When fires initiate or intensify towering thunderstorms, they can inject aerosols into the lower stratosphere that were once thought to originate only from volcanic plumes.
Wildfire, and its relation to weather, climate, and society, is a topic of increasing interest and attention. For instance, the Hayman Fire (Colorado) exploded from a human-caused ignition into a firestorm that burned 24,000 ha and advanced 31 km in its first 24 h (Graham 2003). Australia’s capital, Canberra, was overwhelmed by a lightning-started bushfire in January 2003 that brought death and wholesale destruction of property (Webb et al. 2004). San Diego, California, was under siege in October 2003 by the human-caused Cedar Fire, which consumed an area unprecedented in California history (U.S. Forest Service 2004). In 1988, 558,000 ha of the Greater Yellowstone Area were torched by wildfires that were historic in their intensity and community impact (Alexander 2009). Fires in Greece in 2007 and 2009 were major news events; in 2009, the government faced strong criticism for the recurrence of death and destruction after just 2 yr.

Global and regional warming trends have been identified and associated with exacerbated wildfire occurrence and impact (Stocks et al. 1998; Westerling et al. 2006). Attention to this topic has been heightened with growing concern regarding anthropogenic climate forcing and the apparent increase of fire in the wildland–urban interface. Superimposed on this important topic is a relatively new discovery. In 1998 a remarkable manifestation of extreme wildfire impact was identified: there was smoke in the stratosphere that was historic in their intensity and community impact (Alexander 2009). Fires in Greece in 2007 and 2009 were major news events; in 2009, the government faced strong criticism for the recurrence of death and destruction after just 2 yr.

Reports of confirmed pyroCbs and stratospheric impact are increasing in the scientific literature, but the entire body of published cases accounts for fewer than 10 events (Jost et al. 2004; Livesey et al. 2004; Fromm et al. 2006; Damoah et al. 2006; Lindsey and Fromm 2008; Cammas et al. 2009). However, since the advent of the “satellite era”2 in 1979, several stratospheric mystery-layer events have been reported (e.g., Bluth et al. 1997; Clancy 1986; Evans and Kerr 1983). Moreover, in the literature one can find other cases wherein stratospheric aerosol layers are attributed to volcanic eruptions when no clear evidence of such an event exists (Yue et al. 1994). Even the aftermath of a definitive stratospheric volcanic injection such as the 1991 eruption of Mount Pinatubo has involved aerosol patterns that investigators have had difficulty reconciling with expectations (e.g., Thomason 1992). Finally, the literature contains some reports of thin LS cloud layers inferred to be water–ice residue from overshooting convection (e.g., Neilsen et al. 2007) that have been challenged in terms of pyroCb-caused smoke (available online at www.atmos-chem-phys-discuss.net/6/9003/2006/acpd-6-9003-2006-discussion.html). Might the pyroCb, still in its infancy of understanding, be a contributor to some of these phenomena? Now that the pyroCb has been characterized, does the evidence of such mysterious or challenging stratospheric observations allow us to reinterpret earlier assessments? More generally,

1 “Overworld,” a term coined by J. Holton, is the range of stratospheric altitudes roughly greater than the 380-K potential temperature surface. This threshold generally defines the absolute top of the tropopause region anywhere on the globe.

2 The “satellite era” for our purposes is defined as beginning in 1979, when polar-orbiting weather satellites went into service with imaging and Earth radiation budget instruments, along with other instruments such as the National Aeronautics and Space Administration’s (NASA’s) Total Ozone Mapping Spectrometer (TOMS), and a host of solar occultation devices.
can satellite-era data be exploited to go beyond case studies toward a pyroCb climatology? If so, a broad new understanding of the scale of wildfire activity, its relation to weather, and interaction with climate change is within reach.

Here we identify three individual cases in which stratospheric pyroCb impact has been missed or misidentified. We employ nadir-viewing polar orbiter and geosynchronous satellite image data, satellite-based profile data, and ground-based lidar data in this pursuit. Using these resources we present evidence for a reinterpretation of selected stratospheric mystery-layer or volcanic aerosol reports in the literature. In addition, we present an in-depth characterization of the seasonal occurrence of wildfire, pyroCb, and the resulting smoke plumes in North America.

**PYROCB VERSUS VOLCANO.** The canonical model of LS aerosol is that the ultimate source/pathway for its material is the troposphere, and that material enters the LS by two primary irreversible mechanisms: slow cross-tropopause ascent in the tropics and rapid injection by volcanic eruptions (Thomason and Peter 2006). While there is still uncertainty and active research regarding these and other mechanisms (e.g., Khaykin et al. 2009; Dessler et al. 2007; Wang 2007), models of the lower and middle atmosphere do not take into account any other routine process for troposphere-to-stratosphere transport.

Aerosols, being a basic atmospheric constituent, are a fundamental tracer of polluting processes that affect both the troposphere and stratosphere. Regarding the stratosphere, observational and model analyses of aerosols are a basic means for understanding dynamics (e.g., Trepte and Hitchman 1992), patterns, and trends (e.g., Deshler 2008). Since the discovery by Junge et al. (1961) of a stratospheric “background” of liquid sulfate particles, temporal and spatial changes to this “layer” have been well documented with the aid of space- and ground-based profiling instruments (e.g., Jäger 2005; Deshler et al. 2006; Hofmann 1990; Hofmann et al. 2009; Thomason and Peter 2006). One seasonal/regional stratospheric aerosol peculiarity that has also been extensively studied is the polar stratospheric cloud (PSC). These form generally inside the winter polar vortex and are caused by adiabatic and diabatic cooling of air masses leading to condensation and/or freezing (e.g., McCormick et al. 1981; Browell et al. 1990; Toon et al. 1990).

