specific burning days is very dependent on

weather conditions. Usually burning experiments last two spring days and everything is organised towards the end of May, beginning of June, aiming to avoid rainy days and the beginning of the season when experimental fires are not allowed.

Study area characteristics

The study area was located in Central Portugal (40°15'N, 8°10'W), in a hillside of 'Serra da Lousã' with altitudes between 800 and 950 m. To guarantee the safety of the personnel and equipment involved and to assure a good organisation of the experimental program, the area was divided into 10 plots with regular shapes and different dimensions, which were separated by fuel breaks with widths between 5 and 15 m, to limit fire spread and to keep it inside the desired boundaries (Viegas 2002). These experimental burning plots, represented in the photograph of Fig. 1, were established within Forest Service lands, and within the Gestosa forestry perimeter.

Systematic vegetation sampling was carried out along the plots (Viegas 2002). Non-destructive sampling along linear transects was made to determine vegetation cover, species composition and vegetation height. For each plot, the results from previous destructive samplings were used to evaluate the biomass of each species. Table 1 presents a set of data that characterise the experimental plots. The plots areas range from ~0.4 (plot 517) to 1 ha (plot 520). The plots are SW and SSW oriented, with a terrain slope varying between 17 and 27° (smaller plots present higher slopes). The existing vegetation consists mainly of continuous shrubs (*Erica umbellata*, *Erica australis* and *Chamaespartium tridentatum*) with some isolated *Pinus pinaster* trees. Fuel properties were assumed as homogeneous within each plot (see Table 1).

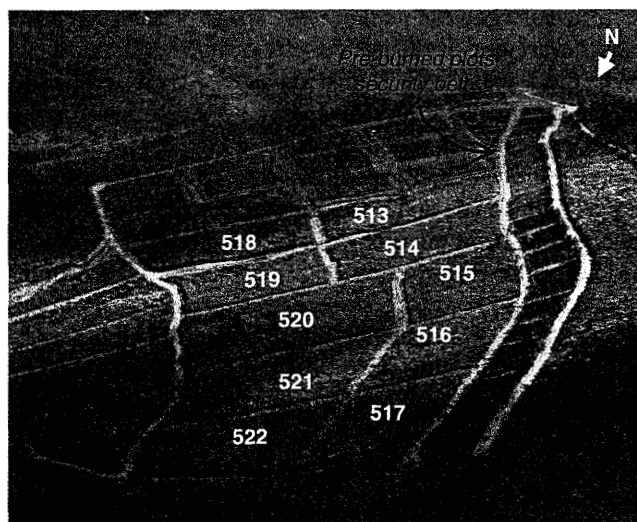


Fig. 1. Gestosa-2002 study area, with location of experimental plots and respective serial numbers.

Measuring equipment and techniques

During the experimental fires, which occurred on warm and dry days, temperature, humidity and wind speed and direction were measured at several locations, near the fire plots. Specific techniques and equipment were used to obtain the concentrations of different pollutants. Table 2 summarises the applied measuring techniques.

Two luggage vans, both located near the burning plots, were equipped with meteorological measuring equipment and air quality analysers. Van 1 measured concentrations of particles with an aerodynamic diameter lesser than $2.5 \mu\text{m}$ ($\text{PM}_{2.5}$), and van 2 measured particles smaller than $10 \mu\text{m}$ (PM_{10}), nitrogen dioxide (NO_2) and nitric oxide (NO), and carbon monoxide (CO). Table 3 presents the distance between the vans and the centre of the nearest plots. In Fig. 2, a photo of van 2 and its air quality equipment is shown.

The continuous acquisition of NO and NO_2 concentrations in air was performed using the automatic equipment Environnement AC31MTM (dual chamber chemiluminescent nitrogen oxides; Environnement S.A., Poissy, France). CO was measured continuously with the Environnement CO11M analyser, whose functioning principle is based on the selective absorption of infrared radiation by the CO molecules. To monitor PM_{10} and $\text{PM}_{2.5}$ concentrations, two Environnement MP101M analysers were used with adequate sampling inlets for each diameter. A β -gauge mass monitor determines the particle's mass.

Taking into account one of the advantages of passive samplers (Radiello equipment), i.e. its portability, a grid of NO_2 and sulfur dioxide (SO_2) samplers was defined according to the local dominant winds, at the top of the experimental field and along two lines, allowing us to obtain a larger spatial cover

Table 1. Main characteristics of the experimental plots (Viegas 2002)

Plot	Dimension (m)		Slope (°)	Mean height (m)	Total cover (%)	Fuel load (kg m^{-2})
	Width	Length				
513	58	97	21	1.2	89	6.3
514	85	90	21	1.1	97	9.9
515	87	53	27	0.8	98	5.4
516	101	51	22	1.1	88	8.7
517	86	52	24	1.3	100	11.1
518	58	108	17	1.6	97	11.0
519	89	91	21	1.2	98	7.8
520	89	109	18	1.2	95	5.7
521	87	99	19	1.3	100	6.6
522	68	90	18	1.2	100	7.2

