SUBTROPICAL COLD SURGES: REGIONAL ASPECTS AND GLOBAL DISTRIBUTION

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Received 25 October 2000
Revised 19 April 2001
Accepted 23 April 2001

ABSTRACT

Shallow surges of cold dry air are frequently observed to the east of the major mountain ranges, moving from mid-latitudes well into the Tropics in about 4 days. Because of their strong impact on weather, regional aspects of cold surges have received considerable attention, particularly over Southeast Asia, to the east of the Rockies and Mexican Sierras, and to the east of the subtropical Andes. Both observational and numerical studies reveal a similar structure and evolution of cold surges in different regions. These common aspects are reviewed in this work, as well as the mechanisms responsible for the development and subsequent advance of cold surges over the subtropics. Atmospheric reanalysis data are used to document the global distribution of cold surges on the basis of their continental-scale imprints on relevant fields, as well as to estimate their contribution on the regional and global energy balances. It is found that cold surges have a major cooling and drying effect over the regions where they are prevalent (e.g. subtropical South America) and represent a sizeable sink of energy for the Tropics. Copyright © 2001 Royal Meteorological Society.

KEY WORDS: cold surges; global distribution; NCEP–NCAR reanalysis; transient heat transport

1. INTRODUCTION

Episodic incursions of cold mid-latitude air that penetrate deep into the Tropics are frequently observed to the east of major north–south oriented mountain ranges (Figure 1). The so-called ‘cold surges’ are confined to the lower troposphere, exhibit horizontal scales in the range 500–1000 km, and can be traced over periods from 2 days to a week. Although their effects are dependent on the geographical area of occurrence, the passage of a cold surge is typically characterized by a rapid and marked decrease of air temperature and dew point at low levels, a sharp increase in surface pressure, and a day or so of moderate to strong equatorward low-level winds. Depending on the thermodynamic characteristics of the pre-incursion environment, cold surges may also have a substantial impact on convection and rainfall. Cold surges are thus one of the most energetic influences on the tropical circulation by the extratropics (e.g., Riehl, 1954; Ramage, 1971; Hastenrath, 1982).

In a broad sense, cold surges reaching low latitudes owe their existence to the interaction of the synoptic-scale flow with the earth’s topography. Mountain–no mountain numerical experiments of cold surges over Southeast Asia (Tilley, 1990), North America (Hartjenstein and Bleck, 1991) and South America (Knight D, 1998, personal communication) have resulted in similar conclusions: in the absence of the relevant mountain range, cold surges do not propagate equatorward, but rather the simulated cold outbreaks progress eastward with the mid-level wave. Early studies interpreted cold surges as rotationally trapped Rossby or Kelvin waves (e.g. Fritsch and Webster, 1985; Sumi, 1985; Hsu, 1987; Tilley, 1990), mainly because of their phase speed (direction and magnitude) and the exponential decay of the wind...
anomalies away from the mountain barrier. In the framework of rotationally trapped waves, vertical motion produces the temperature and vorticity tendencies at the leading edge of the cold air through adiabatic cooling and vortex shrinking, respectively. Nevertheless, the weak upslope flow and the horizontally limited uplifting ahead of the cold air seem incapable of producing the rapid and marked cooling and anticyclonic tendency observed at the edge of the surges. Furthermore, observational and modelling studies of cold surges over Southeast Asia (Sumi, 1985), to the east of the Rockies (Colle and Mass, 1995) and to the east of the Andes (Garreaud, 1999) have shown that the intense low-level cooling is largely dominated by meridional advection, consistent with the similarity between the propagation speed of the surge and the along-barrier wind speed within the cold outbreak. Other processes (such as topographically trapped waves, evaporative cooling, etc.) may be superimposed upon the advection of cold air and explain the observed case-to-case variability and regional dependence of the cold surge evolution.