Decadal studies of stratospheric aerosol loading generally conform to the above mentioned canonical model (Deshler 2008). However, one study reports departures of measured stratospheric aerosol burdens from modeled volcanic decay, with findings that indicate “several limitations in our knowledge of the volcano-atmosphere reactions...” (Bluth et al. 1997). Fromm et al. (2008a) reported that a single pyroCb injection in 2001 deposited into the LS an aerosol mass representing >5% of Northern Hemispheric LS background levels. Hence, it seems our understanding of the LS aerosol processes is far from complete.

**THREE MYSTERY CLOUD YEARS.** In the northern summers from 1989 through 1991, puzzling LS aerosol features were observed from ground and space. Sassen and Horel (1990, hereafter SH90) reported on perplexing lidar signals—depolarizing LS layers—at Salt Lake City, Utah, in August 1989. They concluded that the aerosols were volcanic in origin even though no confirmed LS volcanic injection occurred. In the summer of 1990 there was an impressive and sudden increase in LS aerosol loading in the northern middle and high latitudes, according to Yue et al. (1994). They analyzed an entire season of Stratospheric Aerosol and Gas Experiment (SAGE) II aerosol profiles, which chronicled a 4-month-long perturbation reaching an altitude of 17 km. Yue et al., in accordance with the canonical stratospheric model (and noting that every previous similar observation of SAGE II aerosol perturbation had been associated with a reported volcanic eruption), searched unsuccessfully for a documented volcanic eruption in 1990, and hence concluded that the mystery cloud was attributable to an unreported volcanic eruption in high northern latitudes. In June 1991 Mount Pinatubo’s cataclysmic eruption had a global, multiyear impact (e.g., Hansen et al. 1996). Although this event was thoroughly observed and modeled, a perplexing occurrence of early LS aerosol layers in northern middle and high latitudes formed a subtheme in papers on the resultant LS aerosol loading (e.g., Jäger 1992; Gobbi et al. 1992; Trepte and Hitchman 1992). Indeed, there were sufficient SAGE II observations for Thomason (1992) to characterize a “new mode” of “Pinatubo aerosols” just above the tropopause in northern extratropics. According to Thomason (1992), the new mode particle’s effective radius was between about 0.27 and 0.36 µm (inferred by SAGE II’s wavelength dependence of extinction), which was unique in the SAGE volcanic aerosol record and did not conform to expectations for volcanic sulfate droplets (on the order of 1 µm). Moreover, these new mode particles were observed in a systematically different altitude/latitude regime than the expected mode particles—in northern middle–high latitude and just above the tropopause.
**AEROSOL INDEX: THE UNKNOWN SMOKE SIGNAL.** Soon after the discovery of stratospheric smoke in 1998, a signal of the immediate effect of violent pyroCb explosions began to take shape. The day after a pyroCb was identified the absorbing aerosol index (AI) sensed by the Total Ozone Monitoring Spectrometer (TOMS) highlighted a smoke plume with peculiarly large AI values (e.g., Fromm et al. 2008a). AI is a positive number in the presence of absorbing aerosols, such as dust, smoke, and ash. AI is strongly dependent on plume aerosol optical depth (AOD) and plume altitude (Torres et al. 1998). At any given time on Earth there are optically opaque absorbing aerosol plumes. For example, in the burning season of Amazonia, perhaps the most familiar biomass-burning region, smoke plumes are often expansive and optically opaque. However, optically thick Amazonian smoke plumes have never had an AI > 12 in the TOMS satellite era (TOMS started operating in late 1978 and ended in 2005). In contrast, the “day after” pyroCb smoke plumes of events such as the Chisholm (Alberta, Canada) pyroCb of May 2001 (Fromm et al. 2008a) had AI > 29. In fact, some particularly extreme smoke plumes contain fill/error values in the level-3 (i.e., gridded) AI where the level-2 (i.e., the instrument’s native measurement footprint) AI manifests an even greater intensity. Table 1 shows the ranking of AI in the TOMS era. A listing such as Table 1 is an invaluable tool for investigating causality. Quite simply, it is a matter of looking at satellite image data and weather maps “upstream” one day for a phenomenon that might cause an optically thick, high-altitude smoke plume. Thirteen of the top 20 AI plumes are the results of smoke from documented or otherwise determined pyroCb events. The remaining events are also deep, thick, day-old smoke