Table 3. Average distance between the vans and nearest plots

Van	Plot	Distance (m)
1	515	200
	516	110
	517	65
	521	150
	522	125
2	513	130
	514	75
	515	90
	516	150
	518	200
	519	170
	520	180

Table 2. Summary of air pollutant measurement techniques during Gestosa-2002 experiments

Pollutant	Technique	Type of data	Equipment	Characteristics
NO_x (NO , NO_2)	Automatic equipment, van 2	Continuous measurement: 1 min average	Environnement AC31M TM	Range: 0–10 ppm (programmable) Noise: 0.17 ppb Lower detectable limit: 0.35 ppb Response time: automatic and programmable (minimum 20 s)
CO	Automatic equipment, van 2	Continuous measurement: 1 min average	Environnement CO11M TM	Range: 0–200 ppm (programmable) Noise: 0.025 ppm Lower detectable limit: 0.05 ppm Response time: automatic and programmable (minimum 30 s)
Particulate matter				
$\text{PM}_{2.5}$	Automatic equipment, van 1	Continuous measurement: 15 min average	Environnement MP101M TM	Range: 0–10 000 $\mu\text{g m}^{-3}$ (programmable)
PM_{10}	Automatic equipment, van 2	Continuous measurement: 15 min average	Environnement MP101M TM	Range: 0–10 000 $\mu\text{g m}^{-3}$ (programmable)
NO_2 , SO_2	Passive samplers and laboratory analysis	Variable sampling period (see Tables 5–7): 1 h average	Radiello	
VOC	Sampling in Tedlar bags and laboratory analysis with a FID	Instantaneous sampling		



Fig. 2. Photo of van 2 and its air quality equipment.

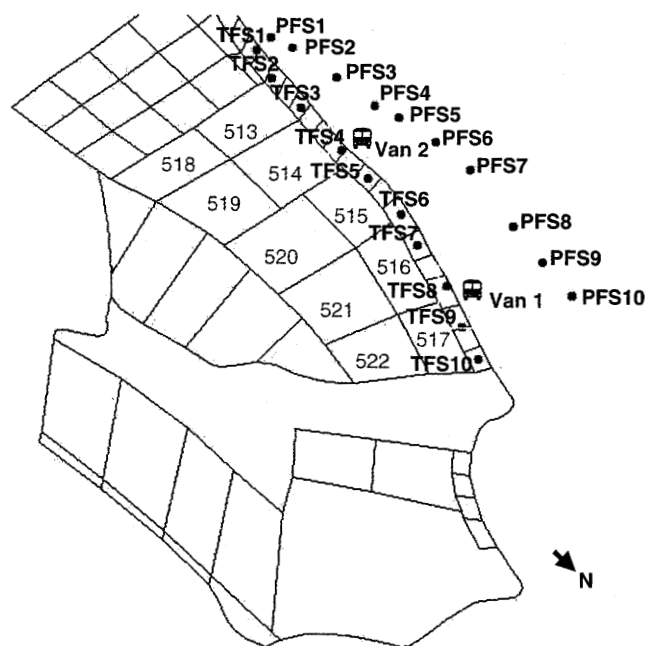


Fig. 3. Gestosa-2002 study area, with location of measuring equipment.

of the pollutants' dispersion. The sampling was performed with triethanolamine diffusivity passive samplers and the subsequent laboratory analyses were made by ionic chromatography. Some of the samplers were changed at various times during a day, aiming to evaluate how the characteristics of the burned plots influenced the air pollutant concentrations. Replicates of each passive sampler were used. In Fig. 3 the location of the two lines of passive samplers is presented; the first one, in which the samplers were replaced one or two times a day (temporary-fixed samplers; TFS), was closer to the burning area, while the other one contained the



Fig. 4. Photo of passive samplers.

devices that sampled during the whole day (permanent-fixed samplers; PFS).

Various firefighters and members of the research team carried a mobile passive sampler (MS) during the experiments in order to estimate the human exposure to NO_2 and SO_2 . Figure 4 shows an example of the passive samplers used during the experiments, namely the fixed sampler at location 6 and a member of our team using one diffusive tube.

Measurements of volatile organic compound (VOC) emissions were carried out during the flaming and the smouldering phases of each plot, by pumping the smoke into Tedlar bags for subsequent laboratory analyses with a flame ionisation detector (FID).

Experimental procedure

Information concerning the set-up and development of the experiments, namely the beginning and the end time of each plot burning, and some particular techniques to ignite and extinguish the fire that were tested during Gestosa-2002, is given in Table 4. Burning time concerns the period from ignition until the extinguishment of the flames.

