



Copertina

Corso di Fisica dello Strato Limite Atmosferico

Strato Limite Atmosferico Situazioni particolari ed applicazioni

Giaiotti Dario

Sommario della lezione

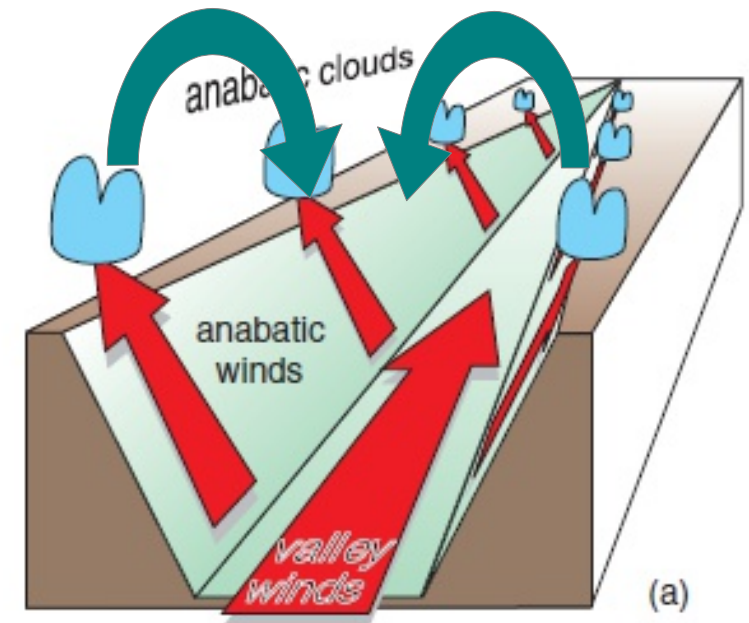
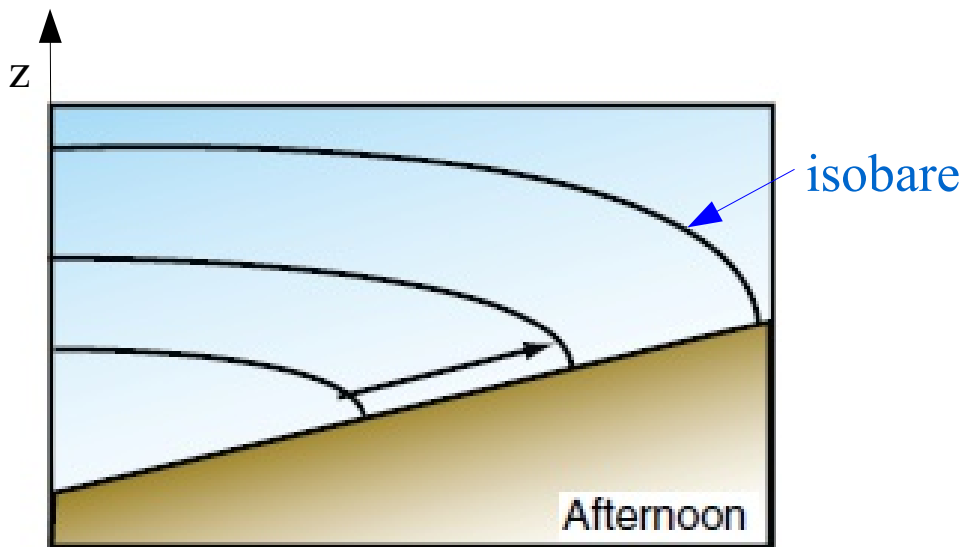
- Strato limite atmosferico in orografia complessa
- Strato limite atmosferico e convezione atmosferica profonda
- Interazioni strato limite atmosferico e flussi sinottici: piogge intense
- Strato limite atmosferico e interazioni mare-terra
- Strato limite atmosferico e cambiamenti climatici
- Risoluzione orizzontale dei modelli e importanza dello strato limite
- Bibliografia di riferimento e per approfondimenti.

Effetti orografici sullo Strato Limite Atmosferico: caso diurno instabile

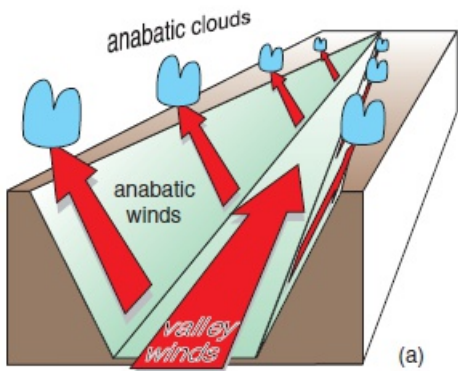
- Riscaldamento dei versanti esposti al flusso ad onda corta
- Formazione di un gradiente termico tra fondo valle e cima
- Circolazione alla mesoscala/microscala (anabatica)
- Formazione di nubi (se il lower condensation level viene raggiunto)
- Chiusura della circolazione con vento discendente (in quota) al centro della valle



Il **gradiente di pressione** lungo i versanti riscaldati è diverso da quello dell'equilibrio idrostatico nella libera atmosfera

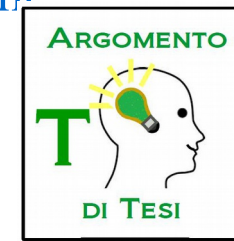


Effetti orografici sullo Strato Limite Atmosferico: caso diurno instabile - Alpi

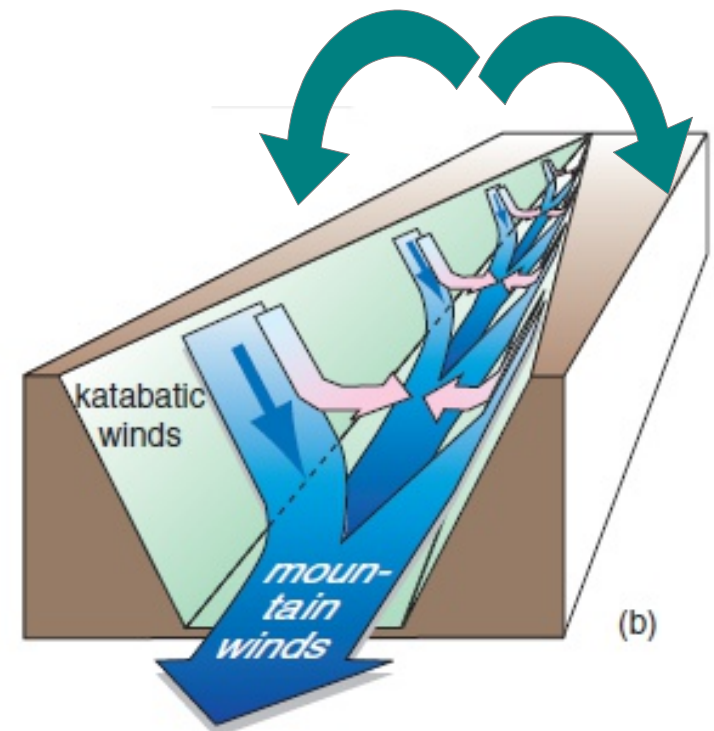
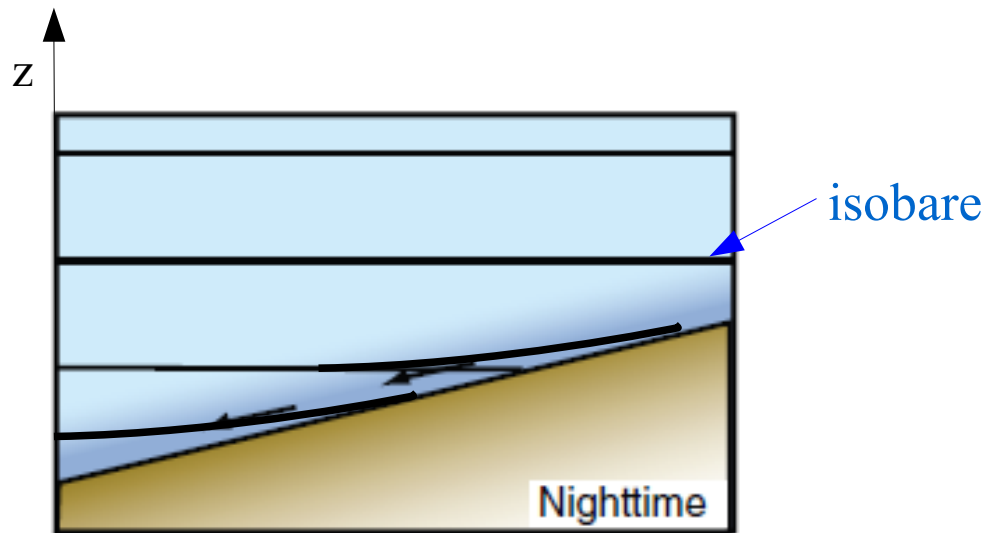


Effetti orografici sullo Strato Limite Atmosferico: caso notturno stabile

- Raffreddamento dei versanti dovuto al flusso ad onda lunga
- Formazione di un gradiente termico tra cima e fondo valle
- Circolazione alla mesoscala/microscala (catabatica)
- Formazione di una inversione termica sopra la valle (irraggiamento verso il centro)
- Chiusura della circolazione con vento ascendente (in quota) al centro della valle

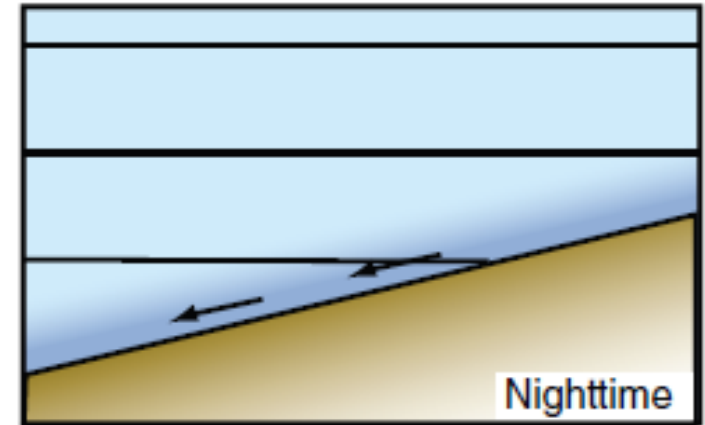


Il **gradiente di pressione** lungo i versanti riscaldati è diverso da quello dell'equilibrio idrostatico nella libera atmosfera



Effetti orografici sullo Strato Limite Atmosferico: intensità dei venti catabatici

Il vento che scende lungo i versanti può essere considerato come uno strato di spessore h e temperatura T_k , mentre l'aria circostante ha temperatura T . C_D è il coefficiente d'attrito ed α è la pendenza del piano lungo il quale scivola l'aria fredda.

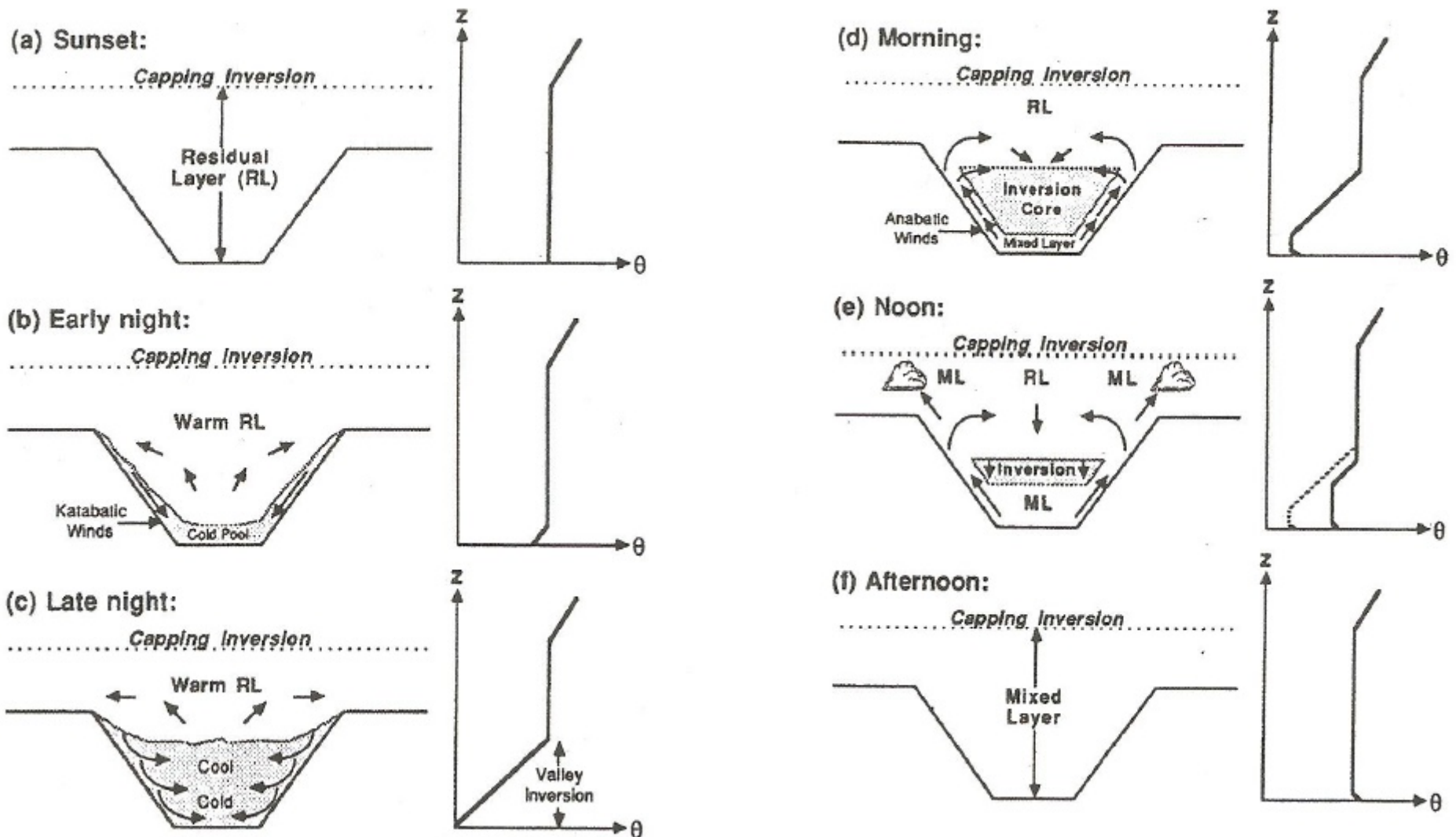


u_k è la velocità del vento in discesa dal versante

$$u_k = \left(\frac{g \cdot (T - T_k)}{T} \cdot \frac{h}{C_D} \cdot \sin \alpha \right)^{1/2},$$

Effetti orografici su ABL: la circolazione tipica in giornate/nottate serene

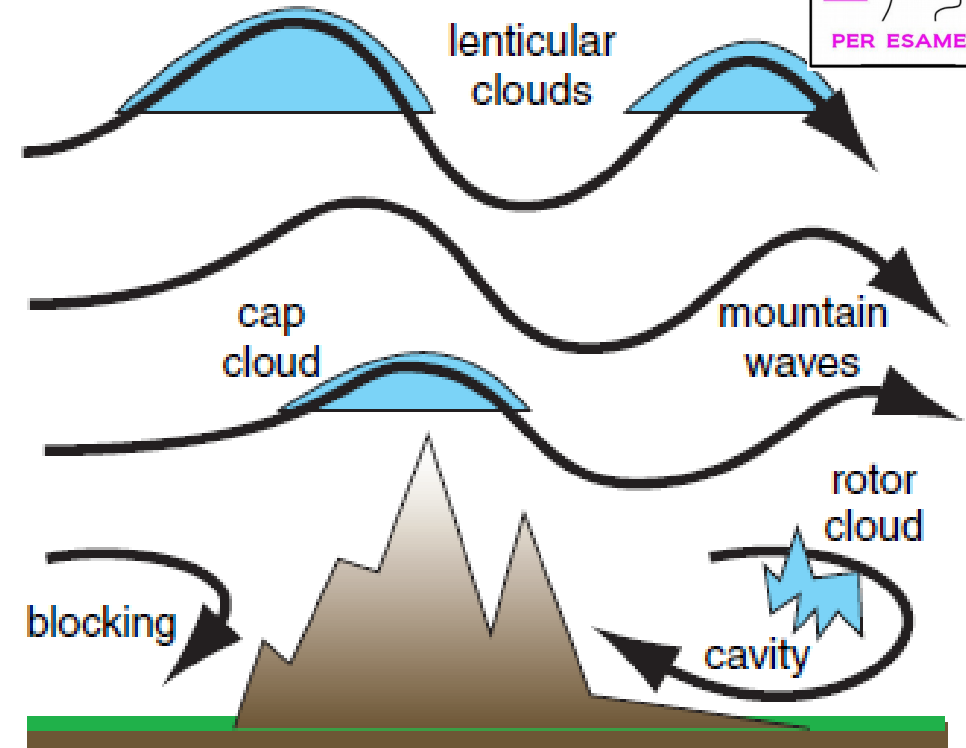
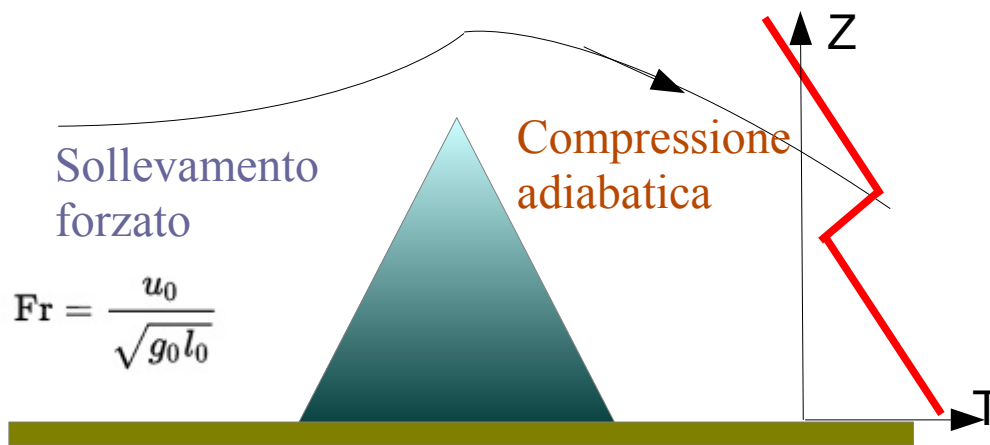
RL residual layer, ML mixed layer



Effetti orografici sullo Strato Limite Atmosferico: in presenza di forzanti sinottiche

L'orografia non modifica solo il flusso sinottico in quota, sopra i rilievi, bensì anche:

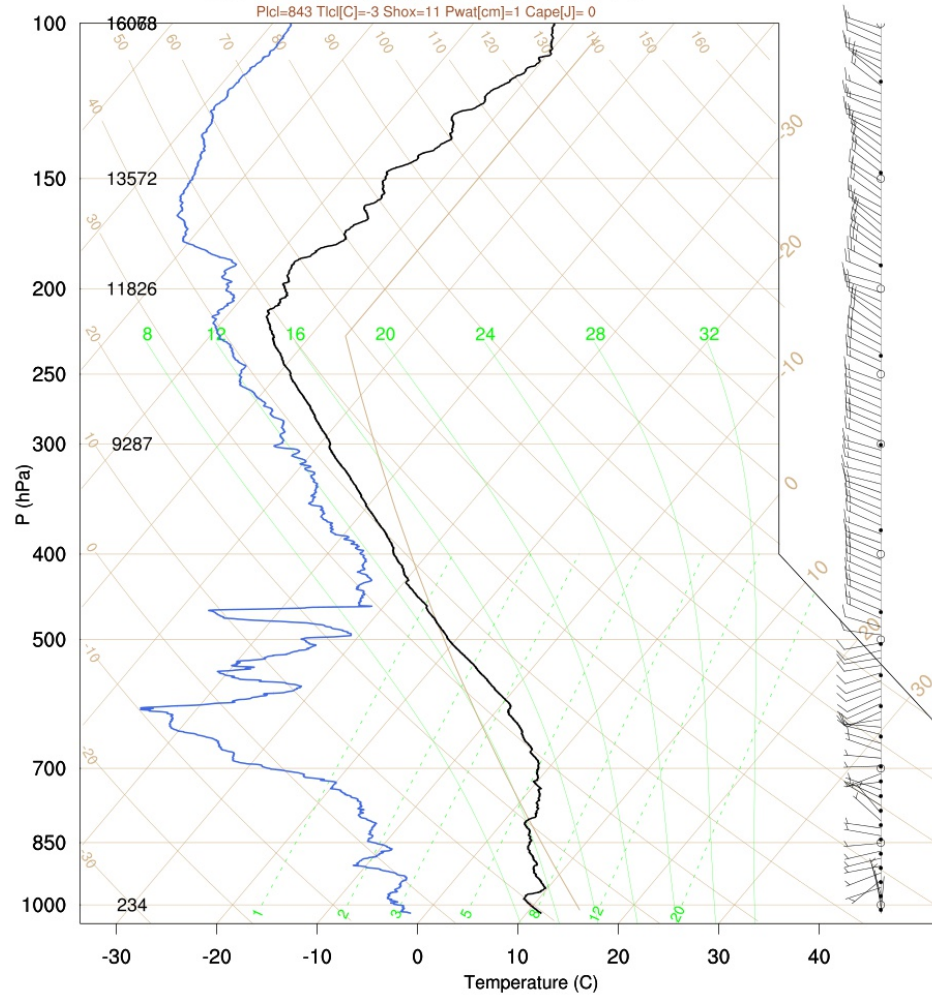
- Lo strato limite atmosferico **sopravvento**
- Lo strato limite atmosferico **sottovento**



Effetti orografici sullo Strato Limite Atmosferico: riscaldamento adiabatico flusso sinottico

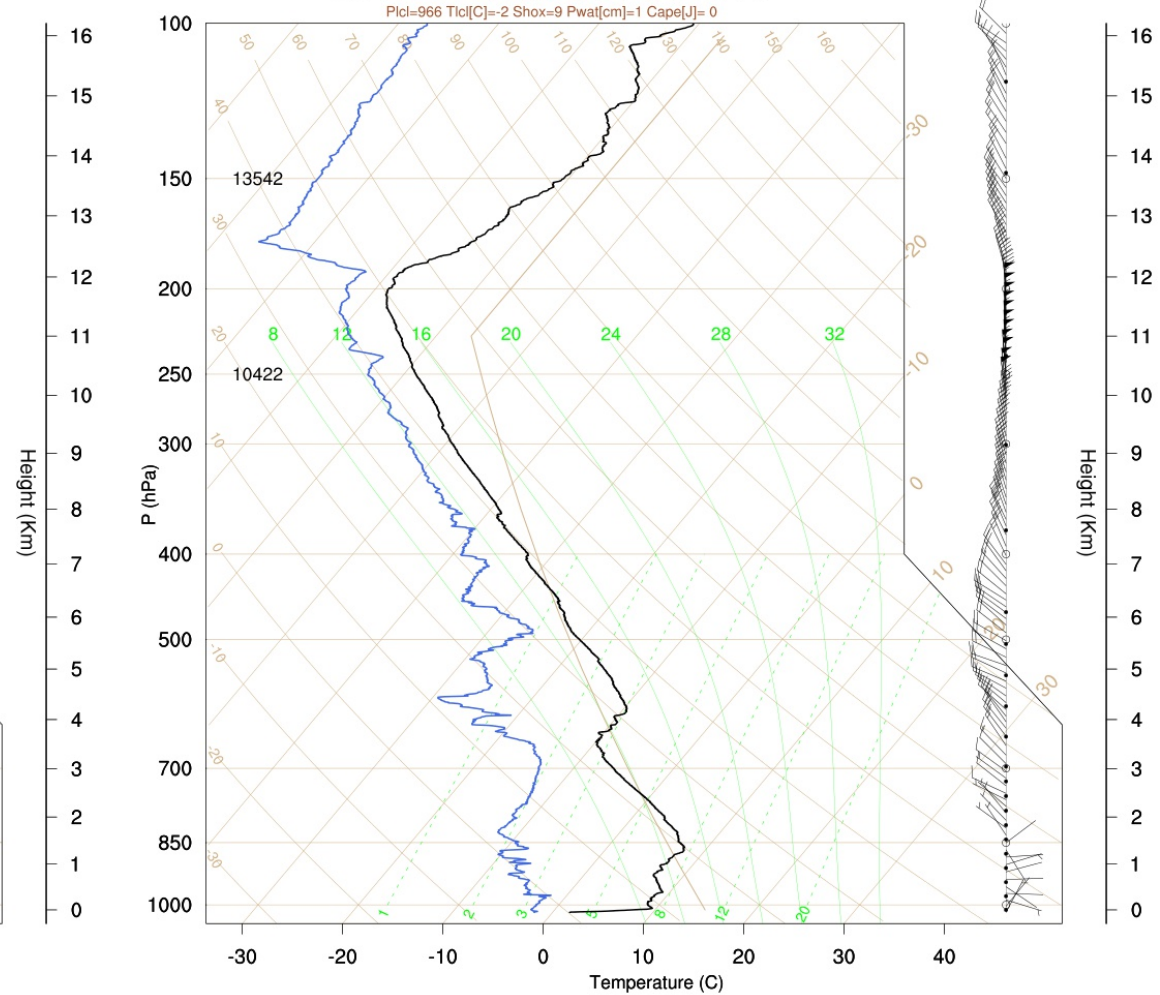
16044 LIPD RIVOLTO 12Z161224

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16044 LIPD RIVOLTO 24Z161224

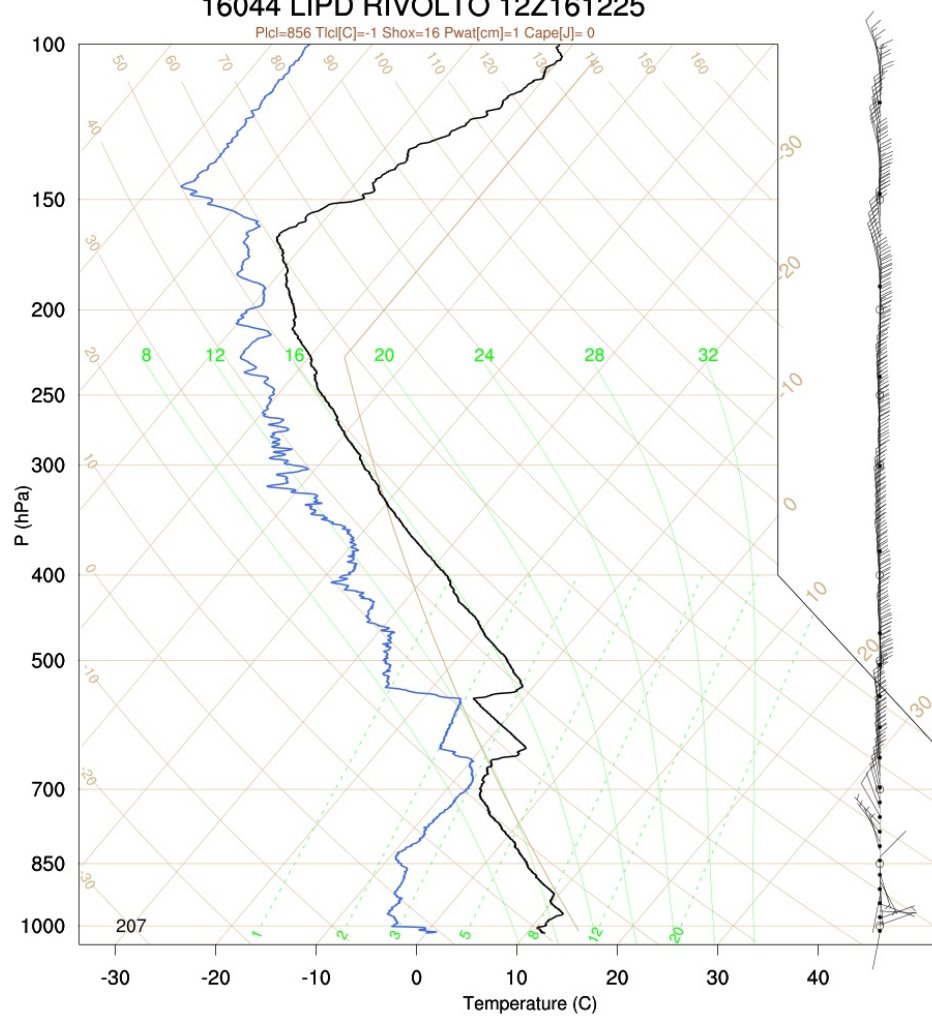
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Effetti orografici sullo Strato Limite Atmosferico: riscaldamento adiabatico flusso sinottico

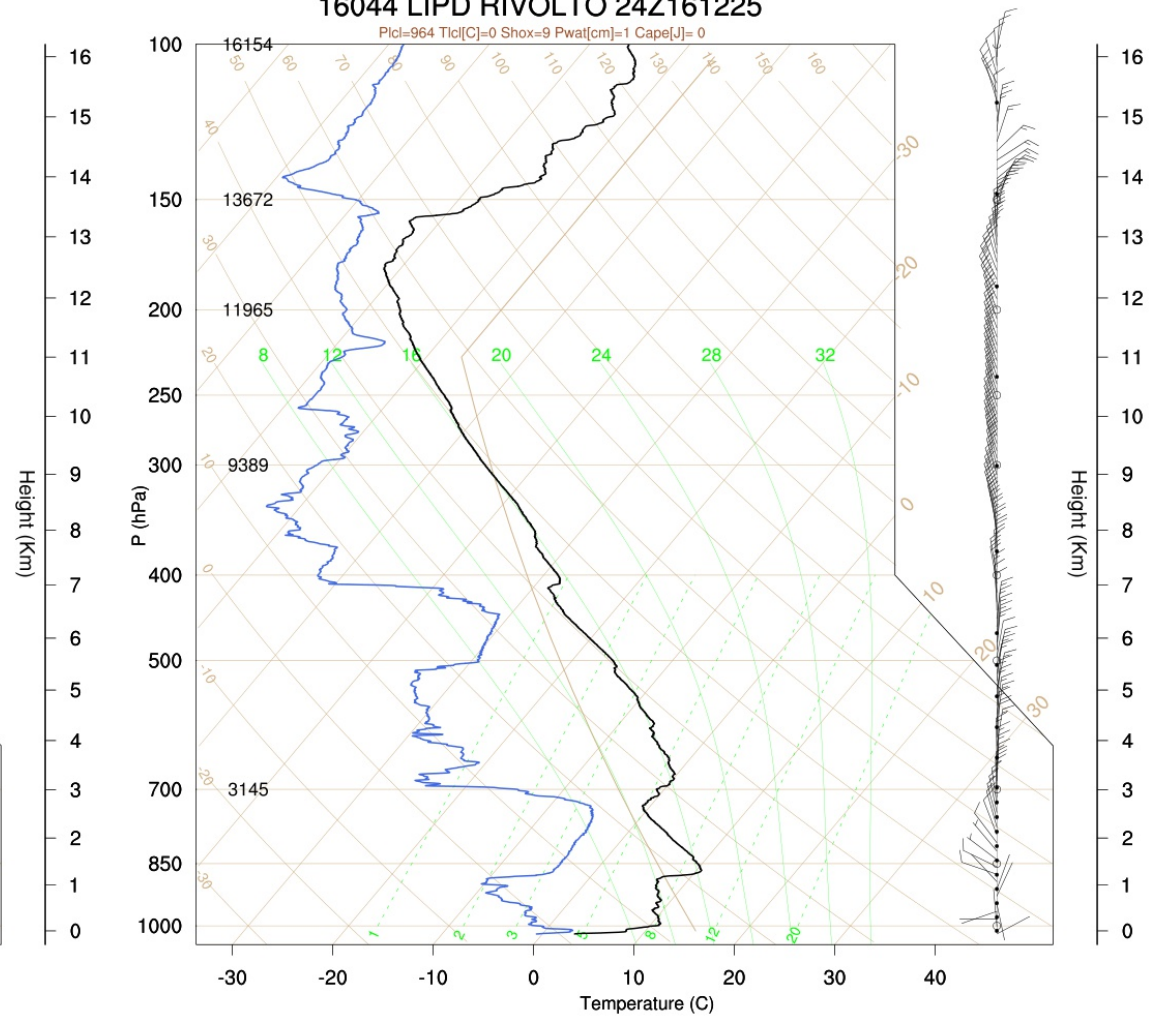
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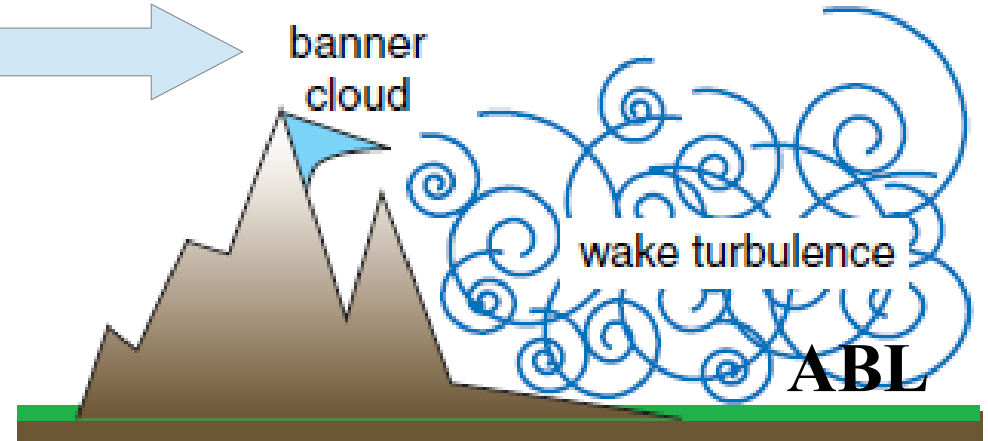
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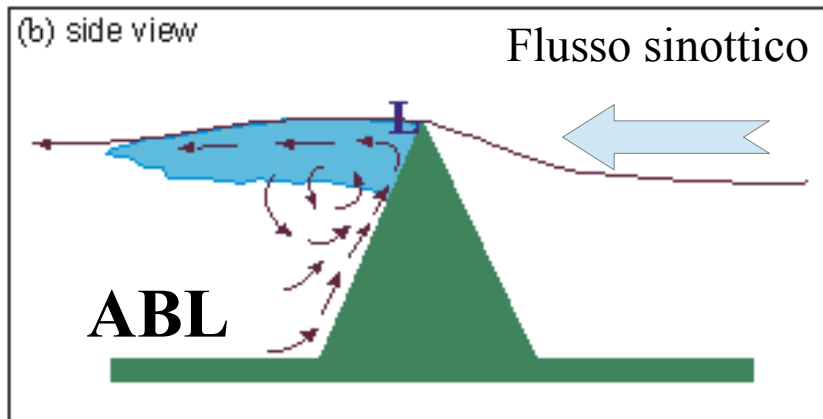
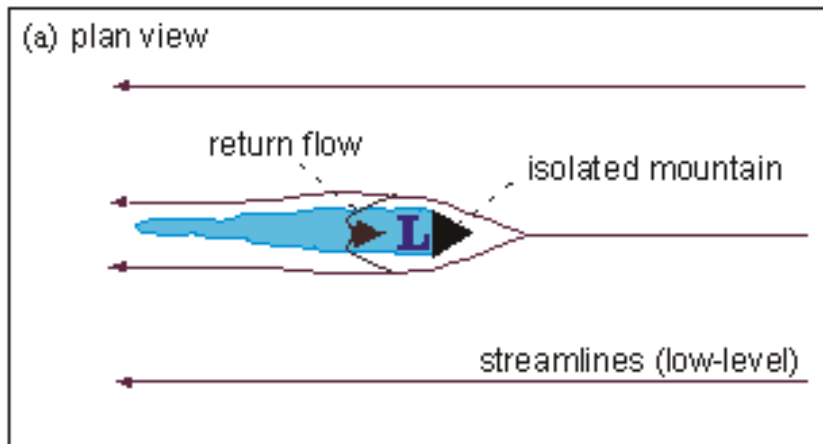
Effetti orografici sullo Strato Limite Atmosferico: rimescolamento e rimozione ABL



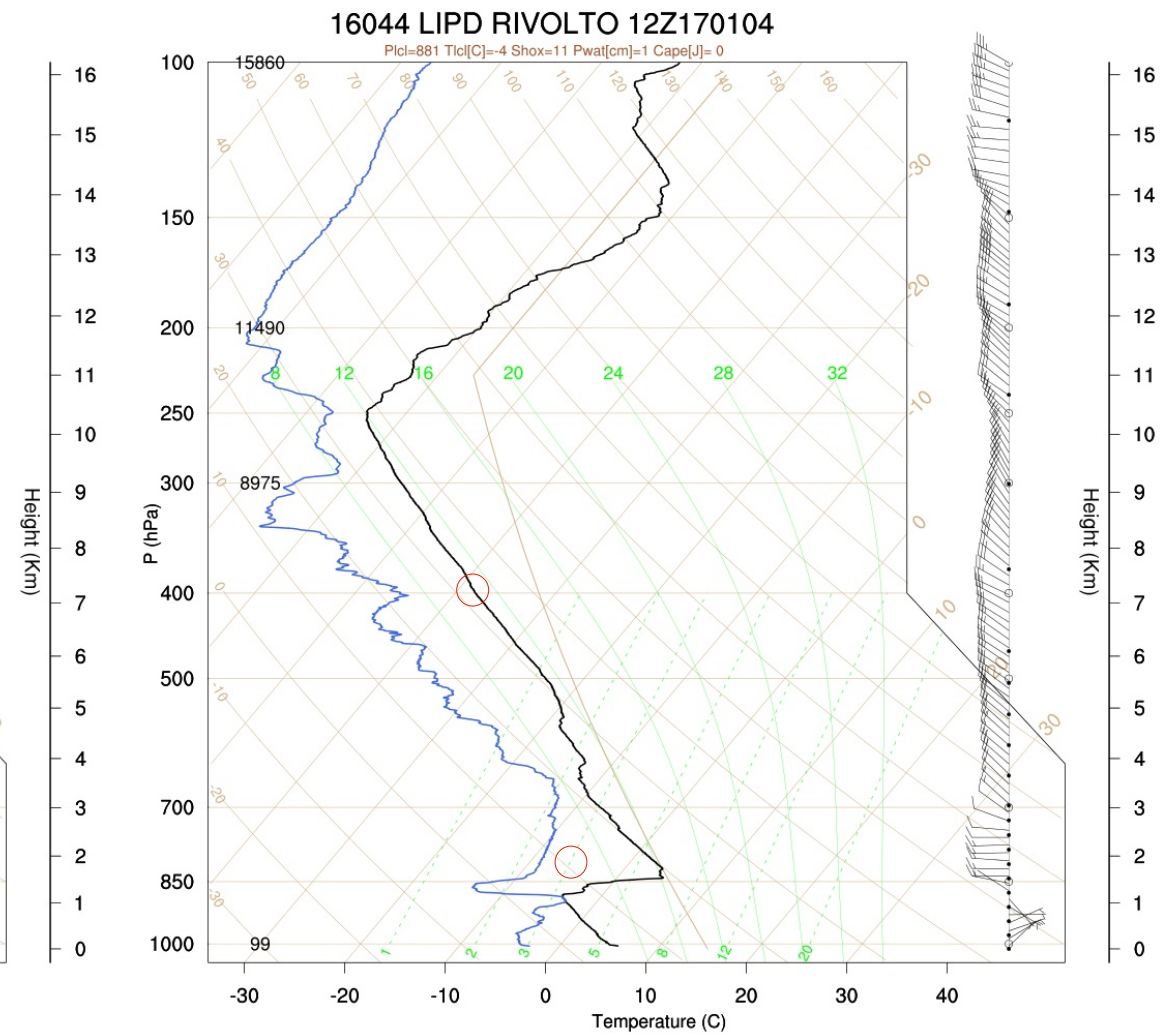
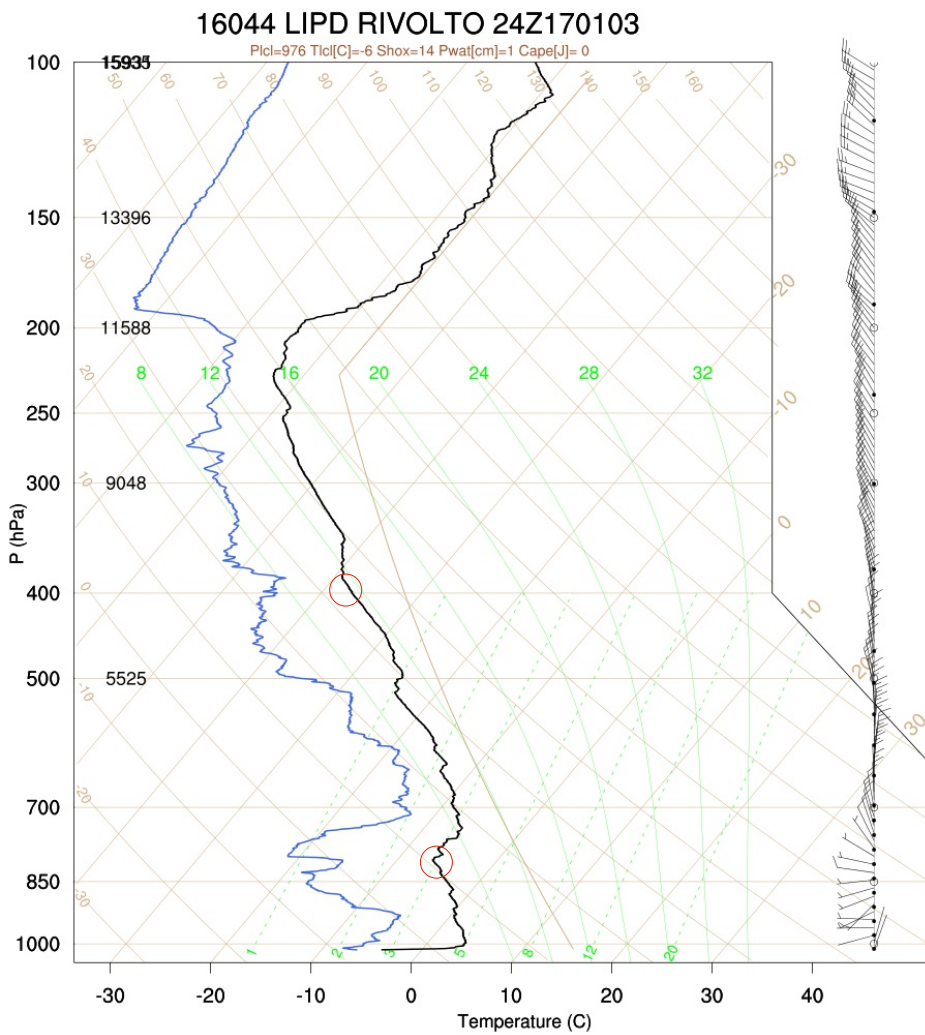
Flusso sinottico