Because of the strong impact of cold surges on weather, literature focusing on their synoptic aspects is extensive, while fewer studies have documented their influence on planetary-scale phenomena (e.g. Chang and Lau, 1981; Lau and Chang, 1987; Iskenderian and Salstein, 1998; Slingo, 1998). Although regional analyses are essential in understanding the structure and dynamics of this phenomenon, it might be useful to place cold surges in a global context by recognizing the common aspects between regions through cross-regional comparison. In this short contribution the aim is toward a global perspective, which will be accomplished by reviewing regional studies (Section 2) and documenting the signatures of cold surges on the global distribution of relevant atmospheric fields (Section 3). This objective survey of cold surges is based on 17 years (1979–1995) of daily average reanalysis fields produced by the National Center for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) (Kalnay et al., 1996). As described in Kalnay et al. (1996), the 6-h reanalysis fields were produced using a frozen weather prediction model and an enhanced observational database, which includes conventional, aircraft and satellite data, and hence they represent one of the most complete, physically consistent datasets of the atmospheric circulation, suitable for synoptic-scale studies. In Section 4, we also take advantage of this dataset to evaluate the energy fluxes associated with cold surges and document their aggregated effect on the tropospheric heat balance. Section 5 summarizes our main findings.
Synoptic-scale cold surges have been mostly described for regions such as Southeast Asia (bordered by the Himalayan Plateau), North and Central America (east of the Rockies and Mexican Sierras), and South America (to the east of the Andes cordillera) (see Figure 1). Shallow surges of cold air are also observed along the coastal escarpment of southeastern Australia (e.g. Baines, 1980; McBride and McInnes, 1993), the east-side of the Appalachians (e.g. Bell and Bosart, 1988) and over the Greek peninsula (Lagouvardos et al., 1998), but their scales are sub-synoptic and as such will not be discussed further here.

Originally, the term ‘cold surge’ was used to describe the rapid drop of air temperature over Southeast Asia and the adjacent South China Sea produced by shallow outbreaks of cold, continental air during the Northern Hemisphere winter (e.g. Ramage, 1971). The surges move southward swiftly, from mid-latitudes (~ 40°N) into the Tropics (as far south as 20°N) in about 4 days, causing rapid weather changes across much of Southeast Asia. The synoptic-scale precursors and structural evolution of east Asian cold surges are summarized by Chang et al. (1979), Boyle and Chen (1987), and more recently by Wu and Chan (1995, 1997), while their dynamics have been addressed in modelling studies, among others, by Lim and Chang (1981, 1987), Sumi (1985) and Tilley (1990). Several studies have also emphasized the impact of these surges on the convective activity over the maritime continent to the south of east Asia (e.g. Chang et al., 1979; Chang and Lau, 1981; Boyle and Chen, 1987) and upon the onset of the Australian monsoon (Williams, 1981; Davidson et al., 1983). Furthermore, outflow from enhanced convection reinforces the upper branch of the Hadley cell and gives rise to the acceleration and westward extension of the subtropical Asian jet. In turn, the intensified jet leads to planetary scale interactions as documented by Chang and Lau (1981), Lau and Chang (1987) and Slingo (1998).

Northerly surges of cold air bounded by the eastern slope of the Rockies have also attracted considerable attention because of their impact over the Great Plains of North America and coastal Mexico from fall to spring (e.g. Murakami and Ho, 1981; Bluestein, 1993). Intense episodes during winter can produce a drop of low-level air temperature as large as 30°C within 24 h and northerly winds exceeding 20 m s⁻¹. Synoptic and sub-synoptic features of North American cold surges are documented, among others, by Dallavalle and Bosart (1975), Hartjenstein and Bleck (1991), Mecikalski and Tilley (1992), Colle and Mass (1995) and Konrad (1996). Under favourable large-scale circulation, described in Schultz et al. (1998), wintertime cold surges originating over extratropical North America propagate deep into the Tropics bounded by the eastern slope of the Mexican Sierras (e.g. Henry, 1979; Hastenrath, 1982; Reding, 1992; Schultz et al., 1997). Central American cold surges can produce substantial air temperature drops and strong, gusty winds over eastern Mexico and Central America (e.g. Schultz et al., 1997), significant cold-season rainfall (Bals-Elsholz and Bosart, 1997) and a decrease of sea-surface temperature over the Caribbean due to wind-driven mixing.