<table>
<thead>
<tr>
<th>AI</th>
<th>Plume date</th>
<th>Lat (°) +N, −S</th>
<th>Lon (°) +E, −W</th>
<th>Cause</th>
<th>Source location</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>29.9</td>
<td>29 May 2001</td>
<td>65</td>
<td>−112</td>
<td>pyroCb</td>
<td>Alberta, Canada</td>
<td>Chisholm Fire; Fromm and Servranckx (2003)</td>
</tr>
<tr>
<td>25.9</td>
<td>19 Jan 2003</td>
<td>−32</td>
<td>163</td>
<td>pyroCb</td>
<td>Canberra, Australia</td>
<td>Pyrotornado; Cunningham and Reeder (2009)</td>
</tr>
<tr>
<td>25.3</td>
<td>5 Aug 1998</td>
<td>73</td>
<td>−64</td>
<td>pyroCb</td>
<td>Northwest Territories, Canada</td>
<td>Norman Wells pyroCb; Fromm et al. (2005)</td>
</tr>
<tr>
<td>18.8</td>
<td>18 Aug 2003</td>
<td>61</td>
<td>−89</td>
<td>pyroCb</td>
<td>Northwest Territories, Canada</td>
<td>Conibear Lake Fire; Wood Buffalo National Park</td>
</tr>
<tr>
<td>17.9</td>
<td>27 Aug 2000</td>
<td>42</td>
<td>−92</td>
<td>pyroCb</td>
<td>South Dakota</td>
<td>Jasper Fire; Black Hills National Forest</td>
</tr>
<tr>
<td>16.5</td>
<td>27 Sep 1998</td>
<td>69</td>
<td>148</td>
<td>TBD</td>
<td>Khabarovsk, Russia</td>
<td></td>
</tr>
<tr>
<td>16.2</td>
<td>18 Dec 2002</td>
<td>−35</td>
<td>144</td>
<td>pyroCb</td>
<td>Victoria, Australia</td>
<td>Big Desert Wilderness Park</td>
</tr>
<tr>
<td>15.9</td>
<td>21 Jun 1991</td>
<td>45</td>
<td>−24</td>
<td>pyroCb</td>
<td>Quebec, Canada</td>
<td>Baie-Comeau Fire (discussed herein)</td>
</tr>
<tr>
<td>15.6</td>
<td>4 May 2003</td>
<td>57</td>
<td>153</td>
<td>TBD</td>
<td>Eastern Russia</td>
<td></td>
</tr>
<tr>
<td>15.6</td>
<td>10 Jun 2002</td>
<td>45</td>
<td>−101</td>
<td>pyroCb</td>
<td>Colorado</td>
<td>Hayman Fire</td>
</tr>
<tr>
<td>15.4</td>
<td>10 Sep 1988</td>
<td>46</td>
<td>−89</td>
<td>pyroCb</td>
<td>Wyoming</td>
<td>Yellowstone National Park</td>
</tr>
<tr>
<td>14.9</td>
<td>7 Jul 1990</td>
<td>70</td>
<td>−152</td>
<td>pyroCb</td>
<td>Alaska</td>
<td>Circle Fire</td>
</tr>
<tr>
<td>14.9</td>
<td>8 May 1987</td>
<td>62</td>
<td>133</td>
<td>TBD</td>
<td>Northern Mongolia</td>
<td>Great China Fire; Cahoon et al. (1994)</td>
</tr>
<tr>
<td>14.4</td>
<td>23 Aug 1998</td>
<td>49</td>
<td>153</td>
<td>TBD</td>
<td>Khabarovsk, Russia</td>
<td></td>
</tr>
<tr>
<td>14.3</td>
<td>27 Jan 2003</td>
<td>−39</td>
<td>168</td>
<td>TBD</td>
<td>Southeastern Australia</td>
<td></td>
</tr>
<tr>
<td>14.3</td>
<td>20 Jun 2002</td>
<td>39</td>
<td>−104</td>
<td>pyroCb</td>
<td>Arizona</td>
<td>Rodeo-Chediski Fire</td>
</tr>
<tr>
<td>14.1</td>
<td>19 Jun 2002</td>
<td>42</td>
<td>−99</td>
<td>pyroCb</td>
<td>Colorado</td>
<td>Hayman Fire</td>
</tr>
<tr>
<td>14.0</td>
<td>6 May 2003</td>
<td>48</td>
<td>142</td>
<td>TBD</td>
<td>Eastern Russia</td>
<td></td>
</tr>
<tr>
<td>14.0</td>
<td>1 Feb 2003</td>
<td>−28</td>
<td>−178</td>
<td>TBD</td>
<td>Southeastern Australia</td>
<td></td>
</tr>
<tr>
<td>14.0</td>
<td>19 Aug 2000</td>
<td>48</td>
<td>−107</td>
<td>pyroCb</td>
<td>Idaho</td>
<td></td>
</tr>
</tbody>
</table>
plumes that have not yet been definitely associated with a specific source or event. Among these there are events in eastern Siberia wherein we suspect a substantial role is played by a vigorous extratropical cyclone spinning up in the flaming zone. This type of investigation, of these and other double-digit AI plumes, led us to a new interpretation of the 1989–91 mystery cloud events.

**Mystery Cloud Year 1: 1989.** The August 1989 LS aerosol layers at Salt Lake City reported by SH90 were shown in the context of meteorological analyses and parcel trajectories to be consistent with anticyclonic LS flow between the tropics and midlatitudes. It was in Guatemala that SH90 found a candidate volcanic eruption consistent in place and time with this LS flow regime. The suspected volcano was Santiaguito, which indeed erupted on 19 July 1989. However, it did not inject material near the stratosphere according to several scientists’ eyewitness reports (Smithsonian Institute 1989). We retrieved Geostationary Operational Environment Satellite (GOES) thermal infrared (THIR) imagery from the time of observed eruption ±5 h and found that the 11-µm THIR brightness temperature at the location of Santiaguito attained a minimum value of approximately −11°C, which, according to the closest radiosonde, implies a cloud top of no higher than 6 km. Moreover, Bluth et al. (1997) list no volcanic eruptions anywhere as having a volcanic explosivity index (VEI) indicating a stratospheric injection in the second half of 1989.