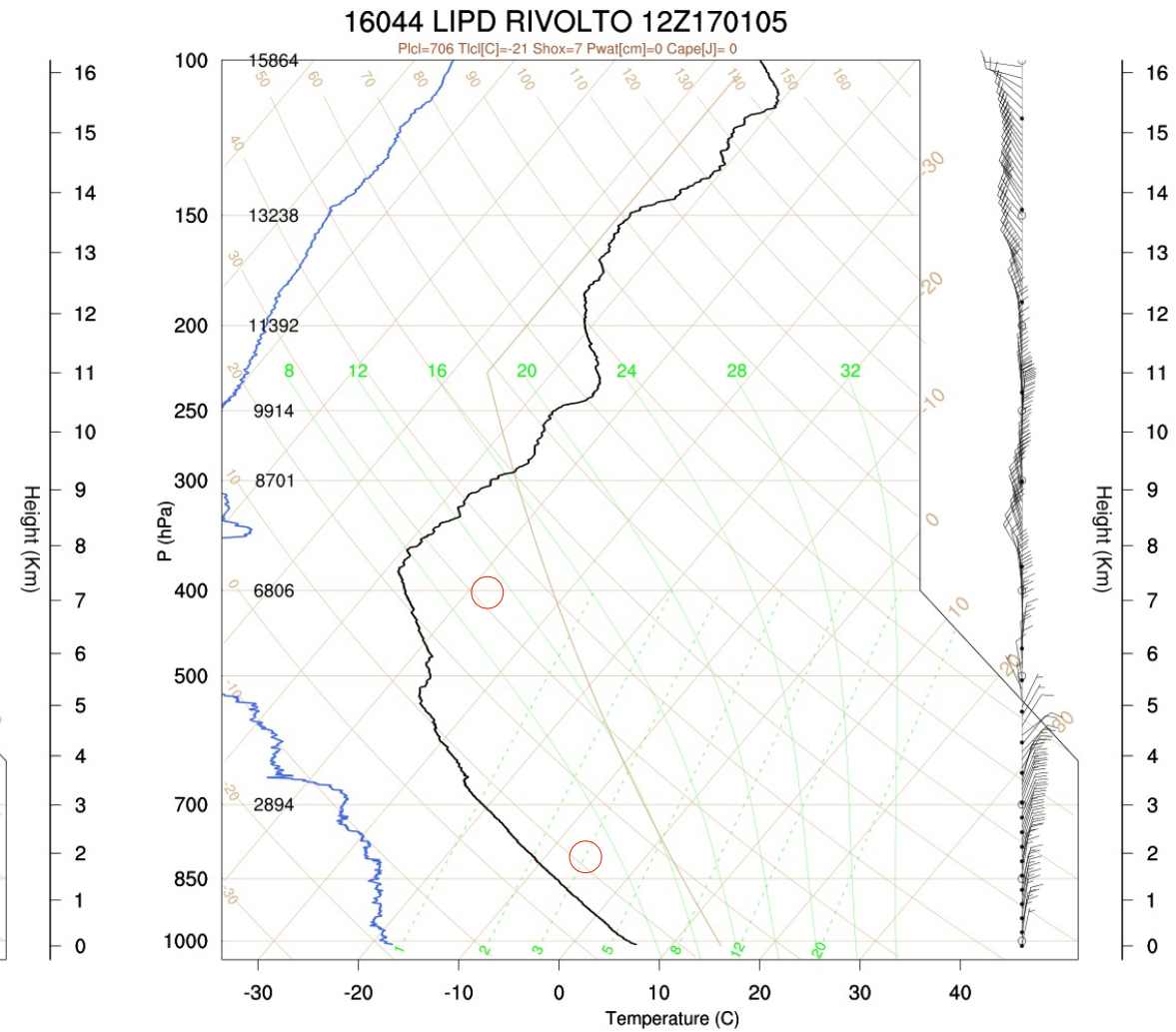
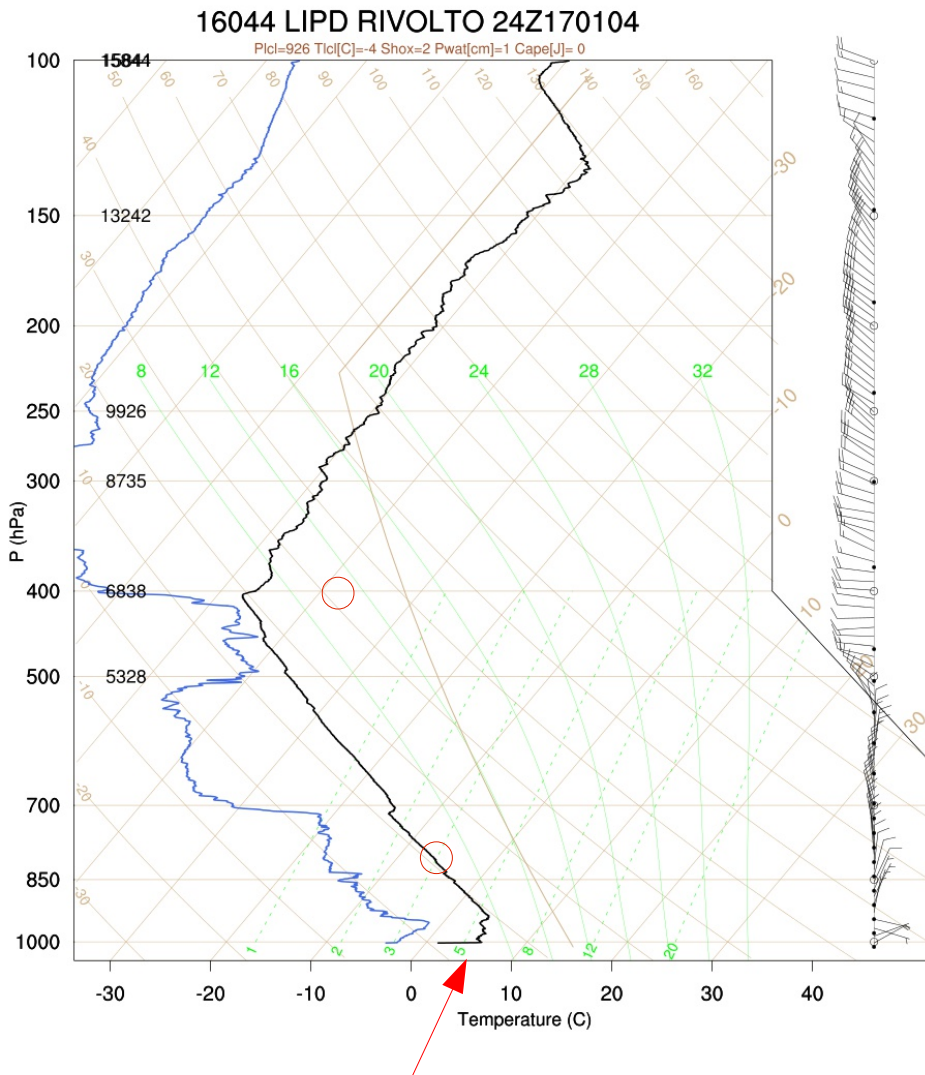
formation of banner clouds



Effetti orografici sullo Strato Limite Atmosferico: rimozione strato limite sottovento



Effetti orografici sullo Strato Limite Atmosferico: rimozione strato limite sottovento



Effetti orografici sullo Strato Limite Atmosferico: rimozione strato limite sottovento

ARPA FVG - OSMER

Stazione: UDINE SANT'OSVALDO
05 gennaio 2017

ora UTC*	Pioggia mm	Temp °C	Umidità %	Vento med km/h	Direzione Vento	Vento max km/h	Direzione Vento max	Radiaz. KJ/m2	Pressione hPa	Bagnatura Fogliare min
1.00	0.0	-0.8	76	6	N	13	NE	7	995.3	0
2.00	0.0	-0.1	80	7	NNW	11	N	1	995.3	0
3.00	0.0	-0.8	79	7	NE	15	E	11	995.4	0
4.00	0.0	-2.3	80	4	ESE	10	ENE	10	995.9	0
5.00	0.0	-3.6	78	4	NE	12	E	8	995.8	12
6.00	0.0	2.1	53	8	NNW	19	NNW	6	996.4	0
7.00	0.0	4.8	29	19	N	39	N	7	998.0	0
8.00	0.0	6.0	15	25	NNW	47	NW	185	999.3	0
9.00	0.0	4.8	37	31	N	55	N	800	1001.7	0
10.00	0.0	5.9	9	36	N	59	NNW	1092	1002.2	0
11.00	0.0	6.4	9	26	N	40	N	1301	1002.1	0
12.00	0.0	6.5	9	32	NNW	55	N	1330	1001.9	0
13.00	0.0	6.2	9	28	NNW	49	NNW	1166	1002.1	0
14.00	0.0	5.7	9	32	N	59	NW	905	1003.2	0
15.00	0.0	4.4	11	21	NNE	40	NNW	365	1003.9	0
16.00	0.0	4.1	10	29	N	46	N	18	1004.5	0
17.00	0.0	3.7	10	28	N	46	NNW	7	1005.4	0
18.00	0.0	3.5	9	25	N	48	NNW	4	1006.1	0
19.00	0.0	2.8	10	28	NNW	45	N	5	1006.5	0
20.00	0.0	2.4	10	23	NNW	44	N	5	1007.5	0

* ora UTC (ora di Greenwich)

per ottenere quella locale bisogna aggiungere 1 o 2 ore rispettivamente nel caso in cui sia in vigore l'ora solare o quella legale

[scarica dati in formato csv](#)

Strato Limite Atmosferico in condizioni particolari (Fronti e Convezione Atmosferica Profonda)

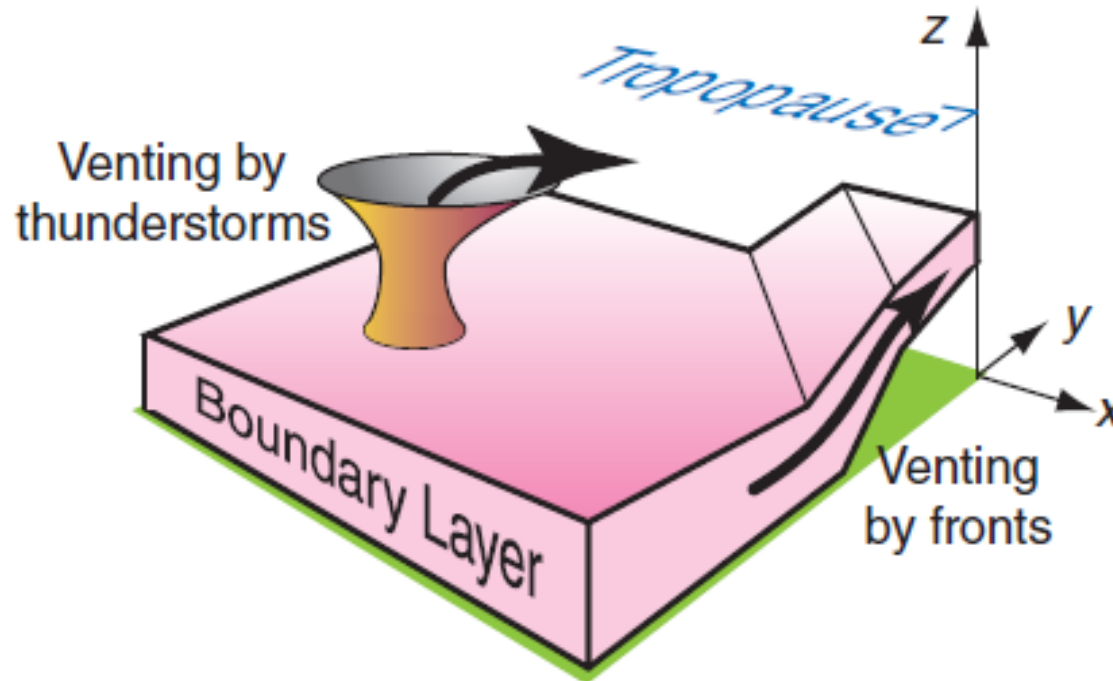


Fig. 9.29 The two dominant mechanisms for boundary-layer air to pass the capping inversion and enter the free atmosphere. [Courtesy of Roland B. Stull.]

Strato Limite Atmosferico in condizioni particolari (Fronti e Convezione Atmosferica Profonda)

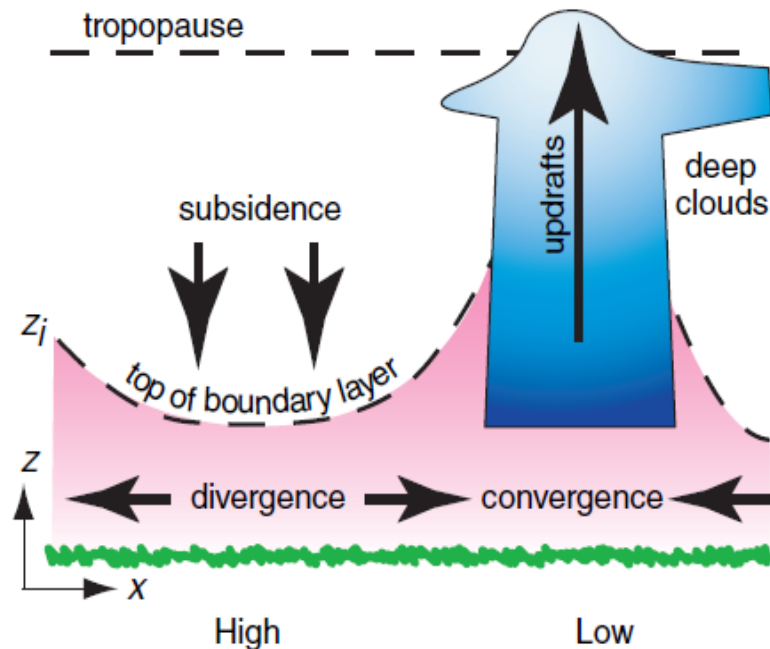


Fig. 9.28 Vertical slice through an idealized atmosphere showing variations in the boundary layer depth z_i in cyclonic (low pressure) and anticyclonic (high pressure) weather conditions. [Adapted from *Meteorology for Scientists and Engineers*, A

