A reconstruction of the meteorological condition leading to the deadly 1996 storm suggests promise for improving the predictability of mountain weather hazards and raises caution about the dangers of stratospheric ozone intrusions.

With a height of 8,848 m, Mount Everest is the highest mountain in the world. For over a century, it has been the subject of exploration for both scientific and recreational purposes (Venables 2003). This exploration began in earnest with the British expeditions during the early part of the twentieth century. At that time, Nepal was closed to foreigners and so these efforts were focussed on finding a route to the summit from the north through Tibet (Hemmleb et al. 2001; Venables 2003). During the 1921 expedition, a potential route to the summit was identified that is now referred to as the North Col-North Ridge route (Fig. 1). It was via this route during the 1924 expedition that Norton attained an altitude of 8,565 m without the use of supplemental oxygen—a record that stood for 54 years until Messner and Habler reached the summit in 1978 (Messner 1999; West 2000). Several days after Norton’s feat, ▶
Mallory and Irvine perished during a summit attempt made with the use of supplemental oxygen (Hemmleb et al. 2001; Venables 2003). To this day, it is still unclear if Mallory and Irvine were able to reach the summit before their deaths (Hemmleb et al. 2001; Messner and Carruthers 2001; Hemmleb et al. 2002). Exploration of Everest continued via this route until the beginning of World War II.

After the war, as a result of turmoil in Tibet and the opening of Nepal to foreigners, the focus of the exploration of Mount Everest shifted to the south and the discovery of a route that followed the large horseshoe-shaped cirque or valley known as the Western Cwm, bounded by Nuptse, Lhotse, and Everest to the South Col, and then ultimately to the summit (Venables 2003). This route, known as the South Col route (Fig. 1), ascends through the Khumbu Glacier and Kumbu Ice Fall into the Western Cwm, up and across the Lhotse Face to the South Col, and then along the south ridge to the summit of Everest.

Please refer to Fig. 2 for a three-dimensional representation of Mount Everest developed by Gruen and Murai (2002) that provides further information on this route. This route was the one used by Hillary and Norgay (Hillary 2003) in the first successful summit of Mount Everest in May 1953. Although other routes to the summit have been pioneered in the intervening years, these two routes remain the most popular (Huey and Salisbury 2003).

Much of the focus of scientific interest in Mount Everest has been on the impact that the low barometric pressure near its summit has on human physiology (Houston et al. 1987; West 1999; Huey and Eguskitza 2001). These studies have shown that above 7,000 m, climbers are at the limits of their endurance and are exposed to significant risks associated with the reduced amount of oxygen available for respiration: cold temperatures and high winds (West et al. 1983a; West 2000; Huey and Eguskitza 2001; Huey et al. 2001).

Perhaps the most dramatic example of these risks occurred in early May 1996 when a storm engulfed Mount Everest, trapping a number of climbers on its exposed upper slopes and ultimately resulting in eight deaths.¹ This event and the ensuing tragedy...
received international attention at the time (Burns 1996; Horsnell and Faux 1996) and became the subject of a number of books, including Into Thin Air (Krakauer 1999) and others (Boukreev and DeWalt 1999; Breashears 1999; Dickinson 1999), as well as figuring prominently in the movie “Everest” (MacGillvray and Breashears 1998).

On the evening of 9 May 1996, a large number of climbers were poised to make summit attempts from camp IV situated at 8,000 m on the South Col of Mount Everest (Fig. 2). High winds had persisted throughout the day and the possibility of summitting appeared low. The winds died down during the evening and the conventional wisdom was that they would remain calm for a period of time. As a result, the decision was made to attempt to summit and the climbers left around 2000 LT on the 9th for the 18–24-h round-trip to the summit. During the afternoon of 10 May, an intense storm with wind speeds estimated to be in excess of 30 m s⁻¹, heavy snowfall, and falling temperatures engulfed Mount Everest, trapping over 20 climbers on the exposed upper sections of the mountain above the South Col (Fig. 2). The unfavorable weather appears to have abated overnight, although winds remained high near the summit. There was however a reintensification during the day on the 11th. Throughout this period, a number of heroic attempts were undertaken to rescue the trapped climbers that were hampered by the high winds and the harsh weather. Tragically, five of the climbers could not be rescued and perished. By the afternoon of the 11th, most of the people on the upper slopes of Mount Everest were sheltered in tents at the South Col (Fig. 2). In this regard, the impact of the weather on the 11th was not as catastrophic as that on the 10th. Supplemental oxygen was however in short supply and there is evidence that even those that had taken shelter were suffering from the lack of oxygen that, as will be discussed in this paper, may have been exacerbated by the unusually low barometric pressure. A number of climbing parties who were attempting to summit along the North Col–North Ridge route (Fig. 1) were also trapped by the storm and three people died. Although the harsh weather abated after the 11th, winds near the summit remained high for next 5 days, hindering attempts to summit.