In South America, episodic incursions of cold air into tropical and subtropical latitudes are a distinctive year-round feature of the synoptic climatology of the region to the east of the Andes (e.g. Kousky and Cavalcanti, 1997; Krishnamurti et al., 1999; Vera and Vigliarolo, 1999; Garreaud, 2000). Extreme wintertime episodes produce freezing conditions from central Argentina to southern Brazil, the chief agricultural sector of the continent, which has motivated case studies by Hamilton and Tarifa (1978), Fortune and Kousky (1983), Marengo et al. (1997), Bosart et al. (1998) and Garreaud (1999). These studies, among others, have shown that cold incursions to the east of the Andes have a structure and evolution similar to cold surges in the previous regions. Summertime episodes produce less dramatic fluctuations in temperature and pressure, owing to the smaller seasonal temperature gradient between mid and low latitudes, but they are accompanied by synoptic-scale bands of deep convection at the leading edge of the cool air (Parmenter, 1976; Ratisbona, 1976; Kousky, 1979; Garreaud and Wallace, 1998; Liebmann et al., 1999) that account for up to 40% of the summertime precipitation over subtropical South America (Garreaud and Wallace, 1998) and can reach as far as the northern coast of the continent (e.g. Kiladis and Weickmann, 1997).

Despite case-to-case variability and geographical dependence, regional descriptions, as summarized above, reveal that cold surges share important aspects of their evolution and structure, schematized in...
Figure 2. First, their origin is invariably associated with the passage of an extratropical disturbance in the middle-troposphere, building up a cold air pool and a poleward large-scale pressure gradient to the east of the mountain range. The subsequent equatorward advance of the cold air occurs in the form of a shallow \((H \sim 2\text{ km})\) dome of cold air with a sharp temperature gradient along its leading edge, moving parallel to the contours of surface elevation, and accompanied by a hydrostatically induced ridge of surface pressure. As the surge moves into low latitudes, strong surface heat-fluxes weaken the cold air anomalies, and the surge may lose its ‘cold’ character. However, strong meridional winds and low dew points remain as clear signatures of the surge. Surges tend to enhance subtropical and tropical deep convection because of the intense low-level convergence along their leading edge.

To assess the blocking effect of the topography upon the low-level flow, Table I includes the Rossby number \((Ro = \frac{U}{f l_m})\), Froude number \((Fr = \frac{U}{N h_m})\) and the Rossby radius of deformation \((l_r = \frac{N h_m}{f})\),

![Conceptual model of a cold surge moving from mid-latitudes into the subtropics along a north–south oriented mountain range in the Northern Hemisphere. The leading edge of the surge is indicated by the surface cold front. The thin curves represent surface isobars; H and L indicate the position of the surface anticyclone and trough, respectively. The dashed lines indicate the position and phase of the mid-level wave. See text for details](Image)

Table I. Relevant topographic and dynamic parameters in three regions of common occurrence of cold surges

<table>
<thead>
<tr>
<th>Parameter</th>
<th>East Asia</th>
<th>North and Central America</th>
<th>South America</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Himalayas 35°N</td>
<td>Himalayas 25°N</td>
<td>Rockies 35°N</td>
</tr>
<tr>
<td>(l_m) (km)</td>
<td>800</td>
<td>500</td>
<td>750</td>
</tr>
<tr>
<td>(h_m) (m)</td>
<td>4000</td>
<td>1300</td>
<td>2000</td>
</tr>
<tr>
<td>(N) (10^{-2}) (s(^{-1}))</td>
<td>2.0</td>
<td>1.7</td>
<td>2.0</td>
</tr>
<tr>
<td>(f) (10^{-5}) (s(^{-1}))</td>
<td>8.3</td>
<td>6.14</td>
<td>8.3</td>
</tr>
<tr>
<td>(Ro)</td>
<td>0.18</td>
<td>0.33</td>
<td>0.21</td>
</tr>
<tr>
<td>(Fr)</td>
<td>0.17</td>
<td>0.54</td>
<td>0.30</td>
</tr>
<tr>
<td>(Ro/Fr)</td>
<td>1.05</td>
<td>0.61</td>
<td>0.70</td>
</tr>
<tr>
<td>(l_r) (km)</td>
<td>900</td>
<td>400</td>
<td>500</td>
</tr>
</tbody>
</table>
based on typical values of the low-level horizontal velocity \((U)\), the mountain height and half width \((h_m \text{ and } l_m)\), the lower troposphere Brunt Väisälä frequency \((N)\), and the Coriolis parameter \((f)\). The Froude number is the ratio of kinetic energy to the potential energy required to pass over the mountain, so that \(Fr \ll 1\) implies topographic blocking of the low-level circulation over a region that extends to about one Rossby radius of deformation \((l)\) from the barrier (Pierrehumbert and Wyman, 1985). Along the subtropical portion of the Andes, the Rockies–Mexican Sierras and Himalayas, the small values of \(Fr\) and large values of \(Ro\) (Table I) indicate that the cold air is totally blocked by the eastern slope of the mountain barriers and that the dynamics of the surges substantially depart from geostrophy in the three regions considered. Further poleward over North America and East Asia, \(Ro < 1\) suggests that the dynamics of the surge lies within the quasi-geostrophic regime, although the cold air is still unable to pass over the mountains because \(Fr < 1\). For the three regions considered in Table I, \(l \sim 500–900\) km, which is in agreement with the documented scale (normal to the barrier) of the cold air dome and the region of downgradient, terrain-parallel low-level flow during cold surges.