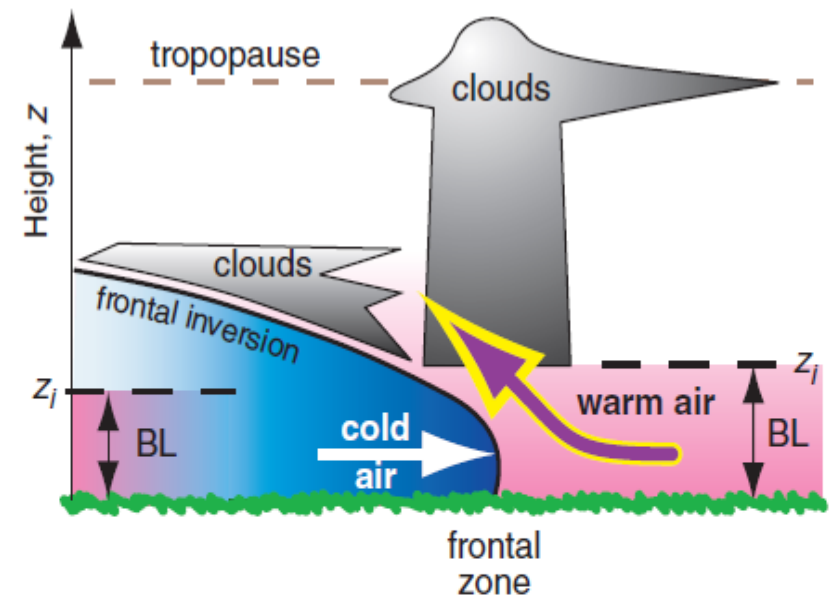
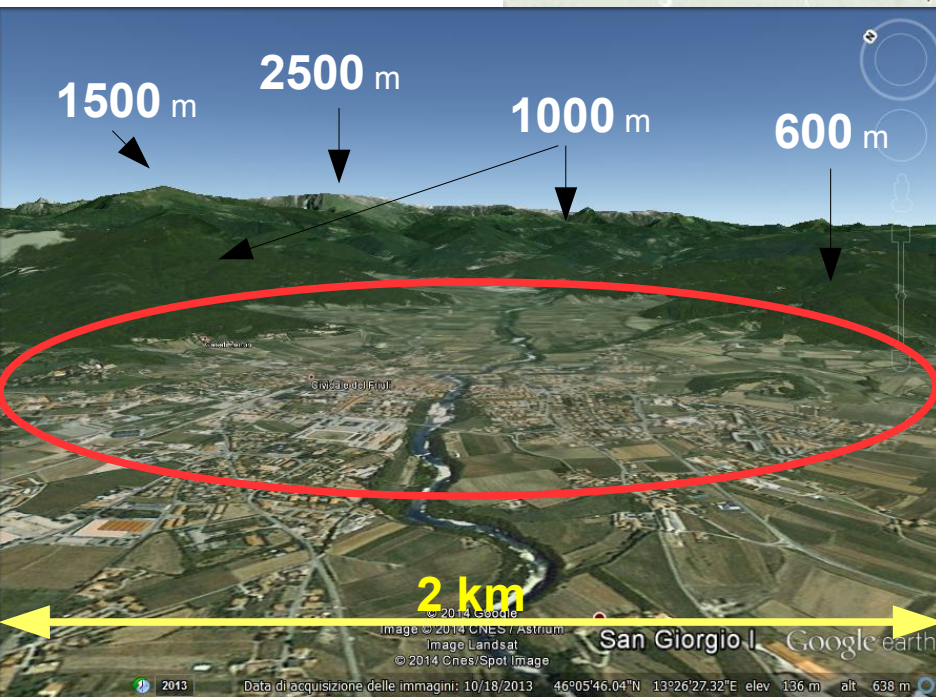
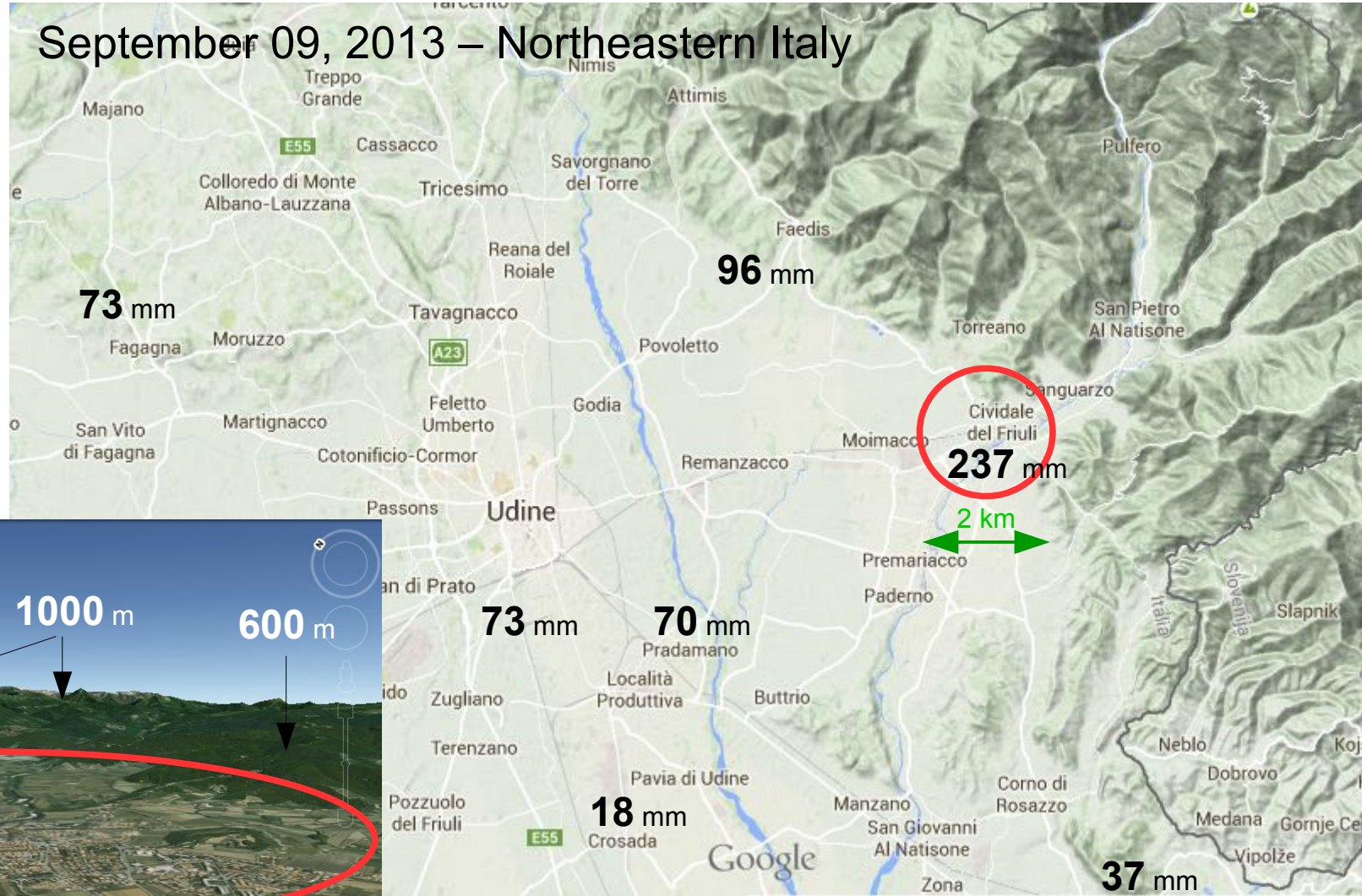
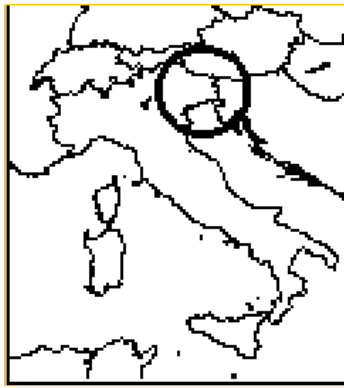


Fig. 9.30 Vertical cross section illustrating the destruction of the capping inversion z_i and venting of boundary layer (BL) air near frontal zones, and subsequent reestablishment of the capping inversion under the frontal inversion after cold frontal passage. [Adapted from *Meteorology for Scientists and Engineers*, A

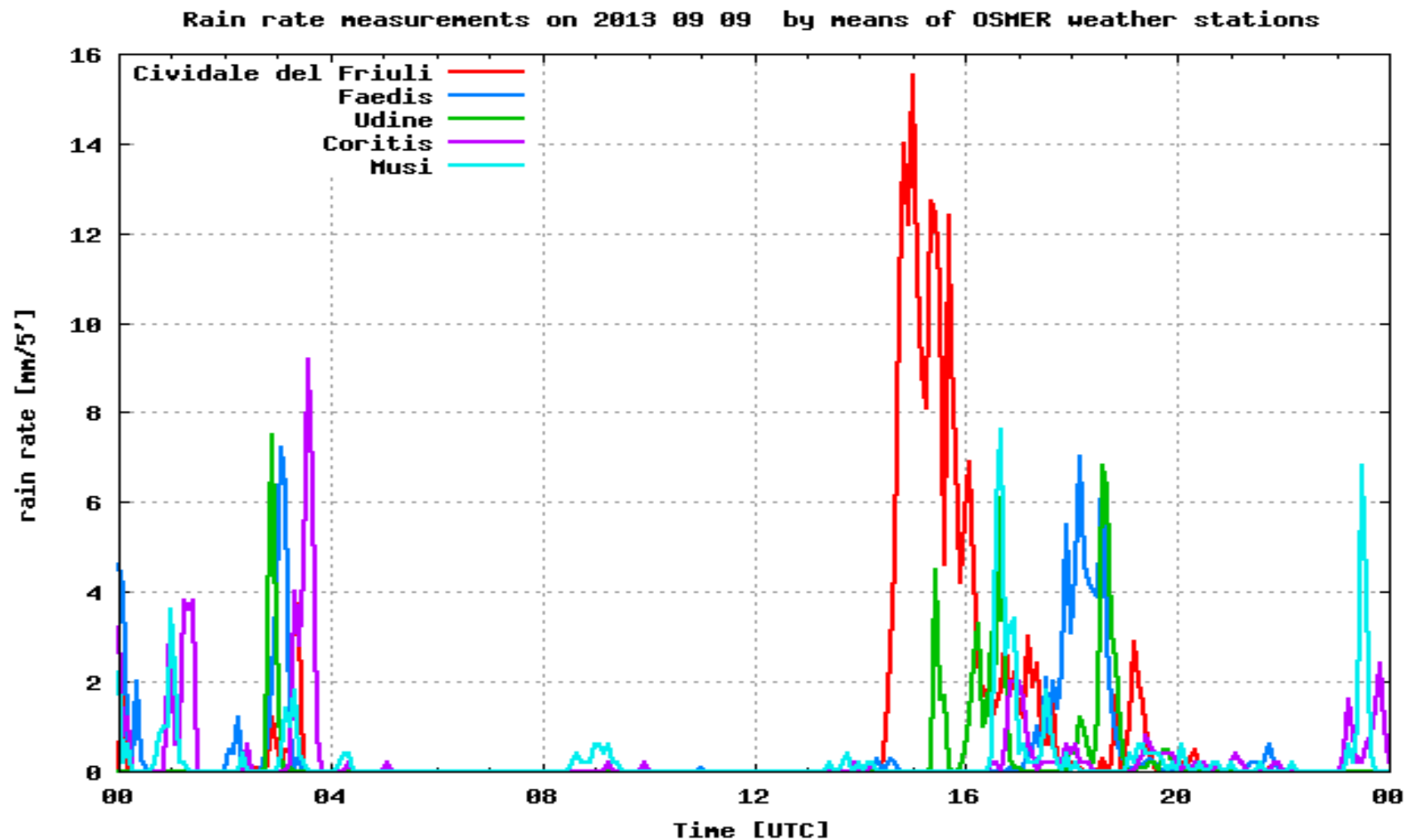
Ruolo dello Strato limite Atmosferico nelle piogge intense e localizzate



100 civil protection volunteers involved to help flooded people
1 MEuro of costs and damages
 At least **one event/year** in the area (100 km x 100 km)

Ruolo dello Strato limite Atmosferico nelle piogge intense e localizzate

- High efficiency in water vapor condensation and precipitation (up to 20 mm/5')
- Persistence for more than one hour
- Large horizontal rain rate gradients (>100% change in about 2 km)



Ruolo dello Strato limite Atmosferico nelle piogge intense e localizzate

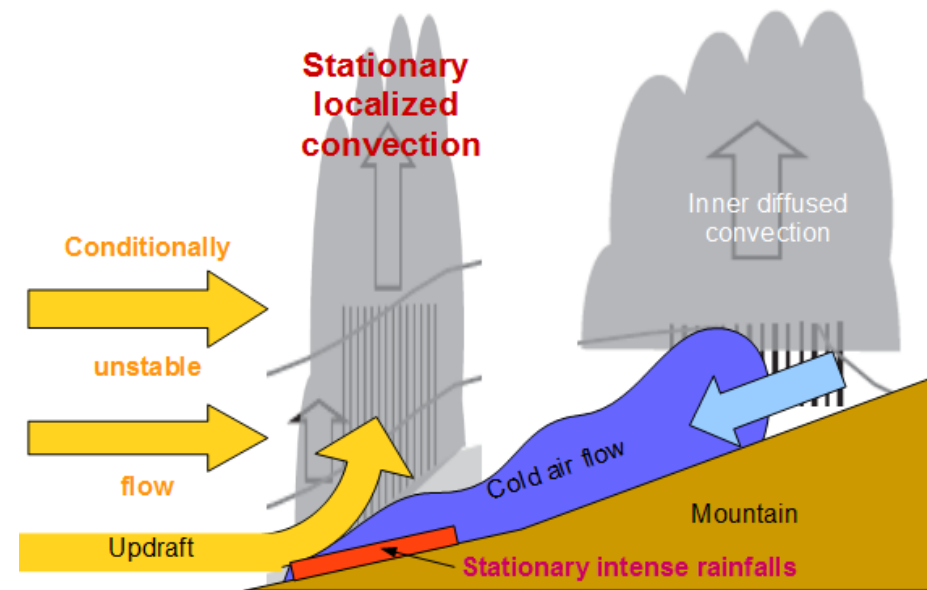
Deep atmospheric convection stationary over the same area

Mandatory elements

- Moist conditionally unstable synoptic flow
- Complex orography

Conceptual model

1. Conditionally unstable air impinging on mountains is lifted up to LFC
2. Deep convection take place and generates diffuse downdrafts
3. Downdraft flows are driven by orography in the boundary layer
4. In the boundary layer, synoptic flows and downdrafts interact lifting unstable air
5. The synoptic flow and the downdraft interact stationary in a restricted area

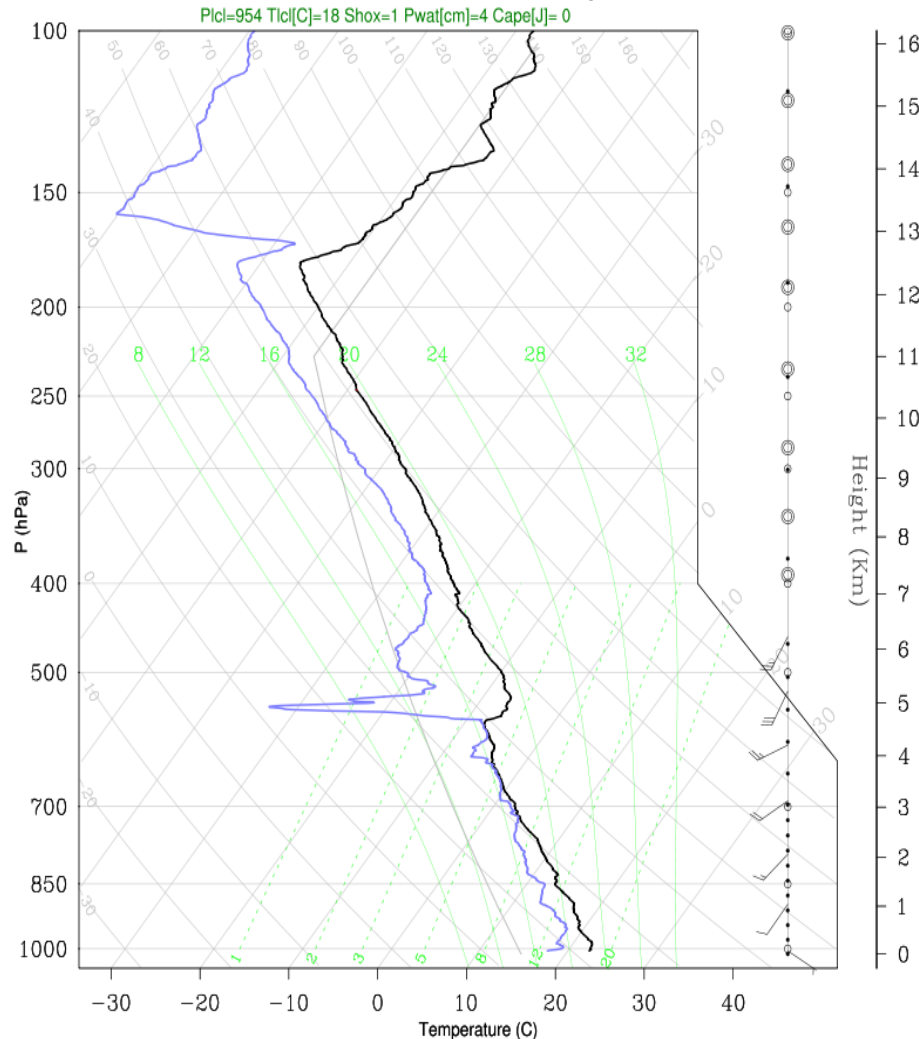


- ☑ The area interested by the interaction is a function of synoptic flow intensity, and stability, the cold air outflow and the orography shape.

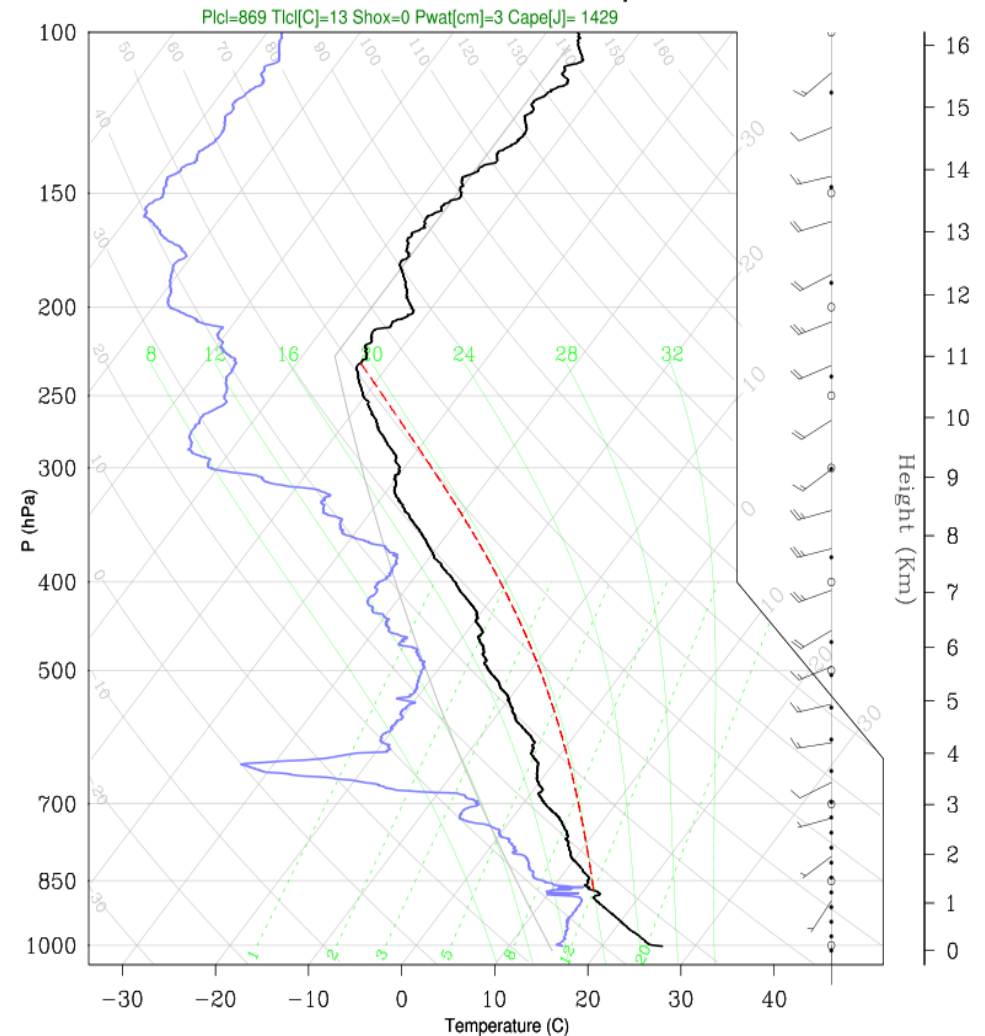
Ruolo dello Strato limite Atmosferico nelle piogge intense e localizzate

The flash flood took place from 14:00 UTC to 16:00 UTC. Here are thermodynamic profiles measured 20 km upstream the event area

16044 RDS at 20130909 00 UTC lifted parcel 150 m amsl



16044 RDS at 20130909 12 UTC lifted parcel 150 m amsl



The synoptic flow was towards the orographic barrier and it was weak 20

Modifica dello strato limite atmosferico in presenza di discontinuità mare-terra

Le differenti capacità termiche della terra e del mare (lago) generano dei gradienti orizzontali di temperatura (quindi di pressione) che inducono moti alla mesoscala (breeze), in assenza di forzanti sinottiche. Lo strato limite atmosferico è fortemente modificato.