Fires in Manitoba and Saskatchewan, Canada, in historically great number were ignited by lightning on 17 July 1989 (Hirsch 1991). Four days later, on 21 July, extreme fire-weather conditions led to pyroconvection at a number of these fires, three of which spawned a pyroCb. GOES imagery (not shown) pinpointed these blowups. Advanced Very High Resolution Radiometer (AVHRR) imagery (Fig. 1a) captured the action in the late afternoon. At least four pulses of deep pyroCb anvils were in evidence. The day-after smoke plume on 22 July contained double-digit AI (Fig. 2a).

Figure 1c shows the smoke plume evolution in the first week after the pyroCb. Evidently the smoke pall is sufficiently high and massive that it can be followed with AI across the Atlantic Ocean to Europe, having a long-range persistence that is similar to other stratospheric smoke episodes (Fromm et al. 2005). We see also that part of the plume was transported south across the United States; the leading edge extended as far as Mexico on 23 July. Remnants of this portion of the plume circulated in the southern United States and Central America. Fortuitously, the smoke plume following this path was sampled by SAGE II on 25 July (Fig. 1b; positions of the SAGE II profiles are marked...
in Fig. 1c). Figure 1b shows a second SAGE extinction profile measured over the Atlantic on 31 July. Both aerosol profiles exhibit a strong layer at 14-km altitude and a wavelength dependence of extinction indicative of particles with radii of less than 1 µm. The back trajectories from both observations make excellent connections with the fire zone on 21–22 July; hence, we have an unambiguous confirmation of stratospheric smoke leading back to this pyroCb event in Canada.

Another fortuitous set of measurements of UTLS aerosols at that time was made in Manhattan, Kansas (39.2°N, 96.6°W), by ground-based lidar during the First International Satellite Land Surface Climatology Project (ISLSCP) Field Experiment (FIFE) Follow-On Project. The volume-imaging lidar (Eloranta and Forrest 1992) operated between late 26 July and 11 August (http://lidar.ssec.wisc.edu/pub_html/fife/vil/1989/index.htm). In relation to Topeka, Kansas, we surveyed the entire set of radiosonde temperature profiles and determined that the uppermost backscattering layers on 26 July, 31 July, and 6 August resided demonstrably above the local tropopause (not shown). We ran a back trajectory from the 26 July observation (Fig. 1c), and its path plus endpoint on 22 July are consistent with the plume transport across the United States and its origin in Manitoba, Canada.

A meteorological perspective for the AI, SAGE II, FIFE, and Salt Lake City observations is presented in Fig. 2. Figure 2a shows the day-after AI smoke plume in the context of 22 July geopotential height contours on the 375-K potential temperature surface (representative of the LS aerosol layers reported here and in SH90). The synoptic-scale LS flow regime straddling the smoke plume involves a trough in the middle United States and quasi-zonal flow eastward through Canada. The contour gradient, proportional to wind speed, is in agreement with the rapid southward transport of smoke across the United States into Central America. Figure 2b shows 3.5-day forward trajectories initialized at the time of the AI plume on 22 July, at 14 km, generally representing the various aerosol layers reported here and in SH90. [The Hybrid Single-Particle Lagrangian Integrated Trajectory (HYSPLIT) model (see Draxler and Rolph 2010; available online at http://ready.arl.noaa.gov/HYSPLIT.php) is used for trajectories in this paper except where noted otherwise.] The path portrayed by this matrix of particles shows the main features of the multiple paths of AI shown in Fig. 1c. Figure 2b also shows a 2-week back trajectory from the earliest SH90 layer at Salt Lake City. It traces a path back to Central America in the timeframe of the forward movement of the Manitoba smoke plume into that region shown in Figs. 1 and 2. Considering all of the meteorological and aerosol evidence presented here and in SH90 (including the depolarizing nature of the Salt Lake City scatterers), we hold that this reinterpretation of
the SH90 conclusions, in terms of LS smoke injected via pyroCb, is convincing. The sporadic measurements of LS aerosol layers provide a conservative hint to the broader stratospheric impact of the July 1989 Manitoba, Canada, pyroCb impact.

Mystery Cloud Year 2: 1990. According to Bluth et al. (1997), there were no volcanic eruptions with stratospheric-level VEI anywhere in 1990 except for Kelut (7.8°S), Indonesia, in February. However, discovery of a pyroCb in 1990 was afforded by the large AI day-after signal (Table 1). On 7 July 1990 AI = 14.9 was located over far northern Alaska. Figure 3a shows AVHRR imagery for that date and location, exhibiting the classic day-after pyroCb plume signature: an ashy gray cloud in visible bands, and very cold in THIR half-a-day-after injection (Lindsey and Fromm 2008). We then examined GOES visible, 3.9-µm (for hot spots), and THIR image loops and isolated a pyroCb generated by a fire called the Circle Fire, located at 65.9°N, 145°W, in the afternoon of 6 July. Figure 3c shows the AI evolution in the week after the pyroCb. The plume drifts north and east over very high Arctic latitudes and then spreads over eastern Canada, the Maritimes, and Greenland. Like the 1989 plume and other pyroCb events, this long-lived and long-transported AI signal represents abundant UTLS smoke aerosols.