3. GLOBAL DISTRIBUTION

From the previous sections it is clear that synoptic-scale cold surges have a pronounced influence on subtropical weather. Therefore, the temporal variance of relevant meteorological variables, over regions where cold surges occur frequently, should be large relative to the rest of the subtropics. To document this aspect, we have used several statistics derived from the daily mean NCEP–NCAR reanalysis fields. The base period for these calculations is 1979–1995. Reanalysis data has proved suitable for describing synoptic-scale phenomena, especially over the continents, where the relatively large number of observations reduces potential errors from the model itself or errors that may propagate from oceanic, data-void areas. For instance, cold surges over subtropical South America selected using NCEP–NCAR reanalysis data correspond well with cold surges identified using surface observations over southeast Brazil (Garreaud, 2000). Nevertheless, quantitative results over regions adjacent to steep topography must be taken with caution, since the coarse resolution of the reanalysis data do not resolve strong, mesoscale circulations. In our analysis, the time series at each grid point were band-pass filtered to retain fluctuations in the \(3–12\) day range, using a ninth-order Butterworth filter. Although there is no clear spectral separation at \(12\) days, the filter segregates the relevant high frequency variability (periodicity of about a week or less) from intraseasonal fluctuations.

Of particular interest for our global survey of cold surges is the standard deviation \((\sigma_T)\) of the air temperature at the 925-hPa level for the boreal and austral winter. Figure 3 shows the zonal mean of \(\sigma_T\) for each semester \((\langle \sigma_T \rangle)\). The zonal mean is at a maximum in mid-latitudes, in connection with the prevalent baroclinic disturbances, and decreases toward the equator, as the synoptic scale perturbations become less intense and frequent. Next, we subtracted the zonal mean from the standard deviation to obtain the global distribution of \(\sigma_T^* = \sigma_T - \langle \sigma_T \rangle\), shown in Figure 4. To aid in the identification of those regions where cold surges account for most of the variance, we have superimposed the skewness coefficient \((\gamma_T)\) field of the 925-hPa temperature, defined as (Panofsky and Brier, 1958):

\[
\gamma_T = \left[ \frac{n \Sigma (T - \bar{T})^3}{(n - 1)(n - 2) \sigma_T^3} \right]
\]

where \(n\) is the number of temporal samples. \(\gamma = 0\) for symmetric distributions. The histogram of temperature in a grid box where \(\gamma_T < 0\) (negatively skewed distribution) exhibits a long tail that extends far to the left, so that extreme cold departures are larger or more frequent than warm departures. The opposite is valid for positively skewed distributions \((\gamma_T > 0)\).

Consistent with the argument at the beginning of this section, \(\sigma_T\) is substantially larger than the zonal mean \((\sigma_T^* \geq 1.5^\circ C)\) to the east of the subtropical Andes, the Rockies and Himalayan Plateau (for absolute values of \(\sigma_T\) see Figure 3). The common occurrence of cold surges in these regions is also signalled by significant negative values of \(\gamma_T\). The clearest case is found over South America during the austral winter.
Figure 3. (a) Long-term zonal mean of the standard deviation of the air temperature at 925 hPa level for the austral winter (JJA, solid line) and boreal winter (DJF, dashed line). The filled symbols indicate the standard deviation at 25° of latitude at 60°W (to the east of the Andes, circle), 85°W (to the east of the Rockies, square), and 110°E (Southeast Asia, triangle) during their respective winter season. The vertical lines indicate the 95% confidence level of the sample standard deviation. (b) As in (a) but for the standard deviation of the meridional wind at 925 hPa. (c) As in (a) but for standard deviation of SLP.
Figure 4. Global distribution of the standard deviation ($\sigma^*$) of 925-hPa air temperature (contours) and skewness coefficient (shaded) for (a) June–August, and (b) December–February. $\sigma^*$ is the departure of $\theta$ from the seasonal zonal mean (shown in Figure 3(a)). Only positive values are shown, at an interval of 0.5°C. The temperature was previously band-pass filtered to retain fluctuations in the 3–12 day range. Dark (light) shading indicates skewness coefficient less than $-0.2$°C (larger than +0.2°C). The thick black lines indicate the axis of large-scale topography (Figure 4(a)), where a well defined cell of high $\sigma^*$ and negative $\gamma_T$ extends from a local maximum over the subtropical plains of the continent ($\sim$ 30°S) to the southern half of the Amazon basin. The subtropical maxima over Central–North America and Southeast Asia are not as marked, in part because they are connected with the entrance regions of the North Atlantic and North Pacific storm tracks. Low-level meridional wind and sea level pressure (SLP) over these three regions also exhibit large values of standard deviation relative to the zonal mean (Figure 3) and their distributions are highly skewed toward the ‘cold surge side’: equatorward flow and high pressure anomalies (not shown).