In Fig. 3, infrared imagery from the National Oceanic and Atmospheric Administration (NOAA) family of polar-orbiting satellites are presented for the late afternoon during the period of 9–12 May 1996. The images are derived from the Global Area Coverage (GAC) level-1B dataset (Kidwell 1998) and have a resolution of 4 km. Brightness temperatures derived from the observed radiances are presented with a color map in which colder (and therefore

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2 The following narrative is based on the published recollections of the storm as contained in Boukreev and DeWalt (1999), Dickinson (1999), and Krakauer (1999). All times are given in local Nepalese time, which is 5 h and 45 min ahead of coordinated universal time.

3 Among those on the upper slopes of Everest was a pilot who was familiar with the appearance of thunderstorms as seen from above. He is quoted in Krakauer (1999) as stating that he believed that the clouds that engulfed Mount Everest were associated with the crown of a “robust” thunderstorm. In two separate incidents during the storm, Krakauer (1999) reports that he and a sherpa observed thunder and lightning.
FIG. 3. Brightness temperatures (°C) derived from NOAA-12 infrared satellite images from (a) 1242 UTC (1827 LT) 9 May 1996, (b) 1221 UTC (1806 LT) 10 May 1996, (c) 1159 UTC (1744 LT) 11 May 1996, and (d) 1317 UTC (1922 LT) 12 May 1996. The location of Mount Everest is indicated by the +.

Figures 3a, 3b, and 3c illustrate the widespread and scattered convective activity over the Tibetan Plateau with India being, for the most part, cloud free. The distinct boundary between the cloud-covered and cloud-free areas lies along the Himalayas (Hirose and Nakamura 2005). The images from 10 and 11 May (Figs. 3b and 3c), the period of the Into Thin Air storm, are dramatically different with the presence of deep convection in the immediate vicinity of Mount Everest. The outbreak of convection on the 11th appears more organized, with the clouds tops higher than those on the 10th. Indeed, the convection in the vicinity of Mount Everest on the 11th has the characteristic circular cloud shield or anvil that is often associated with organized collections of thunderstorms known as mesoscale convective complexes (Laing and Fritsch 1993). On the 12th (Fig. 3d), one has a situation similar to the 9th, with widespread but scattered and shallow convection over the plateau with no evidence of deep convective activity in the vicinity of Mount Everest.

This outbreak of high-impact weather resulted in the largest number of fatalities (8) to occur near the summit of Everest during a single event. In the ensuing years, a passionate and wide-ranging debate has been ongoing in an attempt to understand the factors that contributed to this tragedy (Boukreev and DeWalt 1999; Breashears 1999; Dickinson 1999; Krakauser 1999; Roberto 2002; Mangione 2003; Kayes 2004). Curiously absent has been a quantitative discussion of the meteorological conditions that gave rise to this storm and the storm’s impact on the climbers’ physiology.

Motivated by the 1996 tragedy on Mount Everest, the Media Laboratory at the Massachusetts Institute of Technology (MIT) developed a portable sensor package to monitor weather conditions on Mount Everest (Lau 1998). One such portable weather station was deployed at the South Col of Mount Everest and operated from May to September 1998. Moore and Semple (2004) performed an analysis of these data and compared the observed temperatures and pressures to both the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis (Kalnay et al. 1996) and the European Centre for Medium-Range Forecasts (ECMWF) Re-Analyses (ERA-40) (Simmons and Gibson 2000). They found that the reanalyses were able to capture much of the day-to-day variability in the pressure and, to a lesser extent, the temperature at the South Col of Mount Everest. A high-impact weather event in early May, which resulted in a major impact to operations on Mount Everest with high winds and heavy snowfall reported (W. Berg 1998, personal communication), was captured in the South Col meteorological observations. Moore and Semple (2004) showed
that this event was associated with the passage of an upper-tropospheric jet streak, an elongated region of high wind speed embedded within the subtropical jet stream, near Mount Everest. Moore and Semple (2004) also showed that this event from 1998 was associated with a maximum in the time series of the Total Ozone Mapping Spectrophotometer (TOMS) total column ozone over Mount Everest and furthermore suggested that ozone-rich stratospheric air may have been present near the summit during this event. These results suggest that the summit of Mount Everest resides in the dynamically interesting but relatively unexplored region of the atmosphere known as the “middleworld” (Hoskins 1991), where the stratosphere and troposphere are in direct communication.