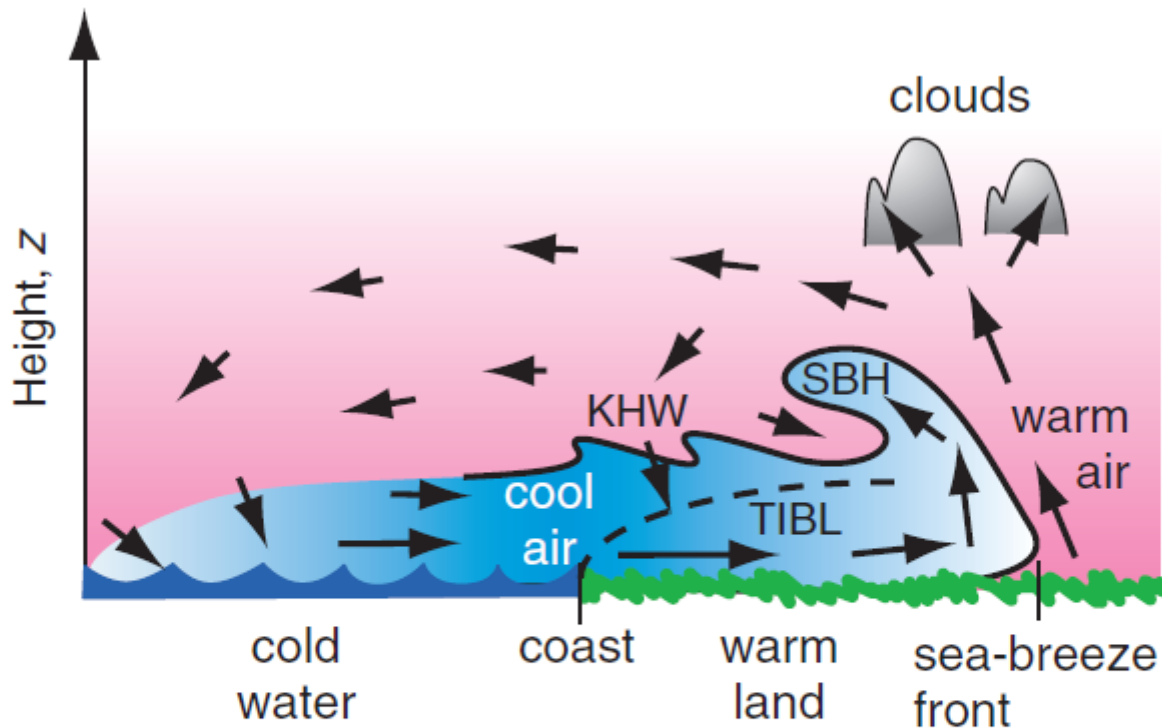
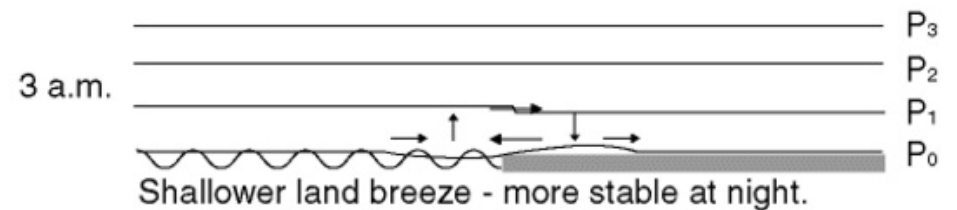
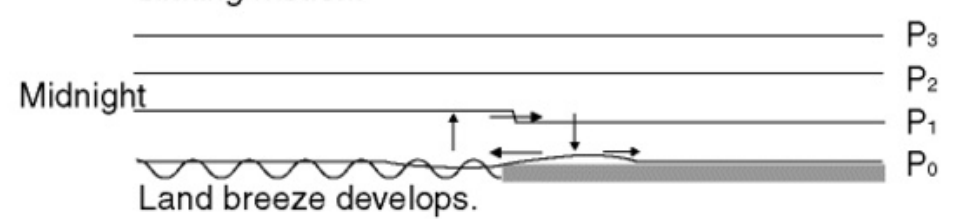
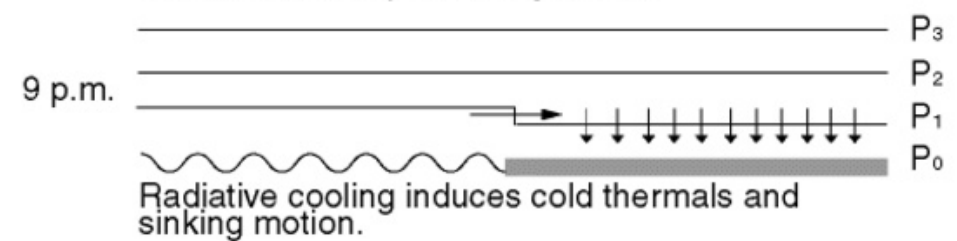
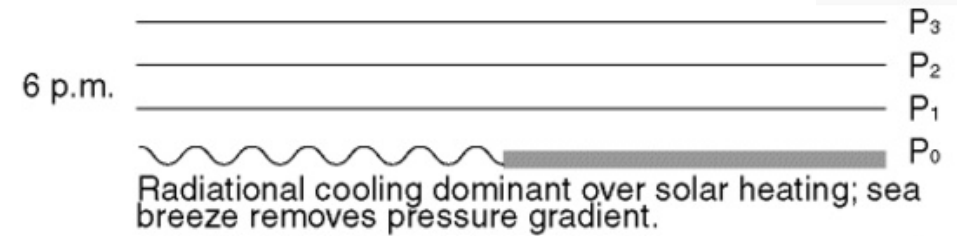
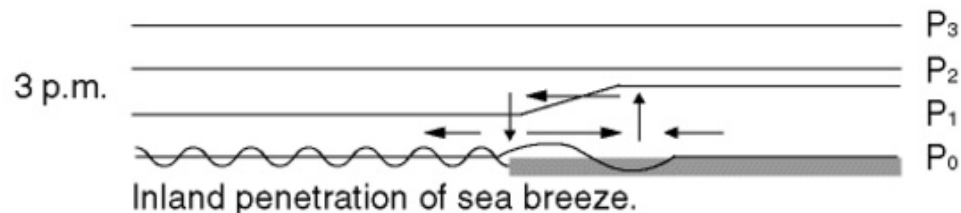
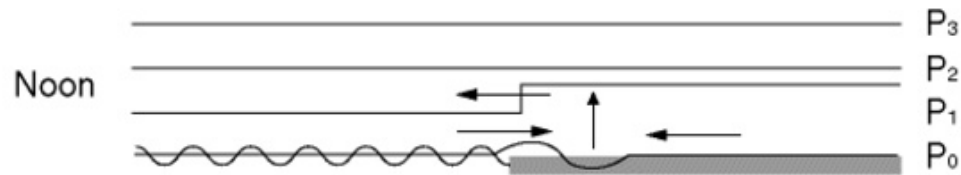
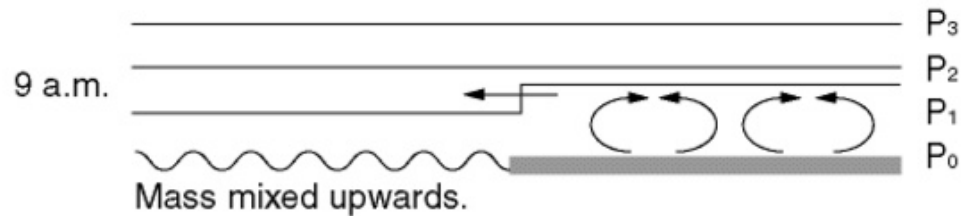
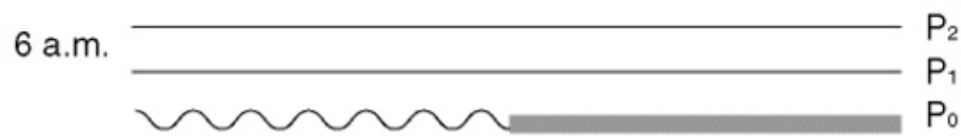


Fig. 9.35 Components of a sea-breeze circulation. SBH, sea-breeze head. KHW, Kelvin-Helmholtz waves. The top of the thermal internal boundary layer (TIBL) is shown by the dashed line. [Adapted from R. B. Stull, *An Introduction to Boundary Layer*



Fronte generato dalla brezza

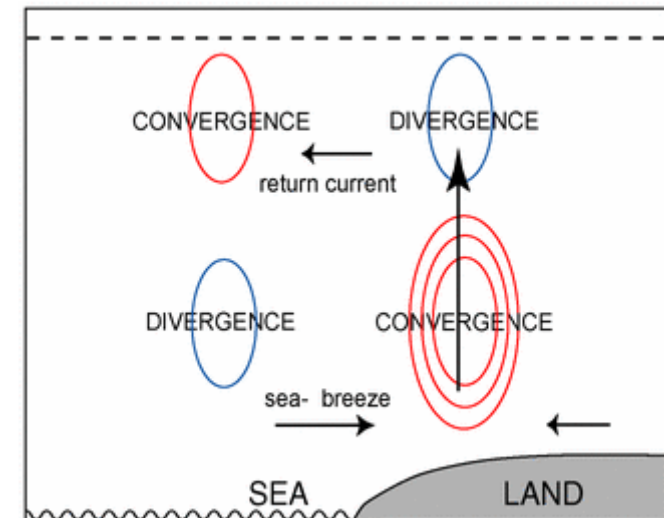
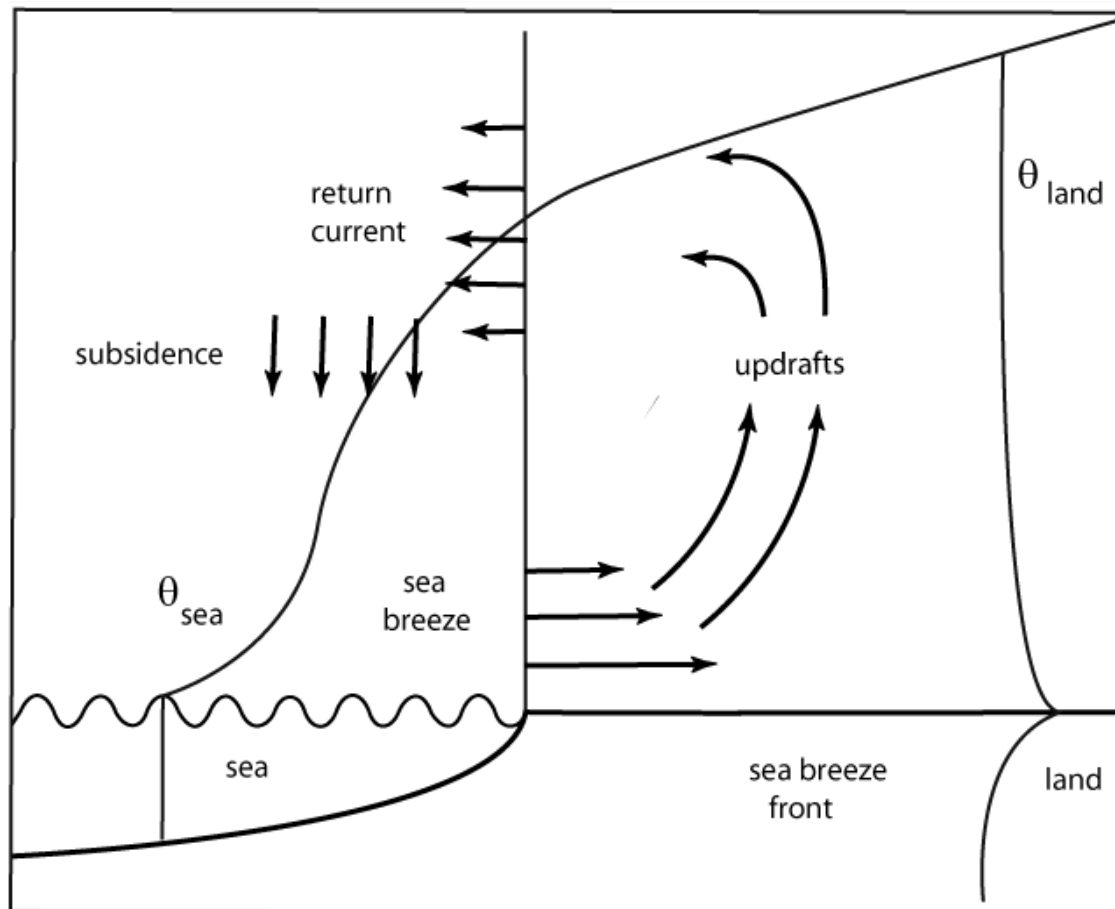
A Sea Breeze normally starts in the morning, a few hours after sunrise, when the solar radiation indirectly heats the boundary layer over land. A classical explanation for the development of a Sea Breeze is the "Upwards" Theory: the differential heating between land and sea leads to the development of a horizontal pressure gradient above the coastal areas (mass convergence of rising air), setting in a flow from land towards sea (left side of scheme above). This upper flow is called a "return current", even though it may develop before the actual Sea Breeze. The mass divergence and resulting pressure fall over land and the convergence and pressure rise over the sea initiate the Sea Breeze close to the surface



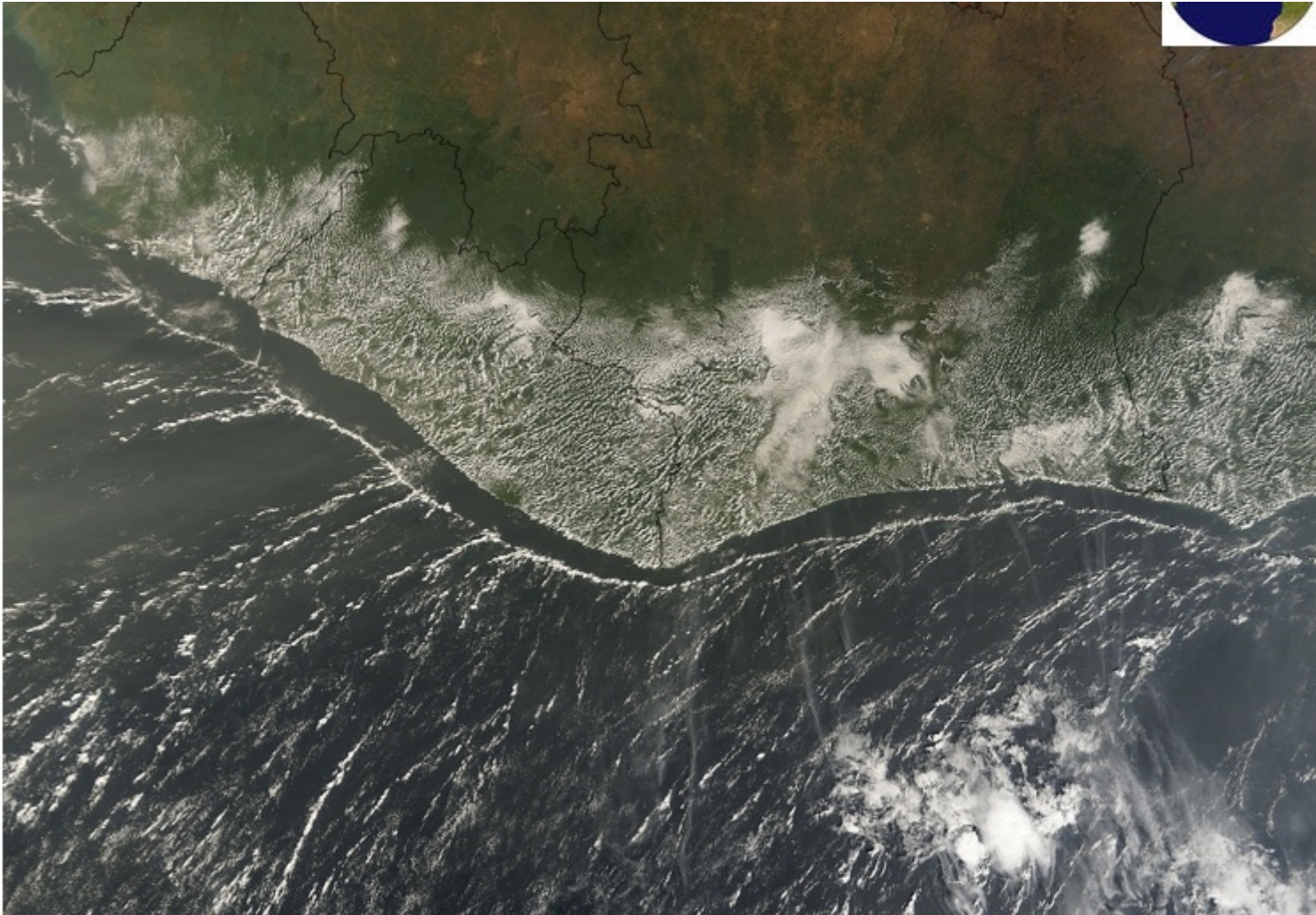
Fronte generato dalla brezza

A Sea Breeze has a typical vertical dimension of 300 – 1000m, with the strongest wind speed at 50 to 200m above the surface. The return current is about twice as thick as the Sea Breeze, leaving for the total vertical dimension of the Sea Breeze circulation typically about 1 to 3 km. In the return current almost the same amount mass of air is returned to the sea, compared with the mass of air transported inland by the Sea Breeze wind.

The scheme in the image above presents all horizontal and vertical processes. At the landside part of the Sea Breeze the vertical speed generally reaches a few m/s and if the LCL is reached convective clouds develop.

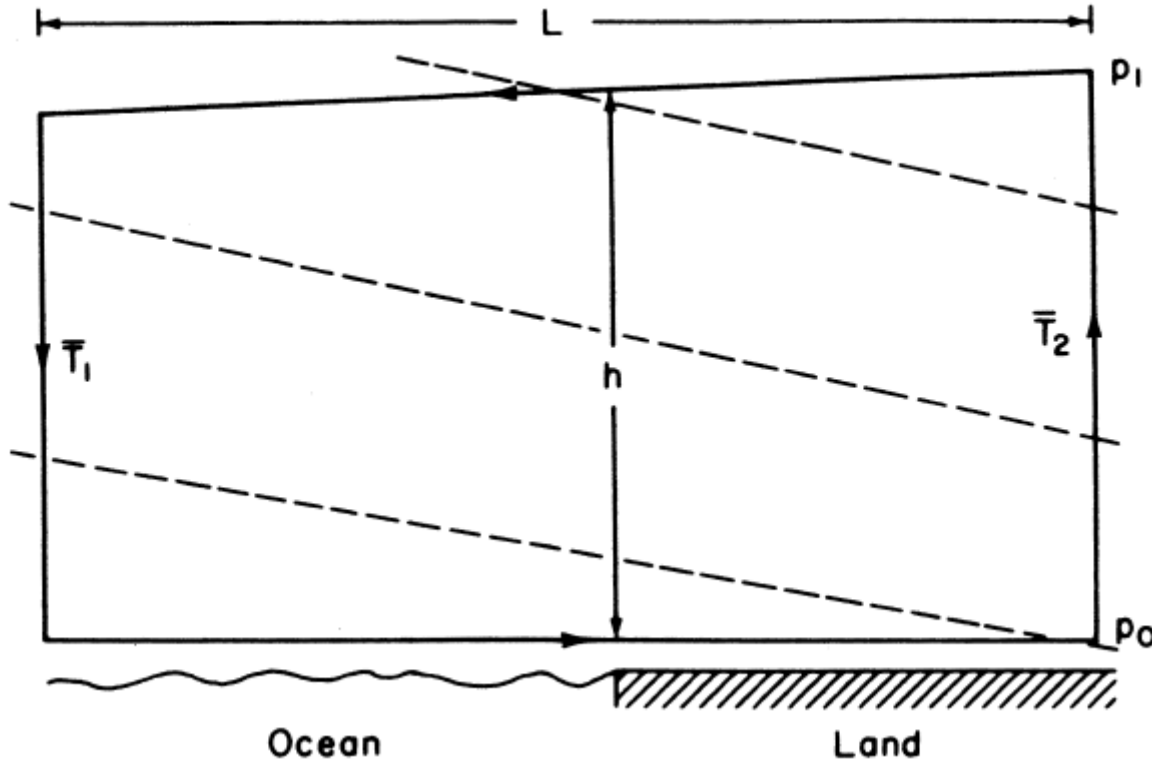


Fronte generato dalla brezza



This MODIS satellite image captures weak convection paralleling the west coast of Africa at 1055Z on December 17, 2006. The convection marked the land-breeze front.