High values of $\sigma^*$ are also found along the southern coast of Australia, along the coast of South Africa, and extending eastward from the eastern coast of Canada, but none of these regions exhibit $\gamma_T < 0$. In Australia, the summertime maximum of $\sigma^*$ and positive values of $\gamma_T$ are likely caused by synoptic changes of the meridional flow, because offshore winds lead to marked advective warming along the coast (e.g. Ryan et al., 1985). Low-pressure cells trapped along the coastal escarpment of South Africa (e.g. Reason and Jury, 1990) might cause large high-frequency variability at these subtropical latitudes ($\sim$ 30°S), but they produce a rather symmetric distribution of temperature anomalies. Finally, the high values of $\sigma^*$ over eastern Canada reflect the enhancement of the synoptic activity at the entrance of the oceanic storm track.

A second aspect that can be used to identify cold surges in a global perspective is the equatorward movement of the surface anticyclones associated with them. To this effect we constructed maps of the
mean propagation vectors of prominent sea level pressure anomalies (SLPa), using the procedure schematized in Figure 5. For each grid point, the propagation vector was calculated as the 3-day displacement of intense anticyclones and cyclones (within the top or bottom 10% of the SLPa distribution) passing over the reference point at day 0. The procedure tracks the centre of intense anticyclonic and cyclonic cells. It can fail to capture the displacement of high-pressure anomalies that are too flat or move too slow, but this is not the case for most anticyclones associated with cold surges.

The results are shown in Figure 6. Similar maps of propagation vectors of pressure systems are presented in Wallace et al. (1988) based on 1-point correlation analysis, but only for the Northern Hemisphere. The propagation field is dominated by the eastward displacement (~10 m s$^{-1}$) of cyclones and anticyclones, particularly coherent along the extratropical storm tracks over the North Pacific, North Atlantic and southern oceans. However, the surface pressure systems acquire a pronounced anticyclonic path around large-scale orographic features, best defined during winter. Such a steering effect of the earth’s topography on synoptic-scale disturbances at low-levels has been earlier documented over the Northern Hemisphere by Whittaker and Horn (1983), Hsu and Wallace (1985), Hsu (1987) and Wallace et al. (1988).

The mean equatorward propagation of positive SLPa to the east of the Himalayas, the Rockies and the Andes, is interpreted as the incursion of cold surface anticyclones from mid-latitudes into the subtropics associated with cold surges. Over these three regions the centre of intense anticyclones moves equatorward at about 10 m s$^{-1}$, roughly the same speed of the meridional low-level wind during typical cold surges. This aspect is in agreement with the hypothesis that horizontal advection is the dominant mechanism in the incursion of cold surges into low latitudes. Negative SLPs, associated with intense low pressure cells, also move equatorward along the east side of the major mountain ranges, probably associated with lee troughing at subtropical latitudes (e.g. Physick, 1981). At subtropical latitudes, however, the mean meridional component of positive anomalies is about twice as large as the meridional component of negative anomalies, an asymmetric behaviour inconsistent with the interpretation of such perturbations as linear Rossby or Kelvin waves trapped along the topography.