Table 4. Plot burning information and time schedule

Day	Plot	Local time		Burning time (min)	Observations
		Begin	End		
30 May	513	10:19	10:44	25	Linear ignition at top and bottom
	517	11:26	11:44	18	Linear ignition at top, upper third pyrotechnic
	516	12:21	12:28	7	Single line at the bottom
	514	13:11	13:30	19	Hoses of explosives filled with water
	515	16:20	16:39	19	Hoses of explosives filled with water
	518	17:04	18:05	61	Stronger wind blowing downslope
31 May	522	11:00	11:30	30	Safety burn
	521	12:00	12:30	30	Point ignition
	520	13:45	14:03	48	Hoses of explosives filled with water
	519	14:32	14:51	19	Oblique ignition

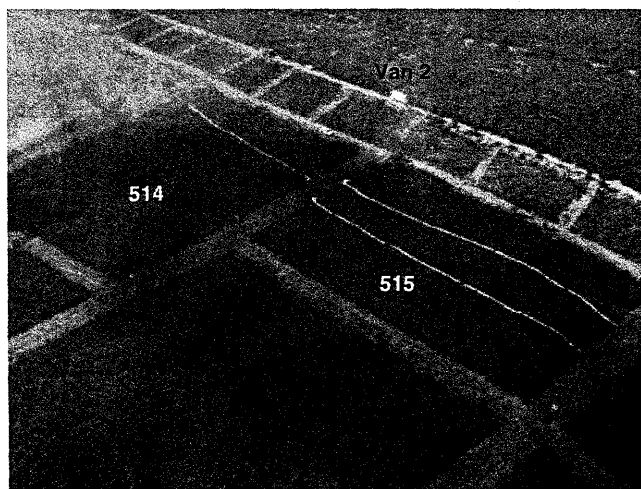


Fig. 5. Aerial view of hoses (white lines) in plots 514 and 515.

On the morning of the first day, plots 513 and 517 were burned. In the first one, linear ignition was used at its top and then bottom. Plot 517 was burned by linear ignition on the top and then linear ignition on its upper third with pyrotechnic devices (Viegas *et al.* 2002).

One of the most striking events of Gestosa-2002 was the test with the German hoses filled with water and with a detonating cord that exploded just before the fire reached them (Viegas 2002). This technique was applied to plots 514 and 515 as it can be seen in Fig. 5. The fire front was extinguished after a short time in these plots where the blasting hoses were used. Aiming to identify the influence of using this extinguishing technique, some of the passive diffusers closer to the burning area were replaced before burning plot 514 (TFS4, TFS5, TFS6 and TFS7) and the others were changed between the plot 514 and 515 burns.

Presentation and analysis of results

Figure 6 presents the meteorological and air quality data acquired during the first day of the experiments by the automatic equipment installed inside the luggage vans. Only

data acquired during the first day of experiments are presented because for the second day, smoke from only one burning plot impacted a single van, and measured values were similar to those of the first day. Dotted lines represent the beginning and the end time for each burning plot. Between 14:30 and 16:00 there are no data because it was lunchtime and experiments were stopped. Limit air quality concentration values settled by the European Legislation are also represented in the graphs, namely the daily average for PM_{10} ($50 \mu g m^{-3}$), established in the Council Directive 1999/30/EC; the 8 h average for CO ($10 mg m^{-3}$) defined by 2000/69/EC; and the hourly average for NO_2 ($200 \mu g m^{-3}$) implemented by 1999/30/EC. However, concerning NO_2 a margin of tolerance, established for the year 2002, permits a maximum hourly average of $289 \mu g m^{-3}$.

It is possible to verify that the burns occurred with weak winds (rounding $2 m s^{-1}$) blowing from NE to SE in the morning hours and changing towards NW during the afternoon. In the afternoon, burning of the first and second plots was influenced by a variable wind direction. This wind behaviour is closely related to the air pollutant concentrations measured in each van.

Notwithstanding the close location of the two vans, separated by 250 m, they were affected by emissions coming from different plots according to the wind direction and demonstrating the considerable effect of topography over the wind field and, consequently, the smoke dispersion. Van 1 registered the influence of emissions from plots 517 (Figs 6, 7a) and 516, and van 2 the effects of plots 514 and 515 (Figs 6, 7b,c). In Fig. 7, photos of plots 517, 514 and 515 burning are presented.

In general, air quality equipment registered the effect of fire emissions only after the end of each burn. This can be explained by the fact that, due to the extremely high temperatures reached during the flaming stage, which induce the formation of a convection column, the smoke plume was forced to rise, not reaching the vans that were located at the vicinity of the plots. With the diminishing of the amount of heat released, local winds became stronger than these thermally induced circulations, transporting the smoke towards