Unlike the 1989 pyroCb event, there is no aerosol-layer measurement close enough in time to the pyroCb for trajectory-matching analysis. However, Yue et al. (1994) described a large-scale LS SAGE II aerosol perturbation at mid- and high northern latitudes in summer 1990 that, according to their Fig. 2, was still evident in October. We reanalyze the SAGE data in terms of LS AOD, defined as the integration of aerosol extinction from 2 to 6 km above the tropopause. Figure 3b shows zonal average AOD, calculated from a single-day complement of SAGE II profiles. In comparison with that from 1989, the 1990 AOD was identical before the pyroCb but nearly doubled afterward by early August. The plot of SAGE measurement latitude in Fig. 3b reveals that the strongest 1990 AOD enhancements were generally in the northernmost SAGE latitudes, indicating a high-latitude source. The anomalous 1990 zonal average AOD exhibits decay but was still evident into November, 4 months after the blowup. We conclude that the true source of this hemispheric LS aerosol increase was the Circle Fire pyroCb on 6 July, not a volcanic eruption. Moreover, a doubling of zonal average LS AOD is qualitatively equivalent to the perturbation caused by the Canberra, and Chisholm, Australian Capitol Territory, Australia, pyroCbs (Fromm et al. 2006, 2008b).

Mystery Cloud Year 3: 1991. Eighth on the list of greatest AI in Table 1 is a smoke plume on 21 June 1991. This plume was located over the Atlantic Ocean northwest of the Iberian Peninsula. One day prior there was also a large AI plume over Newfoundland, Canada. On 19 June there were two pyroCbs in Québec, Canada, spawned by separate fires. One of the fires (Fig. 4a) is about 100 km west of Baie Comeau, Québec, Canada, as evidenced by the largest/brightest hot-spot cluster; the pyroCb blew up...
after this image. A mature pyroCb with smoke-tinged anvil is present in Fig. 4a north of the Baie Comeau fire. Figure 4c shows the A1 evolution of smoke as the plume rapidly crossed the Atlantic and reached Russia within a week of the blowup. On 22 June the core of the A1 plume was situated over northern Europe near Denmark. On that day SAGE II made a measurement slightly east of Denmark (Fig. 4b) that contained a huge aerosol enhancement 2 km above the tropopause. Indeed, this SAGE measurement was the source of a high AOD feature on a global AOD map illustrating the 24 January 1992 cover of Geophysical Research Letters (1992, Vol. 21, No. 2), an issue that was partly dedicated to first Mount Pinatubo measurements. The back trajectory (Fig. 4c) from the 22 June SAGE II layer implicates the Québec pyroCbs, not those from Mount Pinatubo.

In addition to the SAGE II measurements, a number of lidar measurements in the weeks after the Mount Pinatubo eruption also detected LS aerosols that were difficult to reconcile with the volcano. Figure 5a shows that on 1 July 1991 lidars in Germany (Jäger 1992), France (Chazette et al. 1995), and Italy (Gobbi et al. 1992) all detected layers at 14–16 km. Back trajectories (Fig. 5b) from these layers all show a path to the 1 July observations from the northwest, across the North Atlantic Ocean, and crossing into North America (two in Canada), back 8–10 days (and within 1 week of the Mount Pinatubo eruption). Figure 6 shows a time series of the 313-nm backscatter coefficient recently calculated from measurements with the ozone lidar at Garmisch-Partenkirchen, Germany, on 1–3 July 1991 (Carnuth et al. 2002). These data reveal very high backscatter values in the lower stratosphere between 13 and 16 km during two specific periods, but much less in the evening of 1 July when the 532-nm measurement in Fig. 5 was made. The peak backscatter coefficient reached $8 \times 10^{-6} \text{ m}^{-1} \text{ sr}^{-1}$. The strong backscattering is indicative of a young (on the order of days old) and concentrated mass of aerosols. We calculated one hundred and eleven 315-h backward trajectories for this episode at intervals of 3 h, starting at altitudes between 13.5 and 16 km over Garmisch-Partenkirchen. Trajectories from the two relatively strong plumes closely pass over southeastern Québec (not shown). All of the trajectory paths can be generally characterized as westerly; endpoints (between 17 and 19 June) ranged from the western Atlantic Ocean through Central and North America to the eastern Pacific Ocean. The characteristic path of air reaching these three lidar sites is thus entirely inconsistent with the Mount Pinatubo plume, the direction of movement of which was westward from the eruption and constrained within 20° latitude of the equator (Bluth et al. 1992).

Thus, it appears that the pyroCb mechanism offers a reinterpretation for part of the widespread aerosol pollution of the northern LS in the summer of 1991, as well as the mystery clouds in 1989 and 1990. This reinterpretation has implications for how stratospheric aerosol processes and the effect of
extratropical convection on the UTLS are handled in transport, chemical, and climate models.

**HOW FREQUENT ARE PYROCBS?** The lesson of the prior discussion includes a realization that pyroCb occurrence is both greater than expected and a previously unknown contributor to historical smoke plume events. It is also reasonable to conclude that, like “regular” cumulonimbus, pyrocumulonimbus vary in intensity from the relatively rare, deepest stratospheric polluters to more frequent storms of lesser vertical extent. We explore these issues here, where we focus on one season, that of 2002, in North America. Much of southwestern United States experienced particularly intense drought in 2002 (Quiring and Goodrich 2008). During that season, a Canadian pyroCb was shown to be the source for in situ measurements of biomass-burning tracers in the LS (Jost et al. 2004). However, Jost et al. also came to the conclusion that deep pyroconvective activity was also likely to have occurred in the western United States that summer. Partly aided by the TOMS AI record, we surveyed the period of May–September 2002 for other UTLS smoke plumes and pyroconvection.