Figure 5. Schematic of the method for estimating the propagation vector. For a given grid box (represented by an open square) we define the composite SLP pattern as the average map of the days that fall within the top (or bottom) 10% of the frequency distribution. The composite anomaly map is calculated as the composite map minus the long-term mean SLP field. The contours represent isolines of the composite SLPa on days $-1$, 0, and $+1$. The solid circles are the local extrema (maxima or minima) on days $-1$, 0, and $+1$, with values of $P_{-1}$, $P_0$ and $P_{+1}$, respectively. The propagation vector is defined as the vector connecting $P_{-1}$ and $P_{+1}$ (the thin, long vector) scaled by 48 h and translated to the base grid point (thick, short vector).
Figure 6. Propagation vectors for prominent positive (thick arrows) and negative (thin arrows) SLPa for (a) June–August, and (b) December–February. Vectors are shown only at grid points where the SLPa on day 0 exceeds ±1 hPa. Light (medium) shading indicates terrain elevations above 1000 m MSL (3000 m MSL). The reference vector is shown at the top of panel (a). See Figure 5 and the text for details.

4. CONTRIBUTION OF COLD SURGES TO THE SUBTROPICAL AND TROPICAL ENERGY BALANCE

Incursions of mid-latitude air inject cold dry air into the lower tropical troposphere, and therefore they represent an energy sink for the Tropics. In this section, we evaluate the magnitude of this effect based on the mean heat and moisture fluxes over the three regions where cold surges are frequent. The heat flux by transient eddies is defined as $F = T' u' i + T' v' j$, where $u'$ and $v'$ are the transient zonal and meridional components of the wind, respectively, and $T'$ is the temperature departure from the time mean. The relevant term associated with cold surges is the meridional transport ($T' v'$) evaluated in the lower troposphere, since during these events $u' \approx 0$. Figure 7 shows the meridional heat transport at the 925-hPa level for the boreal and austral winter, constructed by calculating the daily temperature and
Figure 7. The 925-hPa poleward heat transport by transient eddies \( (\bar{T}v') \), primes denote departure from long-term mean) for (a) May–September, and (b) November–March. Meridional wind and temperature were previously band-pass filtered to retain fluctuations in the 3–12-day range. Black areas indicate terrain elevation above 1000 m MSL.

Meridional wind departures from the time mean at each grid point \((\bar{T} \text{ and } \bar{v})\), and then averaging their product during the 17 seasons of record. The transient heat transport is poleward everywhere, but within the equatorial belt \( (5^\circ \text{S}–5^\circ \text{N}) \), where there are some places with a very small (less than \(2^\circ \text{C m s}^{-1}\)) equatorward heat transport. Maximum poleward heat transport in the lower troposphere is found along the mid-latitude storm tracks. Over much of the subtropics the field is weak and rather flat, except for wintertime maxima of \( \bar{T}v' \) to the east of the Rockies, over Southeast Asia and, particularly well defined, to the east of the Andes. A similar distribution is found for the meridional moisture flux by transient eddies \( (q'v') \) in the lower troposphere (not shown). Consistent with the shallow character of the cold surges, the subtropical maxima of meridional heat (and moisture) flux are confined to below the 600-hPa level as shown in longitude-pressure cross sections of \( \bar{T}v' \) at \(25^\circ \text{S} \) and \(25^\circ \text{N} \) on Figure 8 (see also Figure 9).

Daily (12:00 universal time coordinated (UTC)) data from a radiosonde station at Resistencia \( (27.3^\circ \text{S}, 60^\circ \text{W}, 52 \text{ m mean sea level (MSL)}) \), within the path of the South American cold surges, from May to September 1994, allow a preliminary assessment of the errors associated with our reanalysis-based calculation of \( \bar{T}v' \) to be presented. At the 925-hPa level, the observed (radiosonde data) and reanalysed meridional heat flux are \(-25.5^\circ \text{C m s}^{-1}\) and \(-27.4^\circ \text{C m s}^{-1}\), respectively, suggesting an uncertainty on the order of \(\pm 3^\circ \text{C m s}^{-1}\) for our estimates of the low-level meridional heat flux by transient eddies over continental areas.
Poleward heat transport can be produced by transient eddies transferring either cold air equatorward (i.e. cold surges) or warm air poleward, so that to determine the origin of the subtropical maxima, $T'\bar{v}$ was partitioned into four components:

$$T'\bar{v} = T'_\alpha \bar{v}'_\alpha + T'_\beta \bar{v}'_\beta + T'_\gamma \bar{v}'_\gamma + T'_\delta \bar{v}'_\delta$$

(2)

where the subscripts indicate the sign of the departure (for instance $\bar{v}'_\alpha = \bar{v}' > 0$: southerly anomaly). Over Central America, Southeast Asia and subtropical South America the ratio of the components associated with cold surges ($T'_\beta \bar{v}'_\beta$ in the Northern Hemisphere and $T'_\gamma \bar{v}'_\gamma$ in the Southern Hemisphere) to the total flux is in the range 0.6–0.8, indicating that subtropical maxima of meridional heat transport are largely due to cold surges. An example of such a partition is shown in Figure 9 for a column over subtropical South America: the vertically integrated (925–600 hPa) $T'_\beta \bar{v}'_\beta$ accounts for 85% of the total meridional heat flux.