During the 1998 high-impact weather event, there was a relatively large discrepancy between the observed pressure at the South Col and the extracted values from the NCEP–NCAR and ERA-40 reanalyses. Through the analysis of satellite imagery of the event, Moore and Semple (2004) argued that this discrepancy was the result of an outbreak of convection in the vicinity of Mount Everest that was not resolved in the reanalyses. With regard to the initiation of this convection, they that argued it was triggered by ageostrophic circulation associated with the presence of the jet streak in the vicinity of Mount Everest (Uccellini and Johnson 1979; Keyser and Shapiro 1986).

From Moore and Semple’s (2004) analysis of the South Col observations and other work (Lang and Barros 2004; Moore 2004; Kennett and Toumi 2005), it is clear that the NCEP–NCAR and ERA-40 reanalyses are useful datasets with which to study the meteorology of the Himalayas. In addition, it appears that at least some of the high-impact weather that affects Mount Everest is associated with large-scale circulation patterns. In this paper, we will draw upon these results to investigate the weather associated with the May 1996 Into Thin Air tragedy and its impact on the people who were trapped near the summit of Mount Everest.

**FIG. 4.** Variability in the (a) temperature (°C), (b) wind speed (m s⁻¹), (c) pressure (mb), and (d) potential vorticity (PVU) at the summit of Mount Everest during May as extracted from the ERA-40 reanalysis. The solid black lines indicate the mean values over the period of 1958–2002. The dashed black lines delimit the band that contains 90% of all of these values during the period of 1958–2002. Therefore, values above/below the upper/lower black dashed lines occur in approximately 5% of the years. The red lines indicate the values during May 1996. These values are shown every 6 h with the * indicating the data at 0000 UTC each day.

**A SYNOPTIC-SCALE ANALYSIS OF THE INTO THIN AIR STORM.** We begin our analysis with Fig. 4 in which the time series for temperature, wind speed, pressure, and potential vorticity at the summit of Mount Everest during May 1996 were extracted from the ERA-40 reanalysis. All time series were obtained by interpolation of the corresponding three-dimensional fields to the location and height of Mount Everest (27°59′N, 86°55′E; 8,850 m). Also shown are climatological mean time series and estimates of the interannual variability based on the time period of the ERA-40 reanalysis (1958–2002). The first three fields are obvious ones that describe the general meteorological conditions at the summit. Potential vorticity is in a different class because it is, for example, not directly observable. It is however a
relatively simple and widely used diagnostic that allows one to distinguish between air of tropospheric and stratospheric origins (Danielsen 1984; Hoskins 1991; Morgan and Nielsen-Gammon 1998), with values in the range of 1–2 potential vorticity units ($1 \text{ PVU} = 10^{-6} \text{ K s}^{-1} \text{ kg}^{-1}$) usually serving as the cut-off. However, as discussed by Hoskins (1991), the exact value used as a cut-off is not as important as the jump in potential vorticity that exists across the tropopause.

As discussed in Moore and Semple (2004), there will be variability in these fields that is not resolved in the ERA-40 reanalysis as a result of its coarse resolution. These time series should therefore be interpreted as being representative of the synoptic-scale variability in these fields. With regard to the climatology, we observe that a warming, a decrease in wind speed, an increase in pressure, and a decrease in potential vorticity occurred during the month of May. As described by Moore and Semple (2004), these changes are associated with a warming of the troposphere and the transition to the oncoming Indian summer monsoon. All fields are characterized by a decrease in variability during May. With regard to the temperature and pressure, the outliers contain the same basic trend as does the mean. In contrast, the trend in variability of the wind speed and potential vorticity appears to be the result of a reduction in the occurrence of anomalously high wind speed/potential vorticity events. We believe this to be also a consequence of the oncoming monsoon that results in a transition from a regime impacted by the passage of baroclinic weather systems to a more convective regime.