**FIRE SEASON 2002.** Figure 7 shows how daily AI extremes for a fixed geographic area vary with time. Interpreting the spikes as a possible signal of a particularly intense and high smoke plume, we identify candidate events to explore more deeply. Note that the spikes of interest need not be double-digit values of the historically greatest plumes of Table 1; any sharp day-to-day AI increase is a clue to a story worth exploring. It is of course also expected that some noteworthy plumes may be “hidden” among other more intense AI signals over an area as large as North America. Hence, Fig. 7 probably underestimates the number of events because one AI spike may be the result of more than one pyroCb. We investigated the AI spike events (AI > 5) by noting the date/coordinates of the plume, evaluating back trajectories from that location, examining GOES imagery “upstream” on the prior date, and searching fire databases to confirm...
fire location. For U.S. fires we used a compilation of Incident Status Summary (ICS-209) reports maintained by the U.S. Forest Service (C. McHugh 2009, personal communication). For Canada, we used the Large Fire Database (LFDB; Stocks et al. 2002). Pyrocumulus (pyroCu) convection is considered to have occurred if the short-wavelength infrared (SWIR; 3.9 \mu m) GOES imagery contains fire hot spots and if THIR imagery shows clouds, anchored to the hot spots, with colder-than-land brightness temperature (BT); “dry” smoke plumes are transparent to THIR radiation. The pyroCb subclass of pyroCu is indicated when the fire-anchored cloud pixels have BT < −40°C. The likelihood of pyroCb detection is increased by using the SWIR image of the fire-anchored cold (in THIR) cloud, which in daylight conditions will emit as an anomalously high BT (+10°C or more) owing to the peculiarly small particle size within smoky pyroCb anvils (Lindsey and Fromm 2008).

**PYROCONVECTION IN 2002.**

Table 2 gives a listing of the 2002 pyroCbs and “smoking gun” fires discovered by this method. Figure 8 is a map of fires > 200 ha, pyroCu, pyroCb, and the AI spikes highlighted in Fig. 7. The dates of the pyroCu and pyroCb events are annotated on Fig. 7, which shows that from 1 to 25 May, daily maximum AI was relatively low and invariant. AI in October was similarly invariant and small, consistent with light/declining wildfire activity. However, starting on 26 May the AI spike frequency increases strongly and remains the dominant feature through July. On 9 days between June and August, maximum AI reaches double-digit values. The first spike in May is attributable to a complex of fires and pyroconvection in eastern Alberta, Canada. Here the pyroCu cloud tops reached a (GOES) BT of −22°C, which according to the nearest radiosonde gives height and pressure of 5.9 km and 470 hPa, respectively. More pyroCu were detected in Alberta, Canada, on 31 May with upper-tropospheric cloud-top heights. Between 2 June and 28 July we identified 17 pyroCbs, 9 of which were in the 2-week period 18 June and 1 July. Noteworthy among these are the Hayman Fire in Colorado, which erupted into a pyroCb within 1 day of being ignited and a second time on 18 June, and the Rodeo–Chediski fire complex in Arizona. These were the two largest fires in the history of these two states and both were anthropogenic (Graham 2003; Ffolliott et al. 2008). On one occasion, 2 June, pyroconvection and two pyroCbs erupted from three separate fires along the Colorado/New Mexico border. One of these fires (named “unknown”) was not included in the U.S. Forest Service fire database. On four consecutive days between 18 and 21 June, pyroCbs exploded in Arizona, Colorado, and Alberta, Canada. On three consecutive days in mid-July, pyroCbs were found in Colorado and

![Fig. 7. Daily maximum NASA TOMS aerosol index over North America, May–Oct 2002. Isolated spikes with values > 5 are capped with a brown dot. Annotations for dates of pyroCu and pyroCb events.](image1)

![Fig. 8. Map of 2002 pyroCu (green diamonds), pyroCb (red-filled circles), and Canada/U.S. fires > 200 ha (white dots). Also plotted are locations of the AI spikes highlighted in Fig. 7.](image2)
Oregon. Two of these were generated by a single fire, the Burn Canyon Fire, roughly 24 h apart.

PyroCbs are obviously an extreme form of convection, yet their favored environmental conditions differ from those necessary for severe thunderstorms. Table 2 contains two stability measures for the 2002 pyroCbs: convective available potential energy (CAPE; see Bluestein 1993) and the Lower Atmospheric Severity Index (LASI) for wildland fires, better known as the Haines index (Haines 1988). There is no single CAPE threshold for severe convection; however, it is usually associated with values exceeding ~1000 J kg⁻¹, which typically implies a conditionally unstable lapse rate combined with abundant lower-tropospheric water vapor. In contrast, the Haines index (online at http://rammb.cira.colostate.edu/visit/fire/haines2.asp for details), which also includes a lapse-rate and moisture term,