The strong divergence of the transient heat flux at the equatorial border of the subtropical maxima implies a significant cooling in the seasonal energy balance in the lower troposphere over subtropical South America, Central America and Southeast Asia. Recalling that $\bar{T}\bar{v}$ is a measure of the mean meridional circulation over subtropical regions, a scale analysis of the heat flux divergence at low levels yields $\nabla F \approx \partial \bar{T}\bar{v}/\partial y \approx -5 \text{C m s}^{-1}/300 \text{ km} = -1.4^\circ \text{C day}^{-1}$ (during the corresponding winter). Such a cooling rate is comparable to the rate of warming by the mean meridional circulation over subtropical regions $\bar{v} \bar{e}/\partial y \approx (5 \text{ m s}^{-1}) \cdot (2^\circ \text{C/500 km}) = +1.7^\circ \text{C day}^{-1}$. Figure 10 shows an example of such a calculation over South America during the austral winter.

The amplitude of the poleward heat and moisture transport in the lower troposphere over subtropical South America (22°C m s$^{-1}$ and 12 g kg$^{-1}$ m s$^{-1}$, respectively, at 925 hPa) is larger, by a factor $\sim 1.5$,
Figure 9. Vertical profiles of the meridional heat transport by transient eddies at 22.5°S, 60°W (subtropical South America) during the austral winter (JJA). The total heat flux \(Q_T = T^\prime v^\prime\) was partitioned in four components defined in Equation (2). Here \(Q_1 = T^\prime v^\prime_+\) (cold surges), \(Q_2 = T^\prime v^\prime_+\) (poleward warm advection), \(Q_3 = T^\prime v^\prime_-\) (equatorward warm advection), and \(Q_4 = T^\prime v^\prime_-\) (poleward cold advection).

than their counterparts in the Northern Hemisphere, and rather puzzling considering the origin of the air masses in these surges. The air mass in Central–North American and Southeast Asian cold surges resides over high-latitude continental areas (Canada and Siberia, respectively) for a long time, and therefore the temperature and dew point drops associated with these surges can be quite large (up to \(-30^\circ\text{C}\)). In contrast, the maritime origin of the air mass in South American cold surges produces smaller temperature drops in most episodes (\(\sim -10^\circ\text{C}\)). Thus, the well-defined maxima of \(T^\prime v^\prime\) and \(q^\prime v^\prime\) to the east of the Andes suggest a very frequent occurrence of South American cold surges penetrating deep into the subtropics, which is in agreement with synoptic climatologies for this region (see a review in Garreaud, 2000).

Finally, we combine the heat and moisture meridional flux into the meridional flux of moist energy by the transient eddies:

\[
\overline{h^\prime v^\prime} = c_p \overline{T^\prime v^\prime} + L \overline{q^\prime v^\prime}
\]

where \(c_p\) is the isobaric heat capacity of dry air and \(L\) is the heat of vaporization of water vapour. To quantify the amount of meridional energy flux associated with cold surges, the profiles of \(\overline{h^\prime v^\prime}\) at fixed latitudes \(\varphi\) were vertically integrated from the surface to 10 km over different ranges of longitude:

\[
Q^\prime = \int_{z = 0}^{z = 10 \text{ km}} \int_{\lambda_1}^{\lambda_2} \rho \overline{h^\prime v^\prime}|_{\varphi} \ d\lambda \ dz
\]

For the austral winter, the meridional moist energy flux at \(\varphi = 25^\circ\text{S}\) integrated over the whole subtropical wall (\(0^\circ – 360^\circ\text{W}\)) is \(Q^\prime|_{\text{SH}} = -2.5 \times 10^{15} \text{ W}\). Integration to the east of the Andes (\(50^\circ – 70^\circ\text{W}\))
yields $Q'_{SA} = -0.6 \times 10^{15} \text{ W}$, almost a fourth of the total hemispheric transient flux out of the tropical belt. For the boreal winter and $\varphi = 25^\circ \text{N}$, the combined flux to the east of the Rockies and Southeast Asia accounts for about a third of the hemispheric flux $Q'_{NH} = +1.9 \times 10^{15} \text{ W}$.