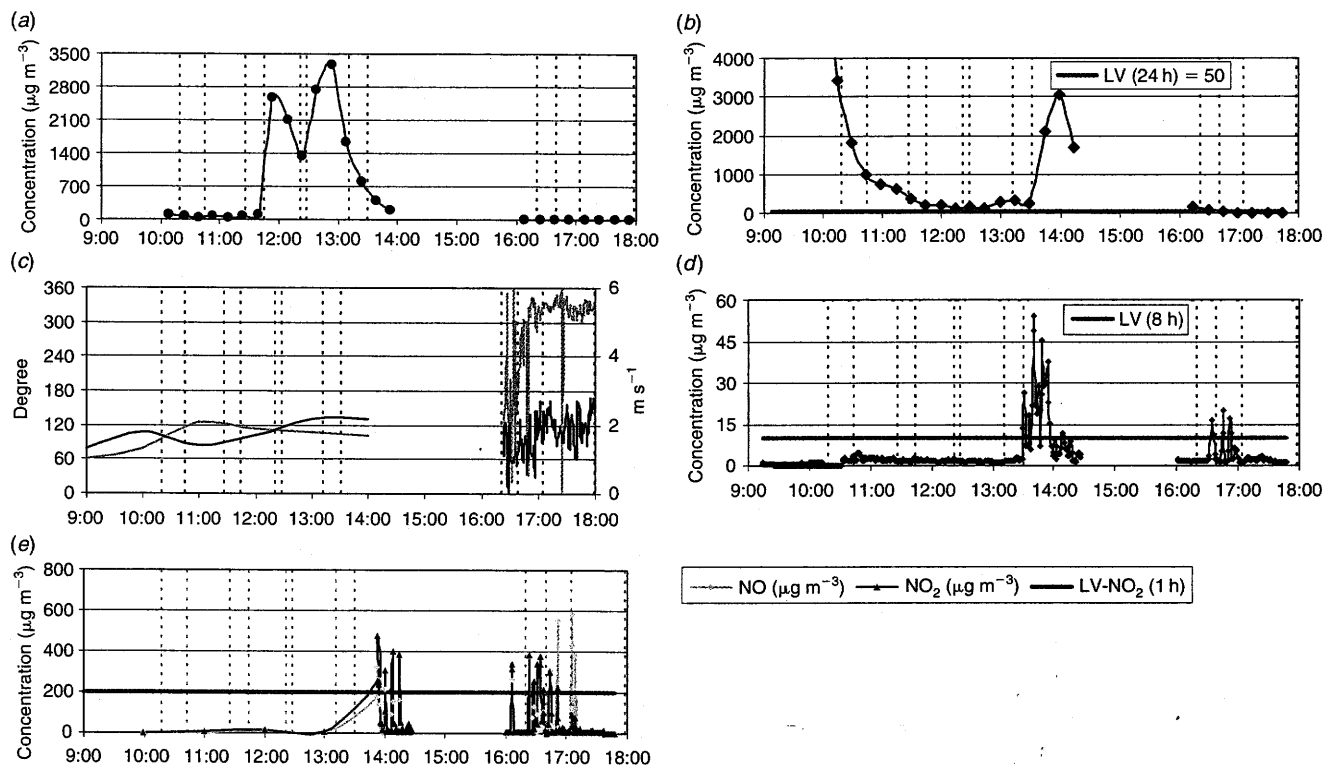


Fig. 6. $PM_{2.5}$ (van 1) (a), PM_{10} (van 2) (b), wind speed and direction (van 2) (c), CO (van 2) (d), and NO and NO_2 (van 2) (e) values measured during the first day of experimental fires (vertical dotted lines represent the burning time of each burning plot).

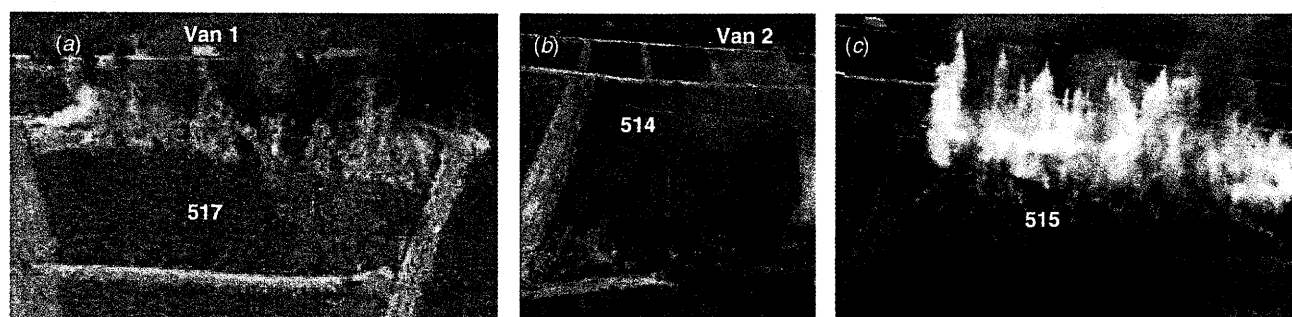


Fig. 7. Photos of plots 517 (a), 514 (b) and 515 (c) (at the precise moment of the explosive hoses detonation) fire front.

the vans. The first burn of the experiments, plot 513, did not affect the air quality values measured in either van because the wind blew from the east and the smoke plume did not reach the vans.

As can be seen in Fig. 7a, van 1 was affected by plot 517 emissions. The van was located just above the plot (65 m distance) and wind blew from the SE, transporting the smoke in its direction. Fifteen minute-averaged $PM_{2.5}$ concentrations reached $2500 \mu g m^{-3}$. The following burning plot also contributed to $PM_{2.5}$ concentration values measured in van 1, reaching even higher values ($3000 \mu g m^{-3}$). The averaged concentration for the time period the equipment was functioning (~ 6 h) is $646 \mu g m^{-3}$. Even considering that during

the rest of the day the concentration was $0 \mu g m^{-3}$, the 24 h average is $160 \mu g m^{-3}$. Although there is no legislated value for $PM_{2.5}$ in Europe, in the USA the legislation established a limit of $65 \mu g m^{-3}$ for a 24 h average. Both the peak and the average values correspond to a hazardous category of air quality and very low visibility, less than 1.4 km according to Core (2001).