| Table 2. PyroCbs in the United States and Canada 2002. |
|---------------------------------|--|--|---|---|---|---|---|---|
| **Name** | **Date** | **Lat** (*°N*) | **Lon** (*°W*) | **BTmin** (*°C*) | **Cloud-top z (km)/p (hPa)** | **LCL z (km)/p (hPa)** | **Haines index (J)/CAPE (kg)** | **RAOB site** |
| **Spring** (6,677) | 2 Jun | 37.0 | 105.0 | −43.0 | 10.4/267 | 5.2/544 | 6/583 | ABQ |
| **Unknown** | 2 Jun | 37.0 | 104.4 | −52.0 | 10.4/267 | 5.2/544 | 6/583 | ABQ |
| **Hayman** (55,749) | 9 Jun | 39.2 | 105.4 | −56.1 | 12.3/200 | 6.1/483 | 6/92 | DNR |
| **Hayman** (55,749) | 17 Jun | 39.2 | 105.4 | −56.2 | 11.6/222 | 4.9/561 | 6/918 | DNR |
| **Hayman** (55,749) | 18 Jun | 39.1 | 105.3 | −53.1 | 11.5/225 | 6.3/490 | 6/0 | DNR |
| **Million** (3,782) | 19 Jun | 37.7 | 106.7 | −58.0 | 12.2/200 | 5.5/519 | 6/287 | GJT |
| **Rodeo/Chediski** (189,651) | 20 Jun | 34.2 | 110.5 | −44.1 | 10.2/270 | 5.1/559 | 6/0 | FGZ |
| **Dobbin** (151,640) | 21 Jun | 56.7 | 104.5 | −58.1* | 11.8/207 | 2.5/760 | 6/135 | YQD |
| **Meadow** (75,483) | 24 Jun | 56.8 | 108.5 | −44.1 | 9.5/290 | 2.2/782 | 5/0 | YSM |
| **Lobb** (62,171) | 27 Jun | 55.3 | 103.3 | −58.0 | 12.4/187 | 2.5/762 | 5/0 | YQD |
| **Nagle** (71,029) | 27 Jun | 56.2 | 105.1 | −61.0 | 12.8/182 | 2.5/762 | 5/0 | YQD |
| **Unknown** | 27 Jun | 56.5 | 108.8 | −58.0* | 12.2/197 | 1.7/819 | 4/11 | YSM |
| **Mustang** (8,109) | 1 Jul | 41.0 | 109.3 | −60.0 | 13.0/184 | 4.1/623 | 6/18 | SLC |
| **Burn Canyon** (12,667) | 13 Jul | 38.0 | 108.4 | −53.1 | 11.9/216 | 6.0/494 | 6/768 | GJT |
| **Burn Canyon** (12,667) | 14 Jul | 38.0 | 108.4 | −53.1 | 12.6/193 | 5.5/532 | 6/420 | GJT |
| **Winter** (14,479) | 15 Jul | 42.8 | 120.8 | −43.1 | 10.7/258 | 3.6/672 | 6/0 | BOI |
| **Florence/Biscuit** (202,169) | 28 Jul | 42.3 | 123.9 | −50.2 | 11.6/232 | 2.4/770 | 6/0 | MFR |
| **Average** | | | | | 11.6/223 | 4.19/628.0 | | |

* BTmin < RAOB Tmin
signals extreme fire behavior only when an unstable lapse rate is matched with a dry lower troposphere, that is, a classic “inverted V” profile. In the case of the 17 pyroCbs in Table 2, CAPE values were indeed relatively small, with roughly half of the cases having a value of zero. However, the Haines index registered its maximum value of 6 (indicating conditions for high rate of fire spread) for all of the pyroCbs in the United States and one out of five in Canada. Of the remaining Canadian pyroCbs, all but one had a Haines index of 5.

In addition to a favorable Haines index, pyroCbs, like most cumulonimbus, also need a trigger for initiation. Sometimes the heat and moisture from the fire itself is sufficient for initiation, but occasionally the midlevel stability is too great for parcels to reach their level of free convection, and pyroCbs do not form. One trigger noted with fire blowups and pyroCbs is a cold front passage (e.g., Westphal and Toon 1991; Fromm et al. 2005). Luderer et al. (2006) modeled a documented pyroCb environment associated with a frontal passage and found that the modeled pyroconvection was substantially influenced by a cold front.

Perhaps fire size is an important metric for predicting pyroCbs. Table 2 lists the final fire size for the “smoking guns.” They were all large fires, but the final burned-area perimeter varies by two orders of magnitude. We did not have access to time-resolved fire-size change for Canada fires; this would be a critical value to have to associate fuel consumed on the days of pyroconvection versus the other days in the fire’s lifetime.

**Stratospheric Smoke in 2002.** In addition to the stratospheric impact Jost et al. (2004) reported from Canadian fires (on 27 June), there is strong evidence of stratospheric smoke from three additional pyroCbs: Hayman (9 June), a pyroCb ensemble between 18 and 20 June in Colorado–Arizona, and the Mustang pyroCb on 1 July (D. Knapp et al. 2009, personal communication). The evidence is from satellite- and ground-based aerosol profiles. For instance, on 21 June, the Purple Crow lidar (Sica et al. 1995) in London, Ontario, Canada, 42.9°N, 81.4°W, detected an aerosol layer between 11.6- and 14.5-km altitude, straddling the tropopause at 13.3 km (Fig. 9). An isentropic back trajectory passes over Colorado close to the Hayman fire on 18 June, the site of a second pyroCb from this fire (Table 2). This supports the contention of Jost et al. (2004) regarding additional occurrences of deep pyroconvection impacting the UTLS in 2002.