Turning our attention to the events in May 1996, bearing in mind that the ERA-40 reanalysis captures only the synoptic-scale variability, we see that the summit temperature underwent an approximate $10^\circ\text{C}$ drop from the 8th to the 13th. This transition was associated with a change from anomalously warm to anomalously cold temperatures. Temperatures remained low until approximately 25 May when they returned to values seen in early May. The summit wind speed time series indicates the presence of several distinct maxima, followed by minima that occur with a period of approximately 1 week. One of these maxima occurred on the 9th when wind speeds at the summit were approximately $22 \text{ m s}^{-1}$. The wind speeds decreased over the next 2

![Fig. 5. Spatial structure of the climatological mean (a) geopotential height (contours, km), potential vorticity (shading, PVU), and (b) potential vorticity (contours, PVU) and horizontal wind (vectors, m s$^{-1}$) and wind speed (shading, m s$^{-1}$) fields on the 340-K potential temperature surface during May. Climatology based on the ERA-40 reanalysis for the period of 1958–2002. The 4,000-m height contour is indicated with the dashed line, while the + indicates the location of Mount Everest.](image-url)
days to approximately 15 m s⁻¹ before increasing to values in excess of 25 m s⁻¹ on the 12th. Wind speeds on the 12th exceeded those observed in 95% of the years in the reanalysis record. The summit pressure was anomalously high in early May and then decreased over the next 8 days, reaching anomalously low values on 12 and 13 May. The pressure remained low for the next 10 days before returning to anomalously high values around 25 May. With regard to potential vorticity, we see that a transition from tropospheric, that is, < 1 PVU, to stratospheric, that is, > 1 PVU, values occurred between 8 and 10 May. The potential vorticity remained high for approximately 3 days before returning to lower values. There were two additional events later in May when high values of potential vorticity occurred.

We now consider the spatial representation of the synoptic-scale flow during the Into Thin Air storm. Motion in the upper troposphere, where the summit of Mount Everest is located, is quasi adiabatic and occurs, to first order, along surfaces of constant potential temperature (Hoskins et al. 1985). Therefore, to visualize the large-scale circulation during this event, we have chosen to view the flow on the 340-K potential temperature surface.

Figure 5 shows the climatology on this surface during May for the period of 1958–2002. With regard to the climatological mean height of this surface (Fig. 5a), we see that Mount Everest is situated in the region of large meridional gradient as one moves northward from the Tropics toward the midlatitudes. The height of this surface in the vicinity of Mount Everest during May is approximately 9.25 km. From the climatological mean potential vorticity on this surface (Figs. 5a and 5b), it is clear that the summit of Mount Everest is typically in the troposphere with the tropopause just to the north over central Tibet. This confirms that Mount Everest lies in the middleworld (Hoskins 1991). From the climatological mean wind speeds on this surface (Fig. 5b), one can see that the region of high wind speeds known as the subtropical jet stream is typically situated just to the north of Mount Everest during May. The region of low wind speeds in this jet stream over Tibet was first noted by Krishnamurti (1961).

In Fig. 6, we present the climatological mean vertically integrated moisture transport field and its magnitude during May. This field, defined as the vertical integral of the product of the specific humidity and the horizontal wind, provides a succinct view of the atmospheric water balance, with convergence of the field on monthly and longer time scales being proportional to the excess of precipitation over evaporation (Peixoto and Oort 1992; Smirnov and Moore 1999). This figure shows that during May there are two main moisture streams in the region—the dominant stream that transports moisture in a northeasterly direction over the Bay of Bengal toward southeast Asia, and a weaker stream that transports moisture eastward from the Arabian Sea over northern India south of the Himalayas. There appears to be some convergence effects associated with the high topography of Tibet and the merging of the two streams over northeast India and Bangladesh, which receives the heaviest rainfall on the subcontinent during May (Parthasarathy et al. 1994).

We present in Figs. 7 and 8 the spatial distribution of the potential vorticity, height, and wind speed in the 340-K potential temperature surface before, during, and after the event. At 0000 UTC 7 May (Figs. 7a and 8a), the height of this surface in the vicinity of Mount Everest was approximately 8.5 km, which is lower than the climatological value of approximately 9.25 km (Fig. 5a). This reflects the higher pressures that
were occurring near the summit (Fig. 4c). At this time, Mount Everest was situated in the troposphere with potential vorticities in its vicinity less than 1 PVU. The wind speed field (Fig. 8a) clearly shows the presence of a jet streak just to the north of Mount Everest with its entrance region to the west of Mount Everest.

