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

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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

$$\begin{aligned} &4C_6H_9O_4 + 25O_2 + [0.322M H_2O + 94.0N_2] \\ &\quad\rightarrow 18H_2O + 24CO_2 + [0.322M H_2O + 94.0N_2] \\ &\qquad + 11.6 \times 10^9\ \mathrm{J}, \end{aligned} \tag{1}$$

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 acknowledgement 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

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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, most 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 Reifsnyder 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:

$$\mathrm{CAPE} = g \int_0^{z_{\mathrm{EL}}} \frac{\theta_{\mathrm{V}}(z) - \overline{\theta}_{\mathrm{V}}(z)}{\overline{\theta}_{\mathrm{V}}(z)} dz. \tag{2}$$

Here g is the acceleration due to gravity; z_{EL} is the equilibrium height (level of neutral buoyancy) for surface air; $\theta_{\mathrm{V}}(z)$ is the virtual potential temperature of surface air after it has ascended to height z (with condensation if appropriate); and $\overline{\theta}_{\mathrm{V}}(z)$ 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 *et al.* (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 *et al.* (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 ΔCAPE values are in units of $\mathrm{J\,kg^{-1}}$

Fire	Day	CAPE	ΔCAPE					
			+2°C	$+2\,\mathrm{g\,kg^{-1}}$	+2/+2	+3°C	$+3\,\mathrm{g\,kg^{-1}}$	+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:

$$\mathrm{DCAPE} = g\int_{z_{\mathrm{sfc}}}^{z_0} \frac{\overline{\theta}_{\mathrm{v}}(z) - \theta_{\mathrm{v}}(z)}{\overline{\theta}_{\mathrm{v}}(z)}dz. \qquad (3)$$

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 DCAPE/Δq_v values are in J g^{-1}

Fire	Day	DCAPE	Δq_v	DCAPE/Δq_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
Dude	26 June 1986	1508	7.4	204
Red Lake #7	28 May 1986	822	9.7	85
	29 May 1986	1018	9.3	109
	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 (DCAPE/Δq_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 DCAPE/Δq_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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