**PyroCb Injection Altitude.** It is abundantly evident, considering the published reports of stratospheric pollution via the pyroCb, that the effective maximum height of a pyroCb’s outflow is at or above the convective cloud-top altitude. A conventional method by which to infer cloud-top height (for optically opaque clouds such as thunderstorm anvils) is by way of cloud-top thermal infrared brightness temperature matched against the environmental lapse rate (Smith and Platt 1978). Even though this method entails uncertainty for clouds in the tropopause region resulting from potential nonsingularity in the temperature–height profile, it can still provide a robust, albeit conservative, value for outflow height. We employed this method for the events in Table 2. The average resulting pyroCb cloud-top altitude and pressure are 11.6 km and 223 hPa, respectively.

**Diurnal PyroCb Behavior.** Of all the factors that influence wildfire behavior, meteorology plays a big role. A strong feature of wildfire behavior is a diurnal cycle of alternation between nighttime relative quiescence and afternoon peak burning activity, driven by surface temperature, relative humidity, and wind speed. These fire–weather factors are basic inputs to the Fire Weather Index (FWI) component to the Canadian Forest Fire Danger Rating System.

**Fig. 9.** (a) Purple Crow lidar aerosol backscatter, 21 Jun 2002 and Buffalo radiosonde temperature profile, 0000 UTC 21 Jun. (b) Back trajectory superimposed on the Al map as in Fig. 1, with white dot showing lidar location. Back trajectory Z = 13 km; endpoint is 0000 UTC 19 Jun.
(Amiro et al. 2004, and references therein). It is therefore important to characterize a typical day in the life of a pyroCb, not only to understand the fire and firestorm behavior, but also to characterize the time change of emission height. This knowledge will inform fire behavior analysts, users of satellite data, and modelers. Since the discovery of the pyroCb, they have been observed by satellite to occur in morning, afternoon, and in middle-of-the-night hours. Even in 2002, among the 17 pyroCb events identified, one occurred at approximately 1100 LT (the Meadow Fire pyroCb on 24 June) and one occurred near local midnight (Burn Canyon, on 13 July). However, the preponderance of pyroCbs reached maturity in late afternoon, around 1800 LT (Table 2). We have analyzed all 17 in terms of local time, using GOES IR imagery to identify fire growth, pyrocumulus onset, and maturity. Here we centered each fire in a grid of GOES pixels approximately 48 km on a side and recorded certain metrics at each image time, for example, the maximum 11-µm BT (BTmax) and minimum 11-µm BT (BTmin). The BTmin metric is generated with respect to radiosonde-derived lifted condensation level (LCL) temperature. Negative values suggest pyrocloud formation; the more negative the value, the higher the pyrocloud. Fire hot-spot size change is tracked with 3.9-µm BT. A qualitative fire-size index is formed by counting hot-spot pixels and dividing by the maximum hot-spot count for that fire/pyroCb.

Figure 10 presents the average over all 17 pyroCb events. The fire-size metric shows that before local noon, fire size is negligible, consistent with the general diurnal behavior of tropical and subtropical vegetation fire (Giglio 2007). Toward midday, fire size increases and peaks in early afternoon. Undoubtedly this metric is impacted by cloud formation and is thus not solely influenced by fire behavior. However, in the mean it is apparent that these fires that erupted into pyroCbs spent the first half of the day being relatively inactive.

The BTmax trace, which likely represents clear-sky pixels, shows morning warming and a peak around 1300 LT. The BTmin curve generally follows BTmax until 1100 LT, when it peaks and begins a steep decline. This signifies the onset of pyroconvection wherein cloud formation in the flaming area begins to modify the diurnal clear-sky radiance progression. At roughly 1330 LT BTmin becomes negative, effectively indicating that an optically thick pyrocumulus cloud fills a GOES 4 km² pixel. Thus, at this point, the emissions from the fire may be assumed to reach as high as the LCL, which on average here is 4.1 km (632 hPa). From this point pyroconvection intensifies steadily (in the average sense) until a peak at roughly 1800 LT, when the pyroCb can be considered in full maturity. At this point the pyroCb is exhausting a considerable amount of biomass-burning emissions in the UTLS.

Thus, in the typical diurnal cycle of fire behavior that includes pyroCb energy, it can be expected that exhaust from this fire will span the troposphere in the course of a day. It is reasonable then to conclude that a considerable proportion of the emissions during the hours of deepest pyroconvection will be injected into the uppermost troposphere, above precipitation/scavenging processes. This is indeed a fundamental reinterpretation of fire vertical injection potential that is not well characterized in regional or global atmospheric models of chemistry and transport.

**SUMMATION.** Since the discovery of smoke in the stratosphere and the pyroCb, only a small number of individual case studies and modeling experiments (Trentmann et al. 2006; Luderer et al. 2006; Cunningham and Reeder 2009) have been performed. Hence, there is still much to be learned about the pyroCb and its importance. With this work we have attempted to reduce the unknowns by revealing several additional occasions when pyroCbs were either a significant or sole cause for the type of stratospheric pollution usually attributed to volcanic injections. Now it is established that pyroCb activity is sufficiently frequent that a measurable stratospheric increase in aerosols attributable to this process occurred in 1989–91, 1992 (Livesey et al. 2010).
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