During the storm, 0000 UTC (0545 LT) 11 May (Figs. 7b and 8b), the height of the 340-K surface in the vicinity of Mount Everest was approximately 10 km, reflecting the lower pressures that were present near the summit at this time (Fig. 4c). This lower pressure was associated with an upper-level shortwave trough that was situated just to the west of Mount Everest at this time (Schultz and Doswell 1999). The source of the elevated high potential vorticity identified in Fig. 4d can be seen from Figs. 7b and 8b to be a filamentary intrusion of high potential vorticity air that extended from the midlatitudes toward the Mount Everest region. Such intrusions are favored in northwesterly flow behind shortwave troughs (Rotunno et al. 1994; Schultz and Doswell 1999). Evidence of this intrusion and the associated shortwave trough can be seen, albeit in a weaker form, over the Arabian Sea on the 7th (Figs. 7a and 8a). Two jet streaks were in the vicinity of Mount Everest on 11 May (Fig. 8b). In addition to the jet streak present on 7 May (Fig. 7a), another jet streak was situated to the west of Mount Everest on the 11th. As a result of its eastward propagation, the entrance region of the jet streak identified in Fig. 7a was now situated to the northeast of Mount Everest, while the exit region of the second jet streak was to the southwest.

Over the next several days, the shortwave trough and the filamentary intrusion of high potential air moved to the east and continued to intensify into a well-defined tropopause fold by the 15th (Figs. 7c and 8c). On this date, Mount Everest was again situated in the troposphere.

**Fig. 7.** Spatial structure of the geopotential height (contours, km), the potential vorticity (shading, PVU), and the horizontal wind (vectors, m s⁻¹) fields on the 340-K potential temperature surface at 0000 UTC on (a) 7 May, (b) 11 May, and (c) 15 May 1996. All fields are from the ERA-40 reanalysis. The + indicates the location of Mount Everest.
with the 340-K surface being at a height of 9.5 km, close to the climatological value of 9.25 km. The second jet streak that was situated to the west of Mount Everest on the 11th was now situated to its north, resulting in the high winds that we observed near the summit in the aftermath of the storm (Fig. 4b).

The synoptic situation presented in Figs. 7 and 8 provides an explanation for the observations of the people who witnessed strong winds on 9 May, (associated with the eastern jet streak) and a weakening of these winds late on the 9th as this streak propagated away from Everest, and then a reintensification of the high winds a few days later as the second jet streak passed by. The recollections of those near the summit during the event clearly indicate the presence of high winds in addition to other severe weather elements. This is inconsistent with the picture that we have presented of the synoptic-scale conditions during the storm. Satellite imagery during the event (Figs. 3b and 3c) clearly shows the outbreak of deep convection during the event. In the following section, we show that this convective component to the event was forced by the synoptic-scale flow in the region.

**CONVECTION IN THE VICINITY OF MOUNT EVEREST DURING THE INTO THIN AIR STORM.** Convection is often observed to occur in conjunction with jet streaks (Uccellini and Johnson 1979; Keyser and Shapiro 1986; Doswell and Bosart 2001). The section of a jet streak where the air parcels are decelerating is referred to as the exit region, while the section where the air parcels are accelerating is referred to as the entrance region (Keyser and Shapiro 1986). Convective activity, which is the result of an ageostrophic circulation that develops in the plane perpendicular to the jet streak, preferentially occurs to the left of the exit region and to the right of the entrance region in a coordinate system in which the observer is looking...
in the direction that the wind is blowing (Uccellini and Johnson 1979; Keyser and Shapiro 1986). Indeed, the satellite imagery (Figs. 3b and 3c) clearly shows that an outbreak of deep convection near Mount Everest occurred during this event. In this section, we will argue that this convection was the result of the juxtaposition of the two jet streaks, enhanced by the presence of the shortwave trough and the large flux of water vapor in the vicinity of Mount Everest. We will focus our attention on 0000 UTC (0545 LT in Nepal) 11 May, a time that represents the approximate middle point in the outbreak of harsh weather.

To verify the presence of ageostrophic circulation, we use the Helmholtz theorem (Arken 1985) to partition the horizontal wind field $V$ into rotational $V_r$ and divergent $V_d$ components,

$$V = V_r + V_d = \hat{k} \times \nabla \psi + \nabla \chi,$$

where $\psi$ is the streamfunction, $\chi$ is the velocity potential, and $\hat{k}$ is a unit vector in the vertical direction. In this formulation, the jet streak–induced motion is contained in the divergent wind field $V_d$ with maxima/minima in the velocity potential associated with regions of convergence/divergence. This partition was performed with the SPHEREPACK software package (Adams and Swarztrauber 1999).

In Fig. 9, we present the velocity potential $\chi$ and the divergent wind field $V_d$ on the 300-mb pressure surface at 0000 UTC 11 May 1996 as depicted in the ERA-40 reanalysis. At that time, there is evidence of two distinct circulations emanating from a region of divergence that was situated just to the east of Mount Everest. One circulation had a region of convergence to the west of Mount Everest, while the other had a region of convergence to the northeast. The positions of the jet streaks at this time (Fig. 8b) show that the former straddles the entrance region of the jet streak to southwest, while the latter straddles the exit region of the jet streak to the northeast. The effect of curvature associated with the presence of a shortwave trough is to enhance divergence to the east of the trough and divergence to the west (Schultz and Doswell 1999), and thus would have acted to enhance the ageostrophic circulation associated with the approaching jet streak.