Regarding the concentrations measured in van 2, it should be stressed that the high PM_{10} values acquired at the beginning of the experiments are related to some problems with the warming up of the equipment. The data measured by van 2 were affected by emissions from the plots where explosive hoses were used (514 and 515).

Table 5. Hourly averaged concentration values for permanent-fixed samplers (PFS)

PFS location	30 May 2002			31 May 2002		
	Exposure period	NO ₂ ($\mu\text{g m}^{-3}$)	SO ₂ ($\mu\text{g m}^{-3}$)	Exposure period	NO ₂ ($\mu\text{g m}^{-3}$)	SO ₂ ($\mu\text{g m}^{-3}$)
PFS1	9:30–17:47	40	10	8:40–16:03	30	7
PFS2	9:29–17:47	35	11	8:38–16:00	43	9
PFS3	9:20–17:45	38	15	8:35–15:59	31	9
PFS4	9:10–17:44	47	19	8:33–15:52	32	8
PFS5	9:12–17:42	50	26	8:30–15:50	31	7
PFS6	9:10–17:41	47	13	8:28–15:50	30	8
PFS7	9:07–17:39	53	18	8:25–15:51	30	10
PFS8	9:04–17:37	48	17	8:21–15:48	48	6
PFS9	8:59–17:34	46	15	8:19–15:40	22	–
PFS10	8:55–17:32	45	14	8:16–15:37	32	9

PM₁₀ concentrations acquired in van 2 during burning of plot 514 (Fig. 7b) were very high, reaching a maximum value of 3000 $\mu\text{g m}^{-3}$ and an average value of 940 $\mu\text{g m}^{-3}$ during the time the equipment was acquiring data (~6 h). Making the same assumption made for the calculation of the PM_{2.5} average, if it is considered that during the other 18 h of the day the concentration was 0 $\mu\text{g m}^{-3}$; the 24 h average is 235 $\mu\text{g m}^{-3}$. This value is extremely high when compared with the European legislation, which sets a maximum of 50 $\mu\text{g m}^{-3}$ for a 24 h average value. Also, the concentrations during the time of the experiment correspond to a hazardous category of air quality and very reduced visibility in that area, less than 1.4 km (Core 2001).

After this experiment, the filters had to be replaced and no data are available for the afternoon experiments (namely plot 515). Therefore, data presented in the Fig. 6 graphs, mainly zero values, are not reliable.

Still related to plot 514 are CO, NO and NO₂ values measured just after the end of the burning. CO concentrations are quite high, even exceeding the WHO (WHO 2000) and EPA (NAAQS) hourly limit values for this pollutant (30 and 40 mg m^{-3} , respectively) during approximately half an hour, a fact that raises some concern. However, the 8 h limit value established in European legislation was never exceeded. Wildland firefighters would probably be in contact with these high, and even higher, CO values. Measured NO₂ and NO concentration values attained peak values of 400 $\mu\text{g m}^{-3}$, but the hourly means, which attained a maximum of 189 $\mu\text{g m}^{-3}$, never exceeded the hourly European limit for NO₂. However, these levels could still be dangerous, mostly for firefighters working close to the fire front, because exposure to high NO₂ concentration values harms the lungs and increases respiratory infections (Frampton *et al.* 1991). The explosive hoses placed in plot 515 (Fig. 7c) actuated sooner than those of plot 514; consequently, CO, NO and NO₂ concentrations were not so high.

Because of the wind direction change during the afternoon, smoke from plot 518 was transported in the opposite direction from the measuring equipment, showing the extreme influence of meteorological conditions, particularly

unstable in mountainous environments, over the success of the work programmed.

Concerning NO₂ and SO₂ measurements with the passive sampling technique, Tables 5, 6 and 7 present the hourly averaged concentration values for PFS and TFS locations, and for MS. All passive NO₂ samplers measured values higher than the rural background concentration value 6 $\mu\text{g m}^{-3}$ (Penkett *et al.* 2003), indicating the influence of the smoke plume. The SO₂ concentration values are not significant and are considerably smaller than the European legislated value of 350 $\mu\text{g m}^{-3}$ (hourly average). Figure 8 schematically represents the spatial distribution of the measured NO₂ concentration values for the first day of the experiments.

The comparison of hourly concentration values measured by permanent and temporary fixed samplers allows verification of the effect of distance on the values measured. In fact, although PFS samplers were acquiring during a longer period, their greater distance from the burning area is reflected in lower values. The changing of diffusive tubes, aiming to understand the effect of explosive hoses on SO₂ and NO₂ values, allowed verification that plots where this technique was tested emitted less NO₂ and SO₂, probably because the fire was extinguished sooner.

During the first day of the experiments, higher SO₂ concentration values were acquired in TFS7 and TFS8. This can be related to the emissions resulting from the burn of two cars used for testing a new type of heat shelter. The intensity of the fire line was extremely high and, notwithstanding the successful performance of these innovative protective devices, some uncovered parts of the cars, especially tires, were seriously burned.