Let us consider a tropical box bounded by subtropical walls at $25^\circ \text{N}$ and $25^\circ \text{S}$, the earth-surface and the troposphere at 10 km, as schematized in Figure 11. The difference $\delta Q = Q'_{SH} - Q'_{NH}$ is the vertically integrated divergence of the meridional moist energy flux by transient eddies, and represents one of the
terms of the energy balance in this box (other terms are: the net radiative flux at the top of the box, the surface latent heat flux, and the mean meridional energy flux). During the austral winter \( \delta Q' \approx Q'_{\text{SH}} \), since \( Q'_{\text{SH}} \gg Q'_{\text{NH}} \sim 0 \), while \( \delta Q' \approx Q'_{\text{NH}} \) during the boreal winter. In both cases, however, \( \delta Q' \approx 2.5 \times 10^{15} \) W (out of the tropical box) at the height of the austral or boreal winter. To place the previous numbers in context, we calculated the annual mean latent heat flux at the surface:

\[
Q_{\text{sfc}} = \int L \bar{E} \sin \phi \, d\phi \, d\lambda
\]

where \( \bar{E} \) is the long-term mean surface evaporation from the NCEP–NCAR reanalysis. Performing the above integral on the bottom of the tropical box (0–360°W, 25°S–25°N), we obtained \( Q_{\text{sfc}} \sim +29 \times 10^{15} \) W (into the box). Thus, the sink of moist energy by transient eddies is almost a 10% of the surface energy input into the Tropics.

5. CONCLUDING REMARKS

Surges of cold mid-latitude air moving into the subtropics are frequently observed along the east of the Andes, the Rockies and Mexican Sierras, and the Himalayas. Despite case-to-case variability and regional features, cold surges share important aspects in their genesis, structure and evolution. In all cases, the surges are initiated by extratropical disturbances crossing the mountain range, and their subsequent advance takes the form of a shallow dome of cold air moving parallel to the contours of terrain elevation and with a front-like leading edge. Analyses of individual cases suggest that the advance of the cold air dome into subtropical latitudes is largely due to a two-way interaction between the mass and wind field under blocking conditions of flow normal to the barrier. The strong surface pressure gradient drives the equatorward, ageostrophic low-level wind that advects the cold air and dominates the local cooling. In turn, the cold air raises the surface pressure and strengthens the meridional pressure gradient.

An objective survey of cold surges based on two of their defining features (weather changes and anticyclone deflection) identifies the same three regions where this phenomenon has been frequently described. It must be stressed that our analyses are based on a relatively coarse dataset (NCEP–NCAR...
reanalysis) so that they can not identify disturbances with a horizontal scale less than ~ 300 km. The signatures of cold surges over subtropical South America are much clearer than the signatures of North American and Southeast Asian cold surges. We speculate that the clear signatures over South America are a consequence of the favourable topography and the lack of other subtropical synoptic-scale phenomena. Specifically, the narrow and tall Andes cordillera, extending continuously from the southern tip of the continent to the north of the equator with an almost straight north–south orientation, seems ideally suited to foster the northward advance of the surges.

Cold surges not only have a transient impact, but they also play a significant role in the maintenance of the heat balance over the subtropical regions where they are prevalent, with a low-level cooling effect almost as large as the warming produced by the mean meridional circulation. In a global perspective, it was found that the meridional moist energy flux produced by cold surges along the major mountain ranges is a major contributor to the total (vertically and longitudinally integrated) flux by transient eddies out of the subtropics, which in turn represents a sizeable sink of tropical energy balance (e.g. about a 10% of the surface energy flux). It is concluded that a proper representation of the aggregated effect of cold surges trapped along large-scale topography is necessary for a realistic modelling of the Tropics at regional and global scales.

ACKNOWLEDGEMENTS

NCEP–NCAR reanalysis were provided by the NOAA Climate Diagnostics Center. The author gratefully acknowledges the support and advice of Dr John M. Wallace during the course of this work. Thanks are also due to two anonymous reviewers for their constructive comments and Drs P. Aceituno, L. Bosart, B. Liebmann and G. Compo for careful reading of the manuscript and suggestions.

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