Brezza di mare: un semplice modello basato sul concetto di circuitazione



$$\frac{DC_a}{Dt} = - \oint RT d \ln p$$

$$\frac{DC_a}{Dt} = R \ln \left(\frac{p_0}{p_1} \right) (\bar{T}_2 - \bar{T}_1) > 0$$

$$\frac{D\langle v \rangle}{Dt} = \frac{R \ln(p_0/p_1)}{2(h+L)} (\bar{T}_2 - \bar{T}_1)$$

$\langle v \rangle$ be the mean tangential velocity along the circuit,

If we let $p_0 = 1000$ hPa, $p_1 = 900$ hPa, $\bar{T}_2 - \bar{T}_1 = 10^\circ$ C, $L = 20$ km, and $h = 1$ km, (4.7) yields an acceleration of about $7 \times 10^{-3} \text{ m s}^{-2}$. In the absence of frictional retarding forces, this would produce a wind speed of 25 m s^{-1} in about 1 h. In reality, as the wind speed increases, the frictional force reduces the acceleration rate, and temperature advection reduces the land–sea temperature contrast so that a balance is obtained between the generation of kinetic energy by the pressure-density solenoids and frictional dissipation.

Effetti della vegetazione sullo strato limite atmosferico

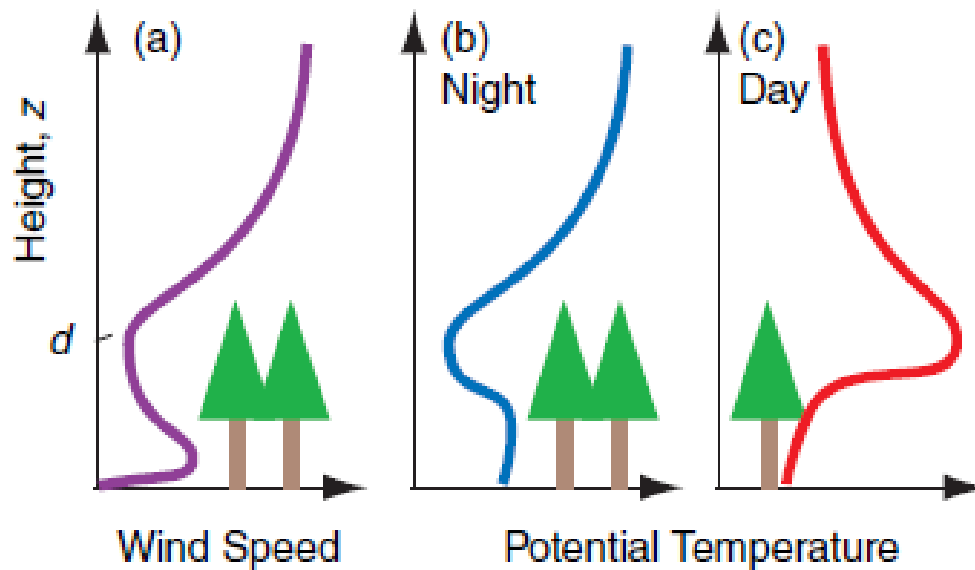


Fig. 9.37 (a) Wind speed in a forest canopy. (b) Potential temperature profile on a clear night with light winds. The level d represents the displacement distance of the effective surface for the logarithmic wind profile in the surface layer above the top of the canopy. (c) Same as (b), but on a sunny day. [Courtesy of Roland B. Stull.]



Effetti degli insediamenti abitativi sullo strato limite atmosferico

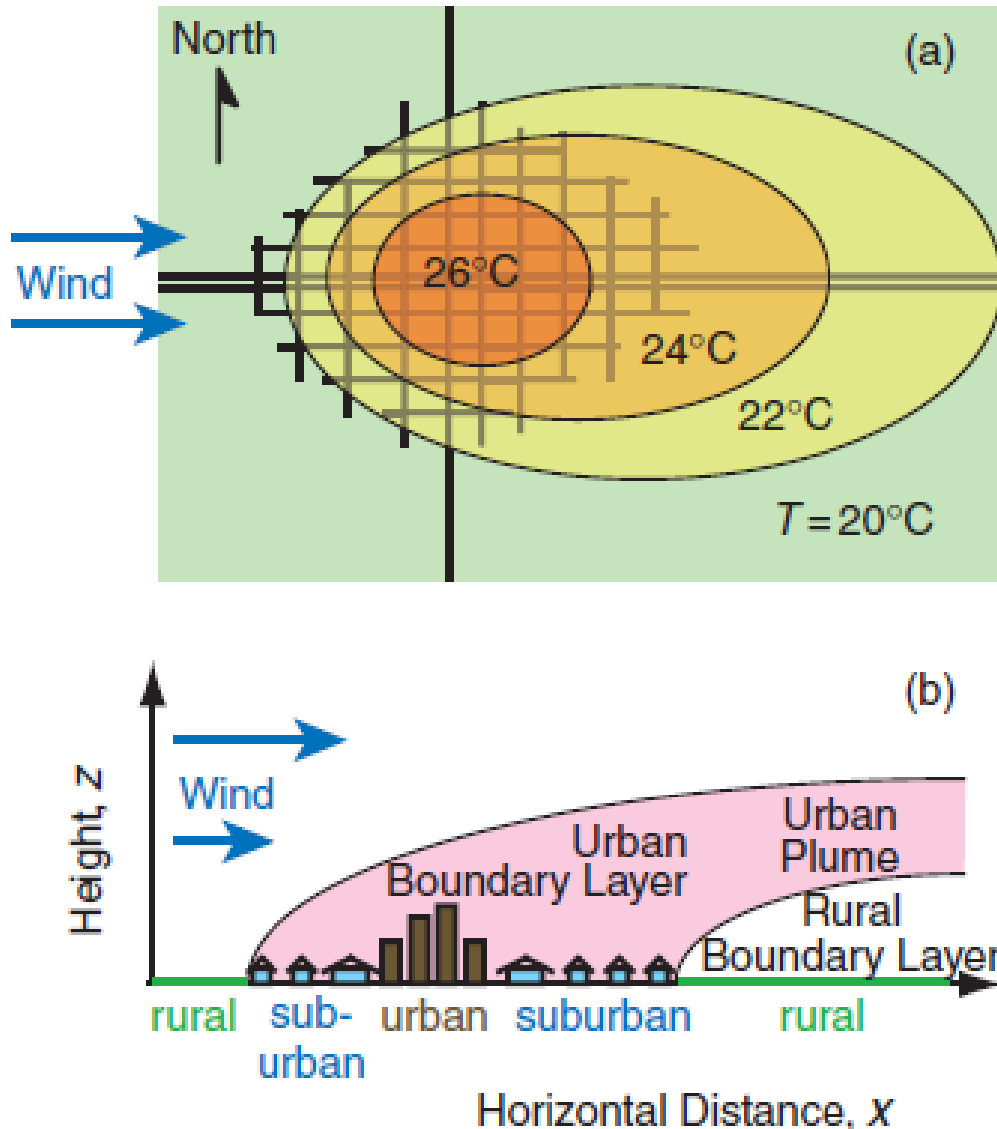
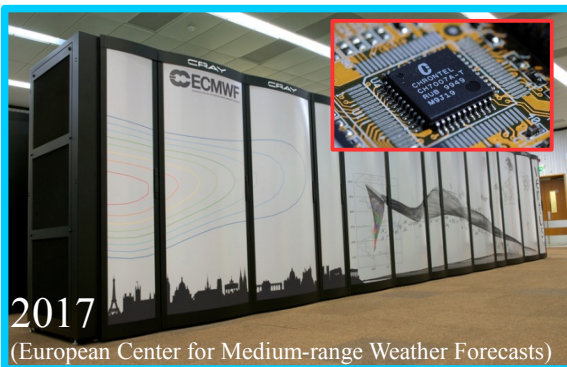
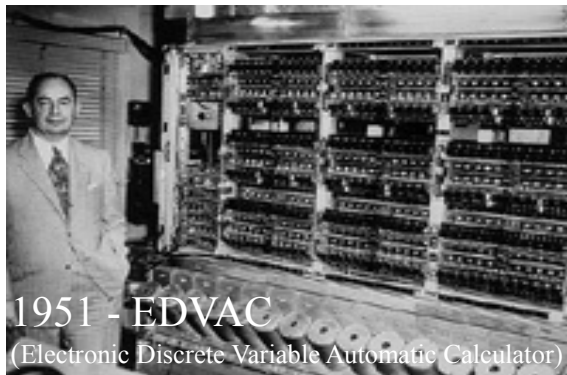
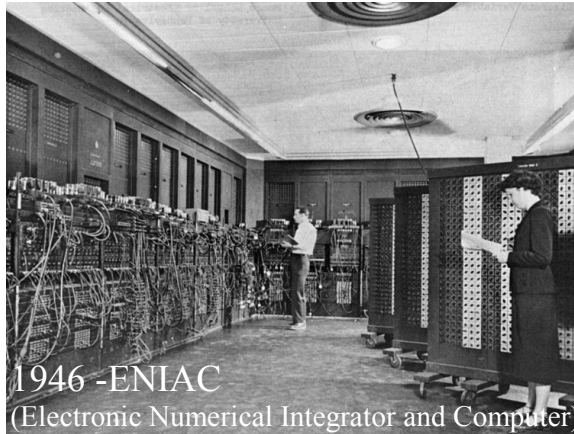


Fig. 9.38 (a) Urban heat island effect, with warmer air temperatures over and downwind of a city. Air temperature excess in the city core can be 2 to 12 °C warmer than the air upstream over rural areas. The grid of lines represents roads. (b) Vertical cross section through a city. The urban plume includes excess heat as well as increased pollution. [Adapted from T. R. Oke, *Boundary Layer Climates*, 2nd Ed., Routledge, New York (1987).]



Simulazioni numeriche dei fenomeni atmosferici: storia di uomini e macchine

Calcolatori



Simulazione di una parte dell'atmosfera 700km x 700km per 2 giorni in 24 ore di calcolo



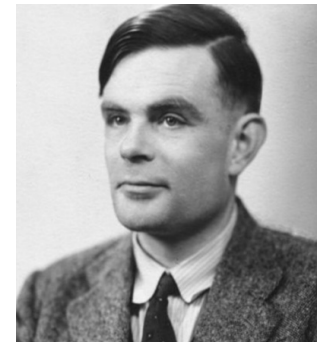
51 Simulazioni su tutta l'atmosfera 12km x 12km per 10 giorni in 1 ora di calcolo

Persone

1922 - Richardson L. F.
Weather Prediction by Numerical Process
La teoria della soluzione delle equazioni




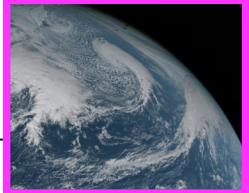
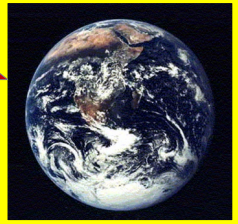




1936 - Alan Turing
On computational numbers with applications ...
La teoria del calcolatore digitale



1951 - John von Neumann
Prime previsioni meteorologiche numeriche
Lo sviluppo del calcolatore digitale programmabile

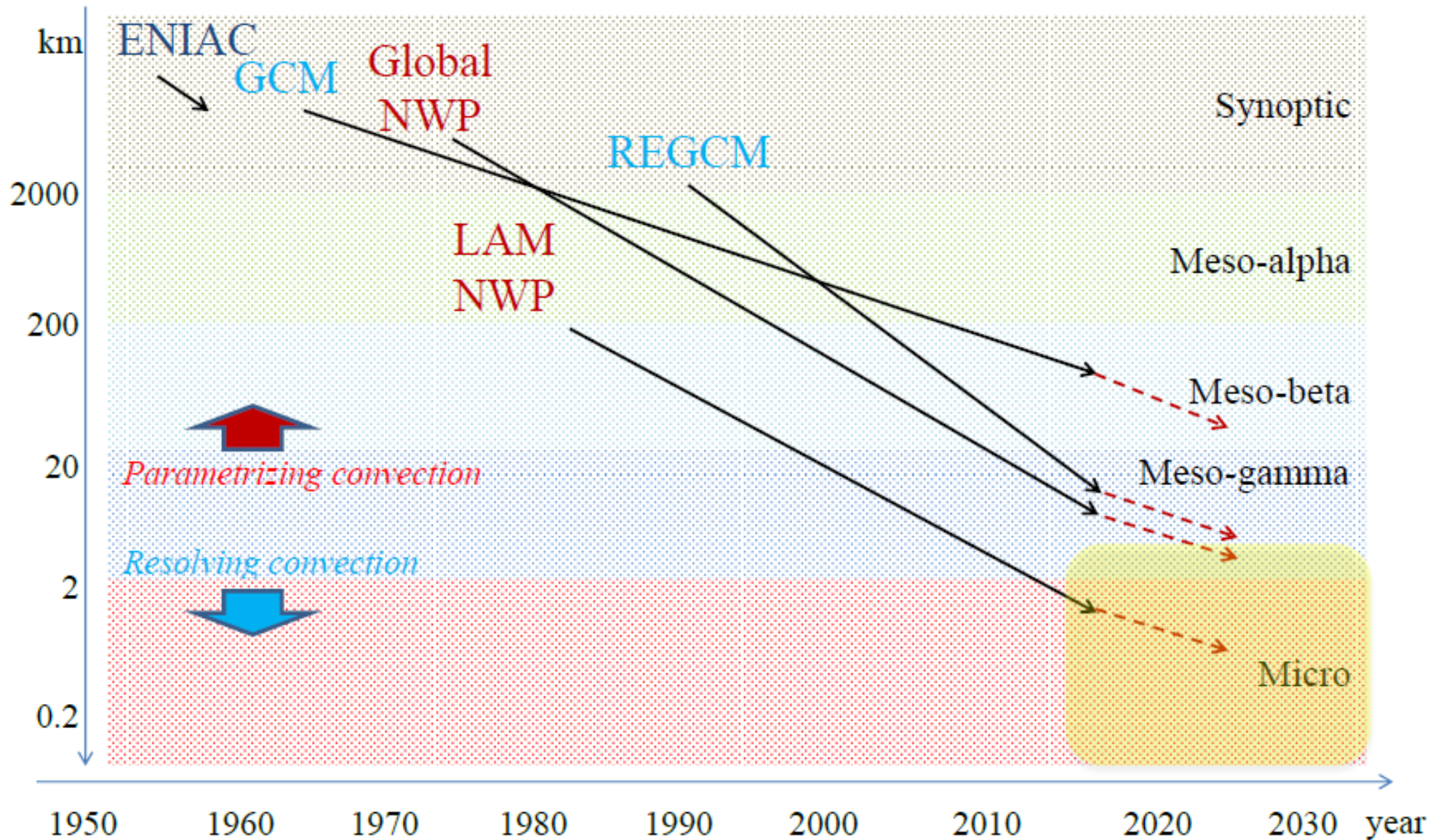


I fenomeni atmosferici alle diverse scale sono interagenti

scale		minuti	ore	giorni	settimane	mesi	anni	km
Macro	α					onde planetarie	riscaldamento globale, EL Nino/La Nina	>10000
	β					cicloni extra-tropicali, onde barocline		2000
Meso	α			frontogenesi	cicloni tropicali, fronti			200
	β			Linee di gruppo, supercelle, temporali				20
	γ			downburst, grandinate			2	
Micro	α		convezione profonda, tornadoes,					0,2
	β		dust devils, strato limite					0,02
	γ	turbolenza					< 0,02	

Cosa è cambiato dai tempi di ENIAC nell'approccio computazionale

Dai tempi di ENIAC (1950) abbiamo aumentato la risoluzione spaziale e questa è ancora **la strategia di sviluppo e di miglioramento dei modelli atmosferici**



Cosa implica l'aumento della risoluzione spaziale dei modelli

Aumentare la risoluzione spaziale significa **simulare processi atmosferici su più scale atmosferiche**

Al modello chiediamo di:

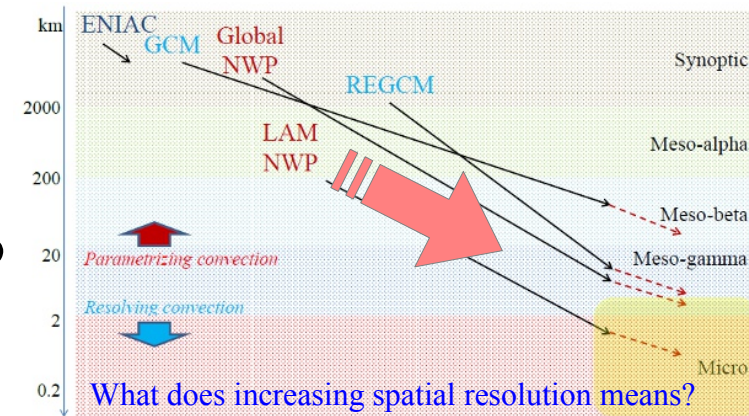
- Trattare più scale atmosferiche
- Simulare le retroazioni tra le scale ed il contorno
- Tenere conto delle condizioni al contorno

Il modello ci chiede:

- Condizioni iniziali più dettagliate e aderenti alla realtà
- Condizioni al contorno più dettagliate
- Più potenza di calcolo

Aumentare la risoluzione, ad oggi ci ha dato:

- Previsioni del tempo più realistiche
- Scenari climatici più dettagliati
- Aumentata capacità di soddisfare gli utenti



Significa

Migliorare i modelli significa migliorare:

- dinamica
- fisica
- assimilazione dati
- processi al contorno
- calcolo

Come valutare se uno schema di strato limite è realistico: le simulazioni e le medie annuali

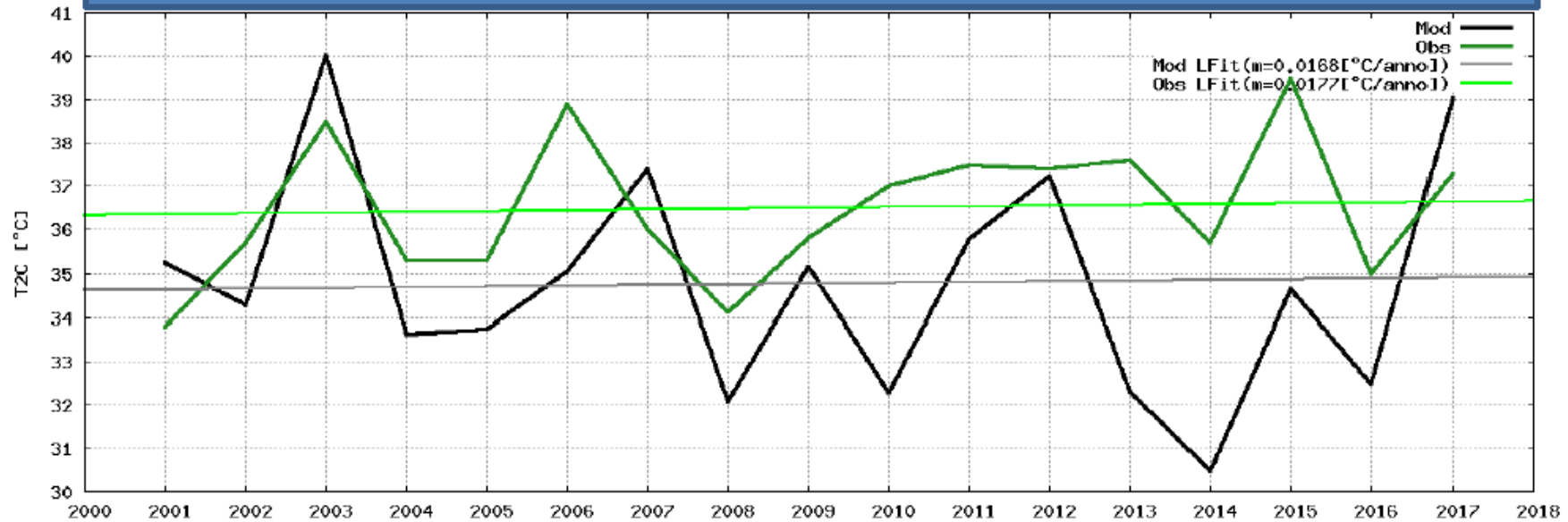
Medie annuali delle medie giornaliere di Temperatura a 2 m ad Udine