On the second day of experiments, NO₂ and SO₂ concentrations were generally lower than those acquired on the previous day for both PFS and TFS. Plots burned on the first day were closer to the passive samplers than those burned on the second day (Fig. 3).

The night period (from 6 p.m. 30 May to 9 a.m. 31 May) was not used to estimate firefighter exposure. Concerning the results of passive samplers used by firefighters and research team members, hourly concentrations of NO₂ and SO₂ were

Table 6. Hourly averaged concentration values for temporary-fixed samplers (TFS)

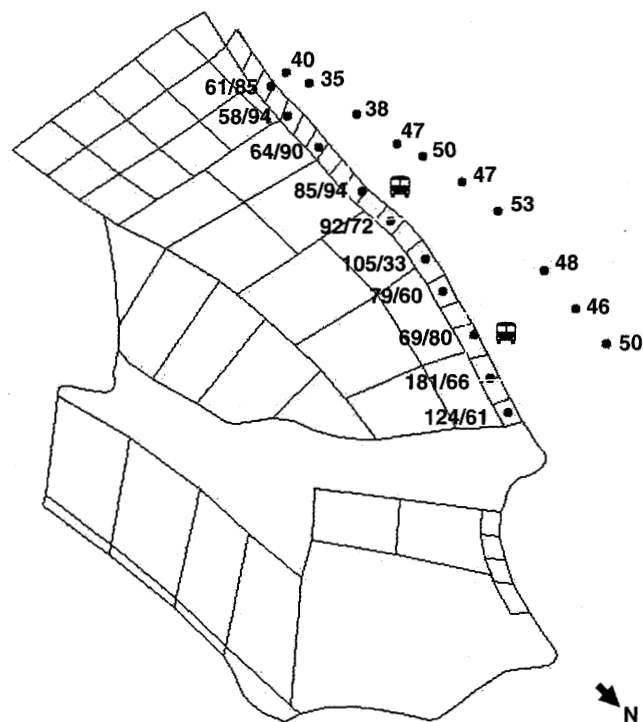
TFS location	30 May 2002			31 May 2002		
	Exposure period	NO ₂ ($\mu\text{g m}^{-3}$)	SO ₂ ($\mu\text{g m}^{-3}$)	Exposure period	NO ₂ ($\mu\text{g m}^{-3}$)	SO ₂ ($\mu\text{g m}^{-3}$)
TFS1	9:19–14:42	61	29	8:22–15:48	24	24
TFS2	9:14–14:40	58	26	8:25–15:47	24	14
TFS3	9:11–14:35	64	–	8:27–15:46	21	12
TFS4	9:09–12:40	85	–	8:30–15:43	24	13
TFS5	9:06–12:41	92	33	8:15–15:45	23	10
TFS6	9:03–12:44	105	36	8:19–15:40	0	10
TFS7	9:00–12:47	79	60	8:23–15:37	27	12
TFS8	8:56–14:40	69	56	8:21–15:38	29	22
TFS9	8:53–14:42	181	35	8:25–15:30	31	13
TFS10	8:50–14:38	124	17	8:36–15:35	35	13
TFS1	14:42–17:47	85	57			
TFS2	14:42–17:50	94	33			
TFS3	14:36–17:53	90	27			
TFS4	12:40–17:40	94	19			
TFS5	12:42–17:58	72	15			
TFS6	12:45–18:07	33	14			
TFS7	12:47–18:05	55	16			
TFS8	14:46–18:04	80	48			
TFS9	14:42–17:53	66	28			
TFS10	14:38–17:59	61	26			

Table 7. Hourly averaged concentration values for mobile samplers (MS)

MS	Exposure period	NO ₂	SO ₂
		($\mu\text{g m}^{-3}$)	($\mu\text{g m}^{-3}$)
Firefighter 1	9:51 (30)–17:30 (31)	38	5
Firefighter 2	9:46 (30)–15:10 (31)	52	6
Firefighter 3	9:46 (30)–15:10 (31)	50	7
Firefighter 4	9:44 (30)–15:10 (31)	48	9
Firefighter 5	9:40 (30)–17:30 (31)	30	4
Team member 1	10:17–15:10 (31)	67	12
Team member 2	10:17–15:10 (31)	57	21

higher in the team members' samplers (Table 7). One possible cause of this unexpected difference could be the fact that those were the team members involved in the smoke sampling with Tedlar bags (for VOC analysis) and, despite the much smaller exposure time, this procedure requires very close contact with smoke. The registered NO₂ values do not seem of concern if compared with the established hourly limit value for the protection of human health by European legislation ($260 \mu\text{g m}^{-3}$). SO₂ concentrations do not seem to be a problem at all when compared with the legislation; its hourly limit value is $350 \mu\text{g m}^{-3}$.

VOC samples were taken just close to the burning plots and the values obtained should be considered as emissions and not air quality values. Table 8 summarises the measured values. Separate VOC samples were taken during the flaming (first sample) and the smouldering (second sample) phases. With the exception of plots 514, 515 and 518, the values acquired indicate different emissions for those two stages, with smaller VOC concentrations emitted during the flaming

**Fig. 8.** Spatial distribution of hourly NO₂ concentrations (morning/afternoon values for temporary-fixed samplers and daily values for permanent-fixed samplers) for 30 May.

phase than in the smouldering stage (Table 8). Concentration values acquired for plots 514, 515 and 518 are very similar for both samplings. In addition, for plots 520 and 519 only one sampling was possible.