Advecting moisture into the region where there is forced ascent is also a requirement for an outbreak of severe convection, preferably in a direction perpendicular to the axes of the jet streaks (Uccellini and Johnson 1979). In Fig. 10, we present the vertically integrated moisture transport and its magnitude as well as the total precipitation at 0000 UTC 11 May 1996 as depicted in the ERA-40 reanalysis. It should be noted that the precipitation in the ERA-40 reanalysis is relatively unconstrained by observations and so is highly dependent on the details of the underlying numerical weather prediction model, resulting in a tendency for the ERA-40 precipitation to have higher values than those that are observed (May 2004; Andersson et al. 2005). Nevertheless, this bias does not appear to be a problem with respect to the Indian summer monsoon and Himalayan rainfall (Challinor et al. 2005; Kennett and Toumi 2005).

On this date, the vertically integrated moisture transport field has a structure similar to the climatological mean for May (Fig. 6), but with a magnitude in the region of interest that is approximately twice as large. Examination of this field on earlier days confirms that the maximum to the southeast of Mount Everest is the result of the merging of two streams—one from the Bay of Bengal and the other from the Arabian Sea (not shown). There is also evidence of a channelling of the moisture flux along the southern edge of the plateau. The total precipitation field has a maximum of 18 mm day$^{-1}$ in the region;
the convective component of this precipitation was relatively small at less than 4 mm day\(^{-1}\) (not shown). The magnitudes of the vertically integrated moisture flux and total precipitation in the region to the southeast of Mount Everest during this outbreak of harsh weather were exceeded only 5% of the time during May for the ERA-40 time period of 1958–2002.

More information on the spatial structure of the atmospheric moisture stream to the southeast of Mount Everest is provided in Fig. 11. Here we show the zonal and meridional components of the water vapor transport field, defined as the product of the specific humidity and the horizontal velocity field (Peixoto and Oort 1992) along 92.5°E as depicted in the ERA-40 reanalysis. The moisture in this stream is confined to the lower 4 km of the atmosphere with only a slight spillover onto the Tibetan Plateau, and with the northward transport being approximately 60% larger than the eastward transport.

In Fig. 12, we show the zonal wind and vertical velocities, as well as the potential vorticity along a meridional cross section through Mount Everest (87°E) at 0000 UTC 11 May 1996, as depicted in the ERA-40 reanalysis. The high zonal wind speeds to the south of Mount Everest were associated with the exit region of the oncoming jet streak (Fig. 8b). The intrusion of high potential vorticity, that is, stratospheric, air associated with this event can be seen to be associated with the presence of a tropopause fold (Keyser and Shapiro 1986) in the vicinity of Mount Everest. With regard to the vertical velocity field, there exists a dipole on this day with a region of rising motion in the vicinity of Mount Everest and a region of descending motion over the Tibetan Plateau to the north. This dipole is consistent with idealized representations of the transverse circulation that is associated with the presence of the jet streak to the northeast of Mount Everest (Uccellini and Johnson 1979; Keyser and Shapiro 1986; Doswell and Bosart 2001), as well as with the divergent circulation illustrated in Fig. 9.

The results presented in this section are consistent with the hypothesis that the convection responsible for the Into Thin Air storm was the result of synoptic-scale forcing. Finally, we note that there is a pronounced diurnal cycle to the convection over the Tibetan Plateau with the most intense activity occurring in the late afternoon and early evening (Laing and Fritsch 1993; Yanai and Li 1994). This is the time of day on both the 10th and 11th when the high-impact weather on Mount Everest was most intense (Boukreev and DeWalt 1999; Krakauer 1999). This diurnal cycle may have also contributed to the high-impact weather.
PHYSIOLOGICAL EFFECTS OF HIGH-IMPACT WEATHER ON MOUNT EVEREST. It has been well established that the hypoxia caused by the low barometric pressures at extreme altitudes may severely limit human performance. Observations from the summit of Mount Everest indicate an inspired partial pressure of $O_2$ ($P_{O_2}$) of 57 mb compared with a sea level value of 198 mb (West et al. 1983b). Furthermore, above 7,000 m extreme hyperventilation is required to provide even these low values of $P_{O_2}$ (West et al. 1983b). This amount of oxygen is barely sufficient for the requirements to maintain even the basal metabolic rate of a human being (West 2000). This has lead to many physiological studies that have concentrated on the ability of acclimatized mountaineers to undergo extreme exertion while in this severe hypoxic state (West 1984; Houston et al. 1987; West 2000). One of the conclusions reached was that the available oxygen levels at these extreme altitudes were exquisitely sensitive to even small variations in barometric pressure. Furthermore, in the physiological literature, the relationship between barometric pressure and altitude has been assumed, apart from seasonal variations, to be relatively constant (West et al. 1983b; Houston et al. 1991; West 1999; Huey and Eguskitza 2001; Huey et al. 2001).