Modello numerico WRF, risoluzione spaziale 2 km x 2 km, dominio AlpeAdria, condizioni al contorno ECMWF 50 km x 50 km su dominio continentale

Come valutare se uno schema di strato limite è realistico: l'utilizzo degli estremi

Massime annuali delle massime giornaliere di Temperatura a 2 m ad Udine

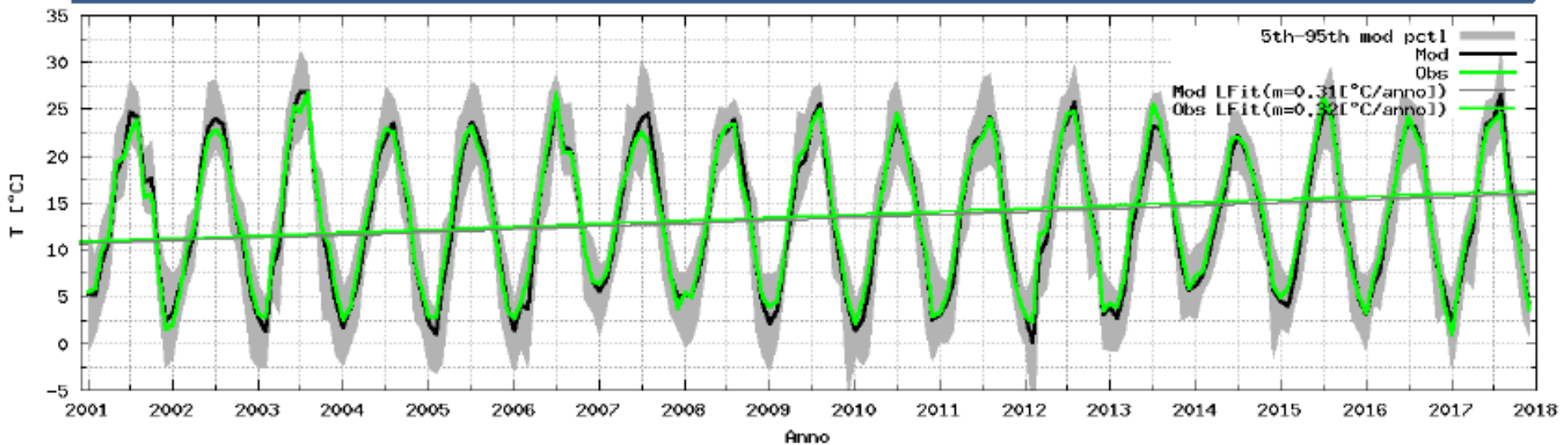


Minime annuali delle minime giornaliere di Temperatura a 2 m ad Udine

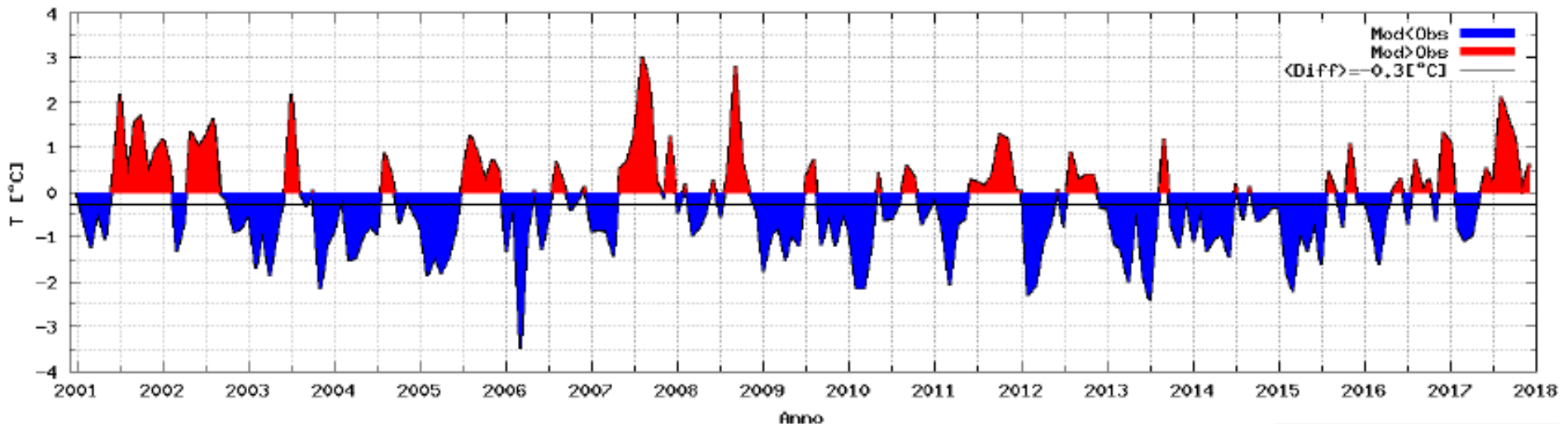


Come valutare se uno schema di strato limite è realistico: l'utilizzo dei grafici

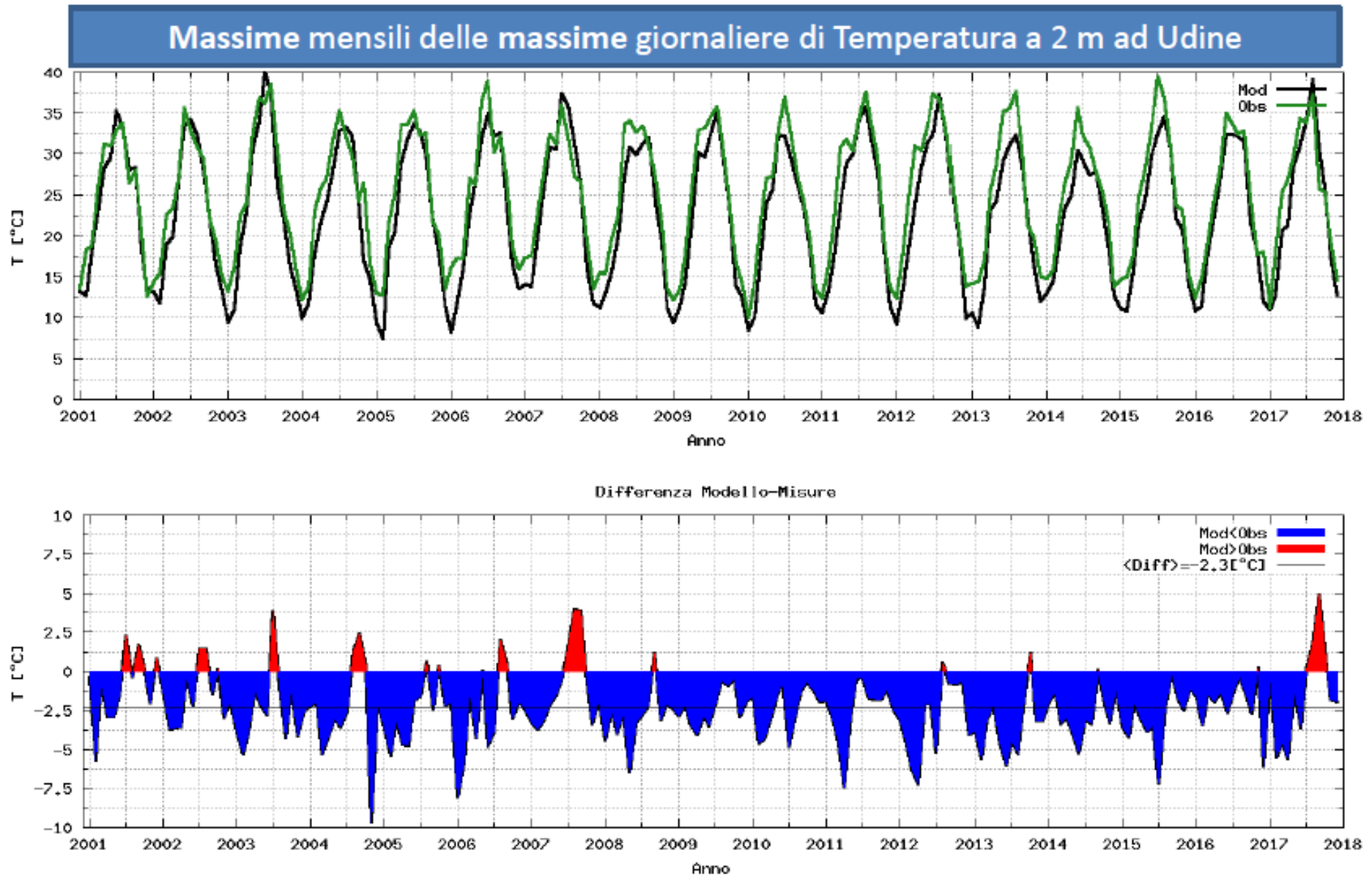
Medie mensili delle medie giornaliere di Temperatura a 2 m ad Udine



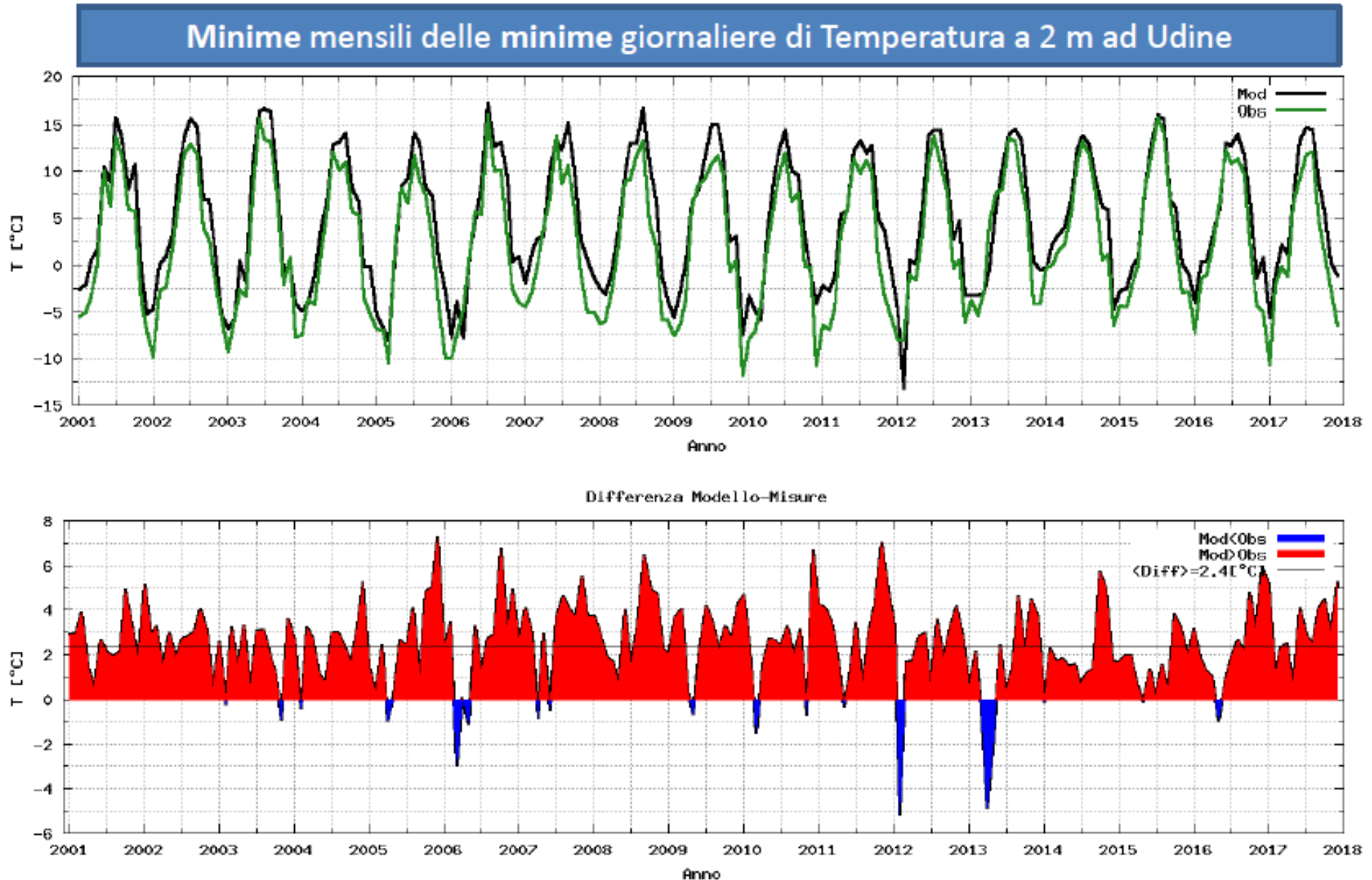
Differenza Modello-Misure



Come valutare se uno schema di strato limite è realistico: l'utilizzo dei grafici per il bias



Come valutare se uno schema di strato limite è realistico: l'utilizzo dei grafici per il bias

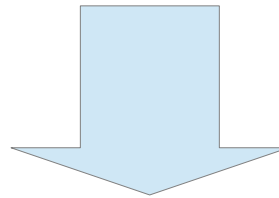


Osservazioni di sintesi

Le temperature simulate hanno una distribuzione statistica con code meno pronunciate rispetto a quella reale

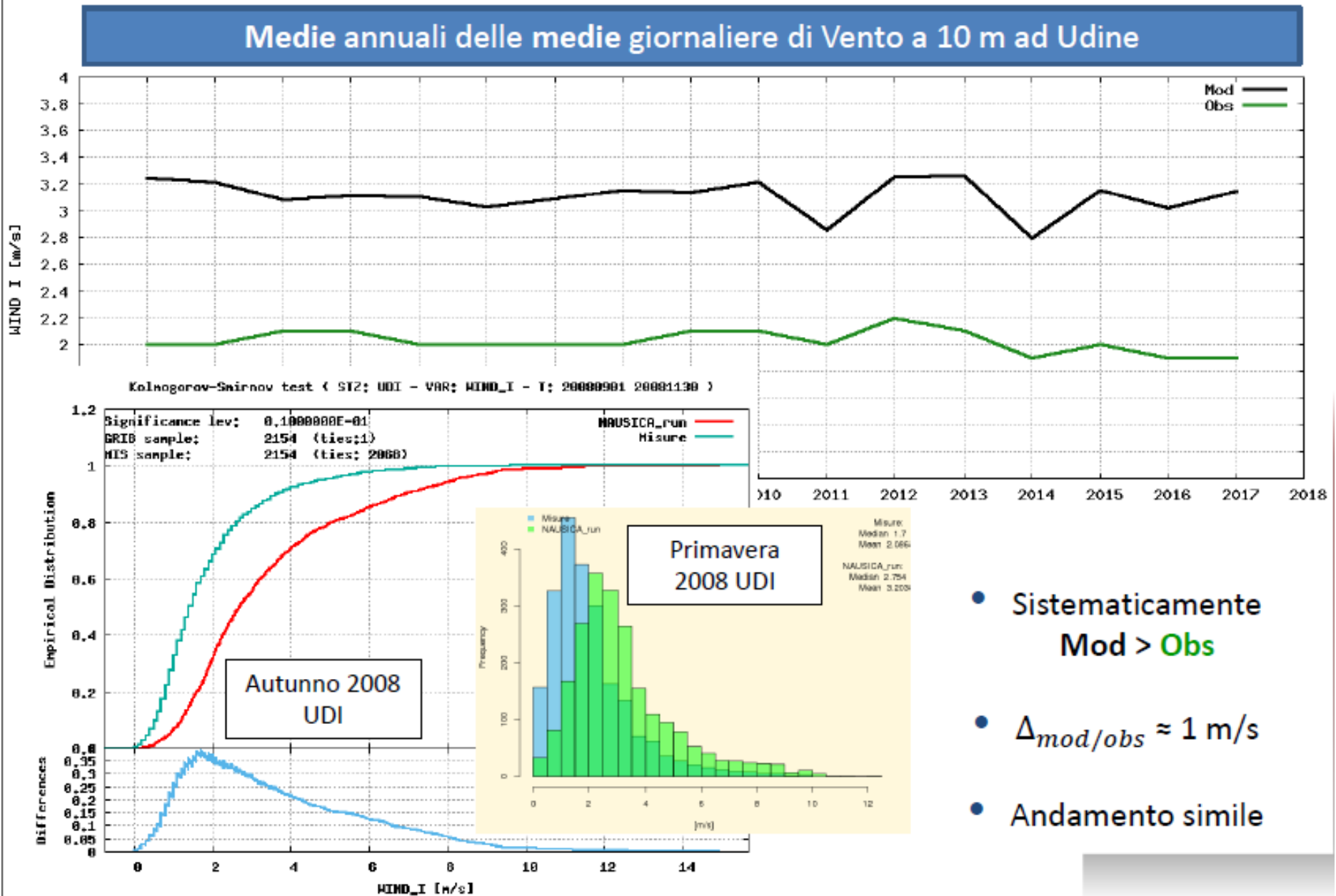
Il bias sui valori medi non è particolarmente importante

La tendenza decennale simulata e quella misurata coincidono



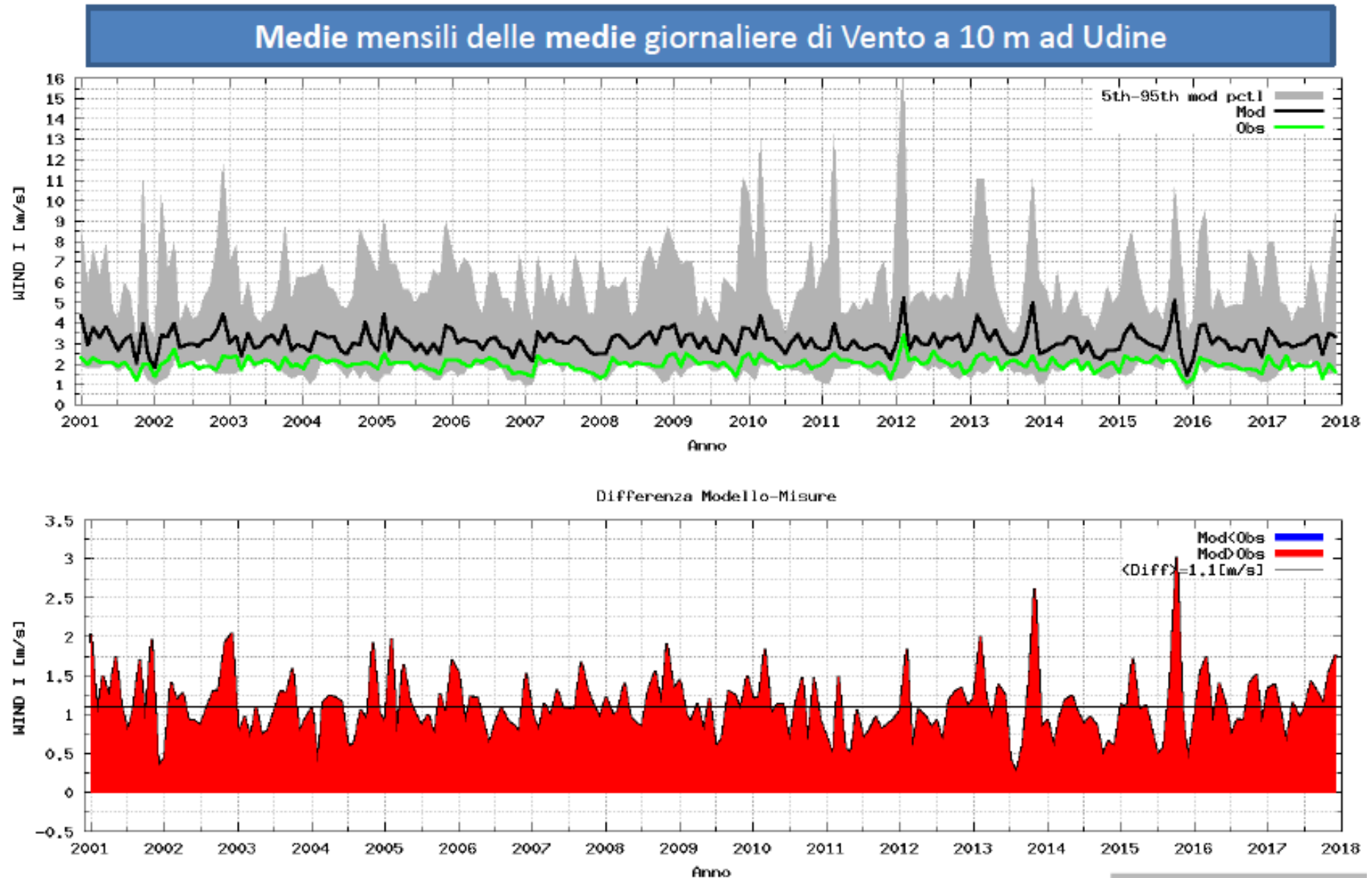
Si deve lavorare sul surface layer probabilmente il rimescolamento turbolento è sovrastimato