Table 8. Volatile organic compound (VOC) concentrations sampled

Day	Plot burned	Local time	VOC (mg Nm ⁻³)
30 May 2002	513	10:30	7.0
	513	10:35	17.7 ^A
	517	11:35	14.3
	517	11:45	27.3 ^A
	516	12:29	8.5
	516	12:30	19.3 ^A
	514	13:28	7.5
	514	13:34	5.4
	515	16:34	15.4
	515	16:36	13.3
	518	17:13	8.3
	518	17:15	6.5
	519	—	5.3
31 May 2002	522	11:32	17.7
	522	11:40	49.8 ^A
	521	12:33	9.5
	521	12:38	32.1 ^A
	520	14:04	22.5
	519	—	5.3

^A Smaller VOC concentrations emitted during the flaming phase than in the smouldering stage.

The Portuguese emission limit value for anthropogenic VOC is 50 mg Nm⁻³. Although the concentrations measured during the experiments did not exceed this limit, even for the smouldering emissions, they are quite significant.

Conclusions

A series of field experiments of fire spread in shrub vegetation on slopes have been performed in Portugal since 1998, in which the University of Aveiro has participated with emissions and air quality measurements. During the 2002 event, it was possible to distinguish the different contributions of flaming and smouldering stages on VOC emissions to the atmosphere. In fact, the concentrations emitted by the latter are more than two times higher, leading to the conclusion that the combustion stage has an intense effect on vegetation burning emissions. This conclusion is in agreement with the results obtained during fire experiments performed by other authors (Crutzen and Andreae 1990; Lobert and Warnatz 1993).

The maximum hourly averaged concentrations of NO₂ and SO₂ reached in the fixed samplers were 181 and 60 µg m⁻³ respectively. These values are less than those established by the European legislation (200 and 350 µg m⁻³ respectively). The maximum hourly concentrations found for these pollutants with the mobile samplers carried by operational people were 67 and 21 µg m⁻³ respectively. However, significantly higher values of NO₂ (of ~265 µg m⁻³) were registered during 2003 and 2004 fire experiments in Gestosa (Miranda *et al.* 2004), allowing the conclusion that, depending on the specific task of each firefighter involved in ground-based operations, the exposure to high levels of pollutants can be expected. A similar conclusion was reached from the

intensive measurements performed by the USDA Forest Service during wildfires and prescribed burns (Reinhardt and Ottmar 2000; Reinhardt *et al.* 2000).

Concerning data acquired by the automatic equipment installed inside the luggage vans, and despite the small size of the burning plots when compared to real wildfires, the measured levels of both categories of particles attained values considered hazardous to human health, 3000 µg m⁻³, 15 min average. Maximum PM_{2.5} and PM₁₀ hourly averaged measured values were, respectively, 2350 and 1430 µg m⁻³. For comparison purposes, the hourly averaged values measured in operational conditions during a wildfire in Greece reached 335 and 1300 µg m⁻³, respectively (NTUA 2004), indicating that in particular the PM_{2.5} value registered in the Gestosa-2002 fire experiments was very high. These measurements reinforce the idea that firefighters are exposed to unhealthy air pollutant concentrations during their daily activity. Regarding the recorded CO and NO₂ values it can be concluded that, in this specific situation, these pollutants did not attain disturbing concentrations. None of the pollutants surpassed the established limit values in the European legislation. In conclusion, it can be inferred that the most critical situation in terms of air quality that occurred during the experimental fires of Gestosa-2002 was posed by the emission of particulate matter.

Experimental field fires represent a valuable tool for understanding wildfires in all their aspects: how they behave, how they affect the environment or health, or how they can be extinguished efficiently. However, in relation specifically to the exposure and resulting adverse health effects of smoke on personnel involved in fire-fighting operations, more information is still needed and field experiments such as Gestosa should be encouraged.