From the South Col observations of the May 1998 storm discussed by Moore and Semple (2004) as well as the results presented in this paper,

**Fig. 12.** Cross section of the vertical velocity (shading, cm s$^{-1}$) and the zonal wind speed (solid contours, m s$^{-1}$) and potential vorticity (dashed contours, PVU) though Mount Everest along 87°E at 0000 UTC 11 May 1996. All fields are from the ERA-40 reanalysis. Please note that the topography along the cross section does not resolve the full height of Mount Everest.
it is clear that there exist pressure fluctuations that are associated with the onset of high-impact weather in the Mount Everest region. From Fig. 4, one can see that the *Into Thin Air* storm was associated with an approximate 8-mb drop in the barometric pressure at the summit of Mount Everest in the ERA-40 reanalysis. This drop in pressure was most likely associated with the presence of the shortwave trough. In a comparison between the ERA-40 and NCEP–NCAR reanalyses and observations at the South Col during the May 1998 storm, Moore and Semple (2004) found an approximate 40% underestimation in the magnitude of the pressure drop in the reanalyses. This error was attributed to the fact that the reanalyses did not capture the pressure fluctuations associated with the convective component of the high-impact weather. A similar result is probably true for this event as well. The summit of Mount Everest is approximately 850 m above the South Col site. Based on the variation of pressure with height in both the NCEP–NCAR and ERA-40 reanalyses, we estimate that during the 1998 event, there was an additional 20%–30% increase in the magnitude of the anomaly at the summit. Assuming similar conditions, we estimate that climbers trapped above the South Col would have experienced a total pressure drop between 10 and 14 mb during the *Into Thin Air* storm.

Such a drop in barometric pressure would result in an approximate 6% reduction in inspired partial PO$_2$ near the summit (West et al. 1983b; West 1999). Nonlinearities in the relationship between PO$_2$ and maximum oxygen uptake ($V_{O_{2}max}$) result in an approximate 14% reduction in the latter for climbers near the summit (West 1999). To put these values into context, similar drops in barometric pressure, PO$_2$, and $V_{O_{2}max}$ are estimated to occur between summer and winter near the summit of Mount Everest (West et al. 1983b; West 1999). These reductions are of physiological significance, and according to some investigators are probably why Everest has never been climbed during midwinter without supplemental oxygen (West 1999). In view of these findings even strong, experienced climbers who were caught near the summit of Everest in a storm of this magnitude would be at a physiological disadvantage. For those climbers without supplemental oxygen or those who perhaps were only partially acclimatized after running out of supplemental oxygen, the physiological impact of the barometric pressure drop associated with such a storm would have been most profound.

In addition, we have presented evidence as to the possible presence of ozone-rich stratospheric air on the upper slopes of Everest during the *Into Thin Air* storm. A similar conclusion was reached by Moore and Semple (2004). Ozone is recognized as posing a risk to human health through the impairment of lung function that comes through the exposure to elevated levels thereof (Devlin et al. 1996; Environmental Protection Agency 1996). In this regard, it is interesting to note that exposure to elevated ozone concentrations at moderate altitudes (2434 m) in the Austrian Alps has been postulated to result in a decrease in pulmonary function in mountain guides (Wittels et al. 1997). Furthermore, ozone of stratospheric origin associated with tropopause folds has been measured in commercial aircraft for over 40 yr at concentrations that can exceed the air quality guidelines (National Research Council 2002; Spengler and Wilson 2003; Spengler et al. 2004). Indeed, concentrations of ozone in tropopause folds of the type that were associated with the *Into Thin Air* storm are in the range of 120–280 ppb (Viezee et al. 1983; Davies and Schuepbach 1994; Ravetta and Ancellet 2000; Roelofs et al. 2003). At these concentrations, which exceed guidelines established by the U.S. Environmental Protection Agency and other agencies, it seems reasonable to suggest that the presence of ozone near the summit of Mount Everest represents a hitherto unrecognized risk to climber’s health.

**CONCLUSIONS.** In this paper, we have investigated the meteorological conditions associated with the *Into Thin Air* storm of May 1996. This outbreak of high-impact weather is by all accounts the deadliest one to occur on Mount Everest, with eight deaths on its upper slopes attributed to it. Our analysis builds on the results of Moore and Semple (2004) who used observations collected at the South Col of Mount Everest during 1998 to show that the global reanalyses of NCEP–NCAR and the ECMWF can capture the synoptic-scale variability in the pressure and temperature near the summit of Mount Everest.

Our results indicate that the high-impact weather during the *Into Thin Air* storm was the result of an outbreak of deep convection in the vicinity of Mount Everest (Fig. 3). The outbreak occurred during a transition from anomalously warm high pressure to anomalously cold low pressure conditions during a period of low synoptic-scale wind speed that was bracketed by periods of high wind speed (Fig. 4). In addition, the event occurred during a period of high, that is, stratospheric, values of potential vorticity at the summit (Fig. 4). Detailed information on the meteorological conditions before, during, and after this event are unfortunately unavailable. However, the narratives of Krakauer (1999) and Boukreev and...
DeWalt (1999) both mention the high winds that were present near the summit on 9 May and the subsequent weakening of the winds just prior to the outbreak of the harsh weather. Dickenson (1999), who was attempting to summit via the North Col–North Ridge route and as a result was in a more exposed position vis-à-vis the westerly wind, comments on the sustained high winds that occurred after the outbreak, which rendered it impossible for him to summit until they weakened later in the month.

We have also shown that there were two jet streaks and an upper-level shortwave trough present in the vicinity of Mount Everest during this event (Figs. 7 and 8). The positioning of these two streaks and the trough was such that the transverse ageostrophic circulation associated with them resulted in a synoptic-scale rising motion near Mount Everest (Figs. 8 and 9). In addition, there was in the lower troposphere an anomalously large transport of water vapor from both the Arabian Sea and the Bay of Bengal into the region to the south of Mount Everest that was oriented approximately perpendicular to the jet streaks (Figs. 10 and 11). We believe that it was the juxtaposition of the ageostrophic circulation associated with these upper-level features near Mount Everest that in combination with the anomalous availability of moisture triggered the organized convective activity (Fig. 12). The harsh weather occurred in the late afternoon, and it is likely that this synoptically forced rising motion was enhanced by that associated with the diurnal cycle of solar heating.

Our results, along with those of Moore and Semple (2004), suggest that it may be possible to use the reanalysis datasets and the models upon which they are based to study and forecast high-impact weather systems in the Mount Everest region, especially in the premonsoon period. This is of great practical benefit because this is the period during which much of the activity on the mountain takes place (Huey and Salisbury 2003). As an example, the summit pressure time series (Fig. 4c) indicates that the 10 May 1996 summit attempt took place during a period in which the pressure was falling to values that were significantly lower than those usually observed. This signal, which is indicative of the large-scale circulation, rather than local wind speed, which may be influenced by local topographic effects as well as misleading as the result of the presence of jet streaks, should have provided a warning as to the possibility of high-impact weather. The analysis of wind speeds as well as the divergent circulation on constant pressure or potential temperature surfaces may provide evidence of the presence of jet streaks, shortwave troughs, and their associated ageostrophic circulations that could assist in determining the potential for harsh weather.

There exists evidence that fatalities on Everest typically occur during descent and that the probability of death is significantly higher for those not using supplemental oxygen (Huey and Eguskitza 2000; Huey et al. 2001). Operational constraints dictate that descents from the summit of Mount Everest typically occur in the late afternoon or early evening (Breashears 1999; Krakauer 1999). As we have discussed, this is the period of the day when the probability of convective activity is highest over the adjoining plateau (Laing and Fritsch 1993; Yanai and Li 1994). The higher probability of convective activity in the late afternoon when climbers are descending and near exhaustion is another risk factor that has not been previously identified.

It is therefore likely that the falling pressure and the high-impact weather that ensues may play a role in the fatalities that occur on Mount Everest. This is especially true above 7,000 m, where climbers are at the limits of their endurance and drops in barometric pressure of the magnitude that we have established, when compounded by the accumulative effect of hypoxia, fatigue, high winds, extreme cold, and incomplete acclimatization, could shift a coping climber from a state of brittle tolerance to physiological distress. The presence of ozone-rich stratospheric air near the summit of Mount Everest during the event may have provided an additional stress on the climbers through a reduction in pulmonary function.

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