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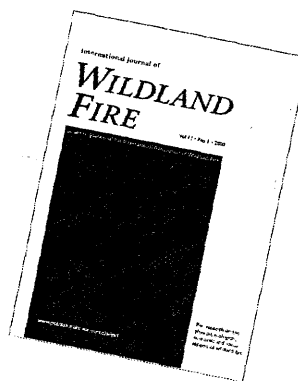
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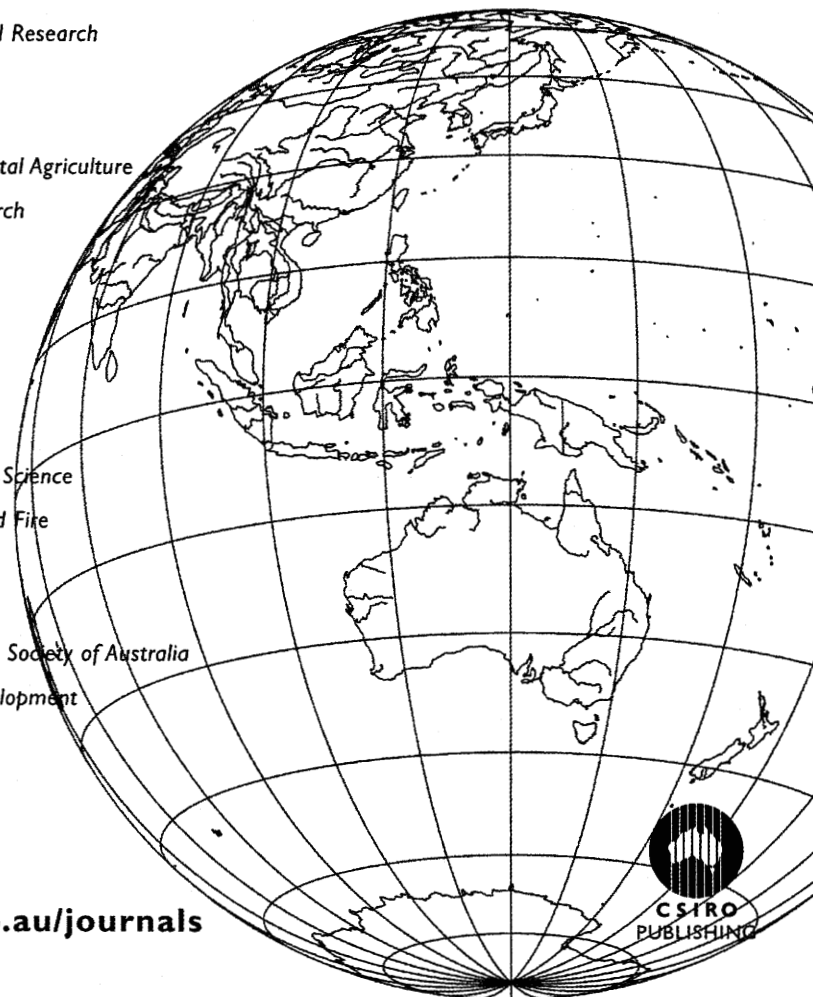
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Fire & Drought

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Australia Burning Fire Ecology, Policy & Management Issues

Edited by:
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David Lindenmayer
Stephen Dovers

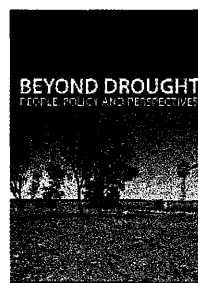
The phenomenon of fire in the Australian landscape traverses many interests and disciplines. At a national level, there is an urgent need for the integration of both the natural and social sciences in the formulation of public policy.

With contributions from 30 leading experts, the book draws together these issues, under the themes:

- Ecology and the environment
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Australia Burning examines key questions that relate to the ecology, prediction and management of fire, urban planning, law, insurance, and community issues, including indigenous and non-indigenous concerns. It looks at what we need to know to inform public policy and explores the avenues for closer integration between science, policy and the community.

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Beyond Drought People, Policy & Perspectives

Edited by:
Linda Courtenay Botterill
Melanie Fisher

Australia's unpredictable climate poses real challenges for practices that were developed on the relative predictability of a European climate. Recent policy has rejected the notion that drought is a natural disaster, favouring an approach based on risk management.

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NOV03

V International Conference on Forest Fire Research

27–30 November 2006, Coimbra, Portugal



First Announcement and Call for Papers

Objectives

The Conference aims to bring together scientists working on different aspects of forest fire, to encourage the presentation of latest advances in research, to promote discussion of methodologies and results, and to increase international cooperation. The format of the Conference will include formal sessions, with keynote lectures, oral presentation of papers, and poster sessions.

The V ICFFR will be held in Coimbra, which is well connected by road and train to Lisbon (200 km) and Oporto (100 km).

Subjects

Papers or posters dealing with topics related to forest fire on a scientific basis will be presented. The following list of subjects is given as a guide for the eventual organisation of the sessions:

**Fire prevention
Fire weather**

**Fire behaviour
Fire ecology**

**Fire management
Fire safety**

**Human and institutional factors
Urban–wildland interface**

Abstracts

Authors willing to submit a paper must send an Abstract written in English, of no less than 300 words, but not exceeding three A4 pages, including key figures and tables, before 31 December 2005. The abstracts will be refereed by the Scientific Committee of the Conference. Authors will be informed of the acceptance of papers by 31 March 2006. Detailed information about the format of the full version will be given at the same time.

Deadlines

Presentation of Abstracts	31 December 2005
Acceptance of Abstracts	31 March 2006
Full version of Papers or Posters	15 June 2006
Second Announcement	31 July 2006
Registration	30 September 2006
Hotel Reservation	30 September 2006
Definitive Program	31 October 2006

Post-Conference Activities

Several Post-Conference activities are planned

Conference Chairman:

Professor Domingos Xavier Viegas
[xavier.viegas@dem.uc.pt].

Conference Secretariat:

ADAI–Associação para o Desenvolvimento da Aerodinâmica Industrial
CEIF–Centro de Estudos sobre Incêndios Florestais
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'Fire and forest meteorology'

Edited by SA Ferguson, TJ Brown and M Flannigan

This issue is dedicated to David W. Goens

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