Problemi di simulazione del vento a 10 metri dal suolo: sovrastima



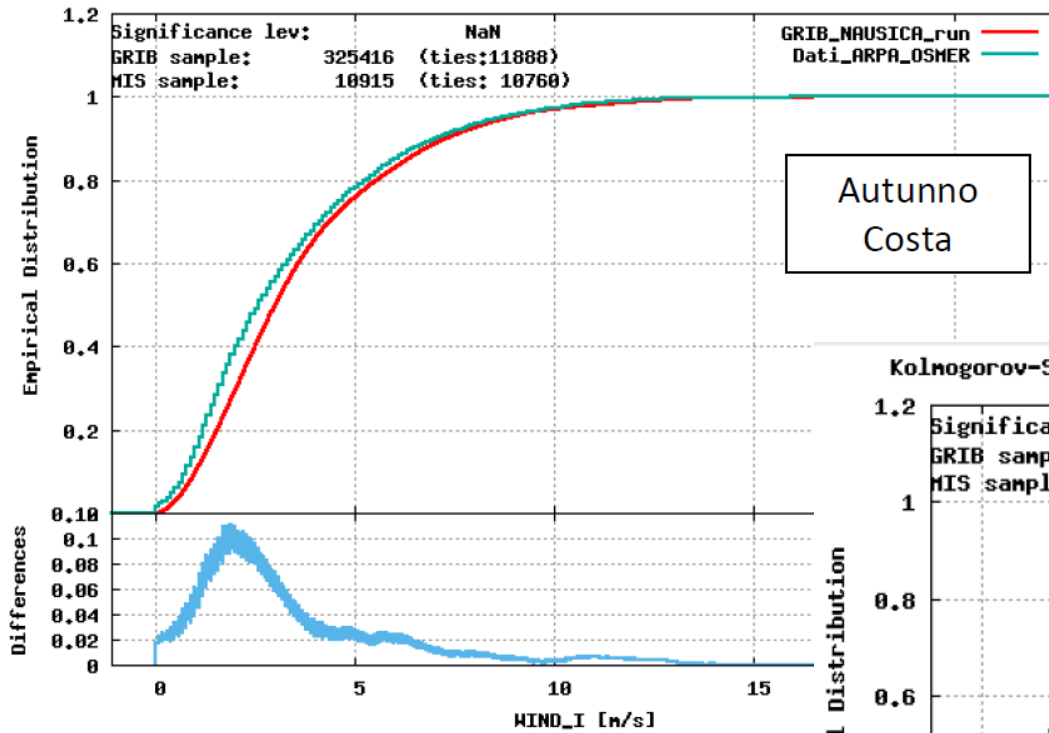
- Sistemáticamente **Mod > Obs**
- $\Delta_{mod/obs} \approx 1 \text{ m/s}$
- Andamento simile

Problemi di simulazione del vento a 10 metri dal suolo: stima del bias

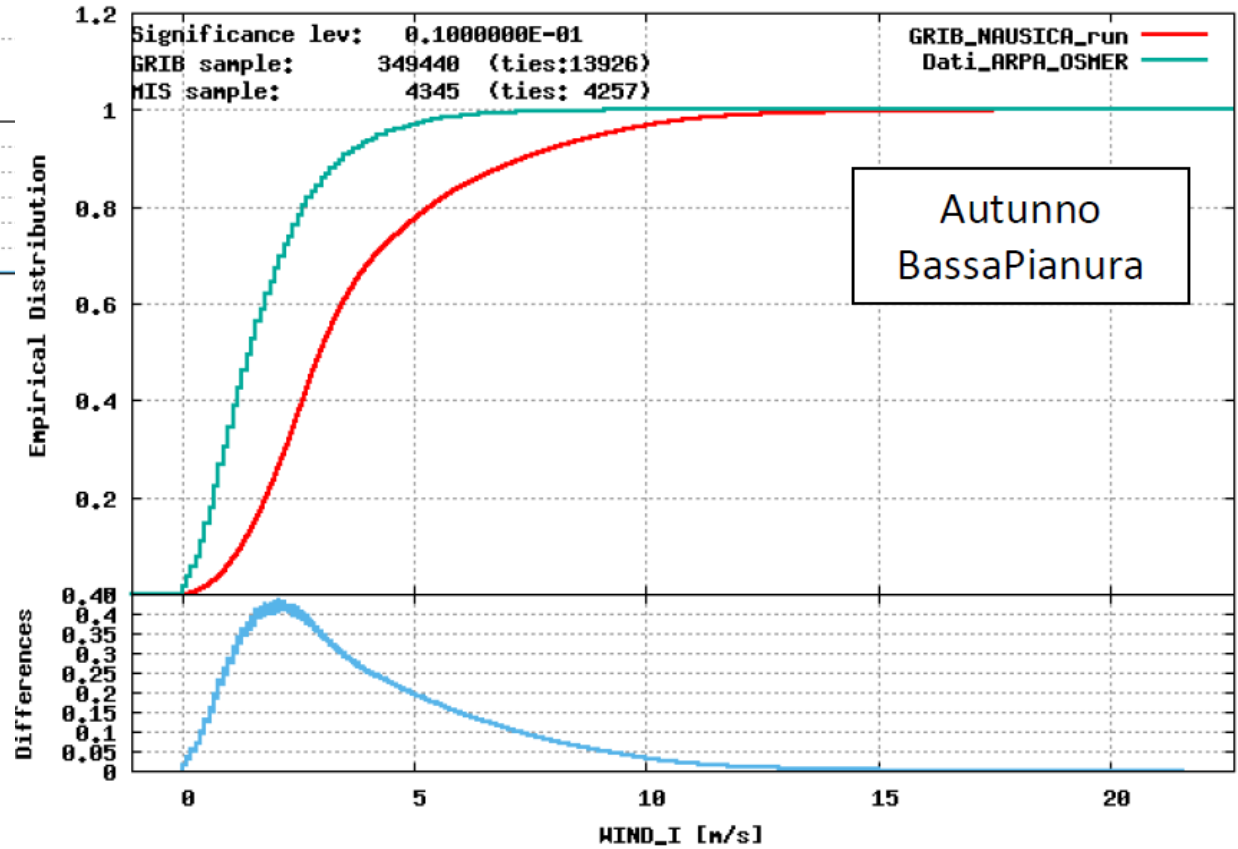


Dove stanno i problemi dei venti simulati nel surface layer?

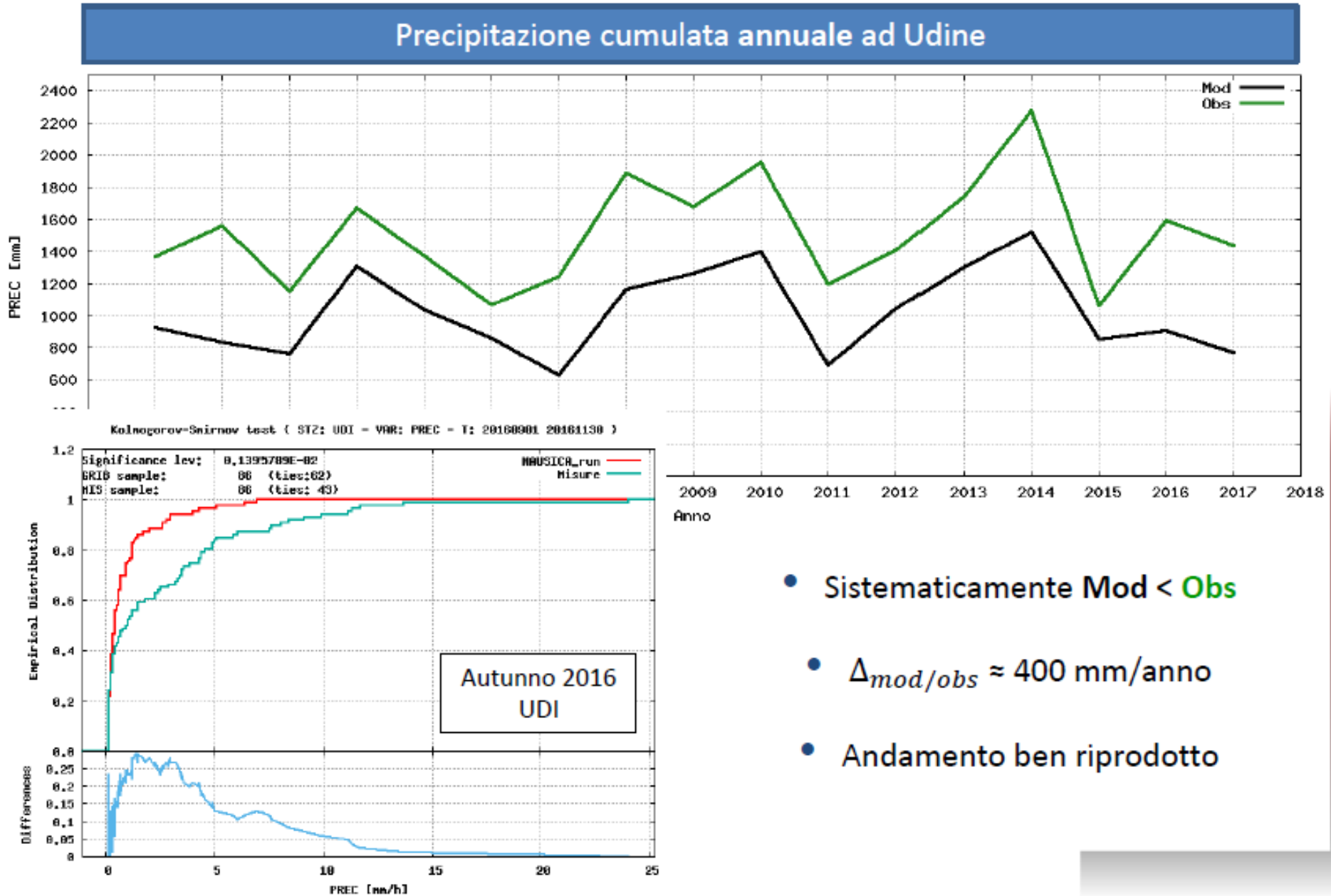
Kolmogorov-Smirnov test (AREA codice AA051501 - VAR: WIND_I - T: 20140901 20141130)



Kolmogorov-Smirnov test (AREA codice AB050301 - VAR: WIND_I - T: 20100901 20101130)

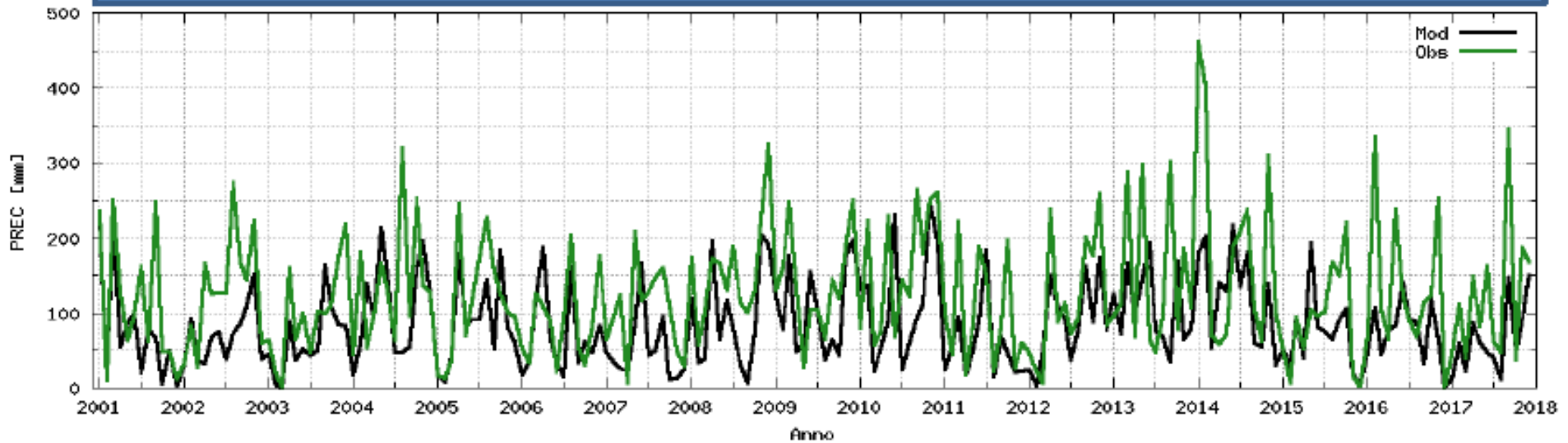


Problemi di simulazione delle precipitazioni deboli: stima del bias annuale

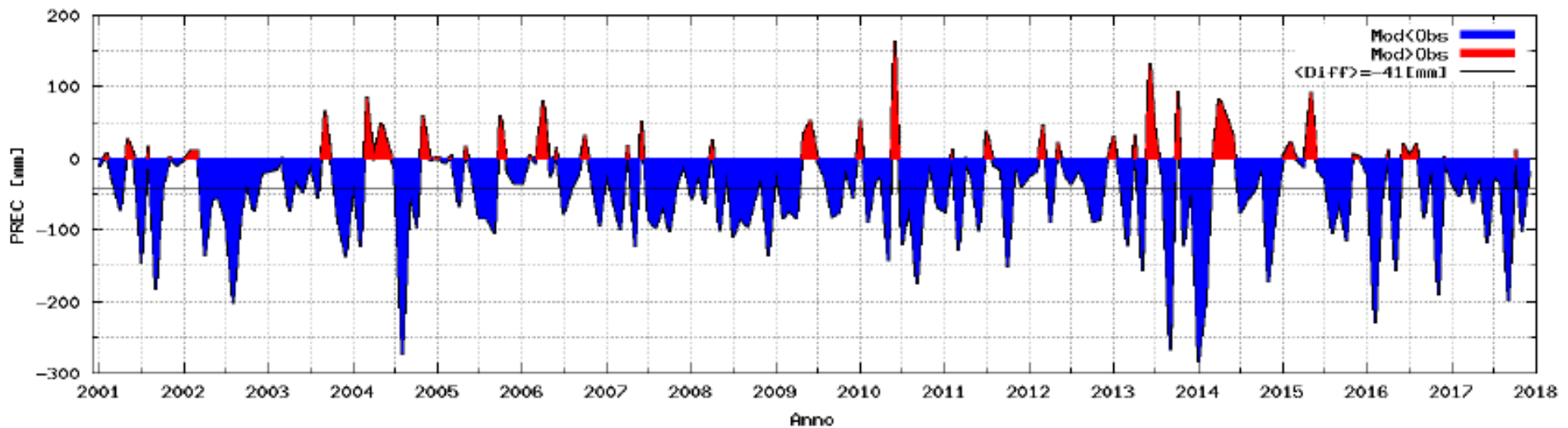


Problemi di simulazione delle precipitazioni deboli: è possibile una correzione sistematica?

Precipitazione cumulata mensile ad Udine

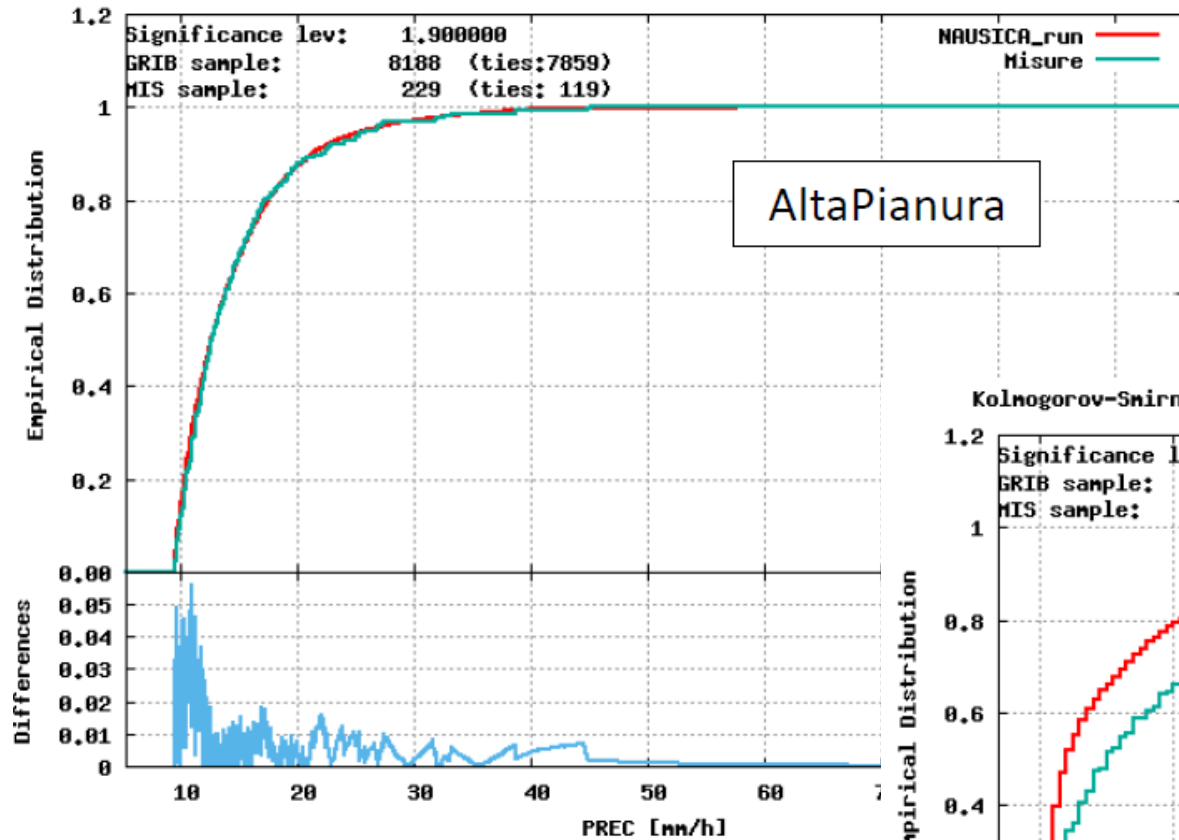


Differenza Modello-Misure



I problemi con le precipitazioni deboli: non è un problema di surface layer

Kolmogorov-Smirnov test (codice AREA:AD053101 - VAR: PREC - T: 20100101 20101231)

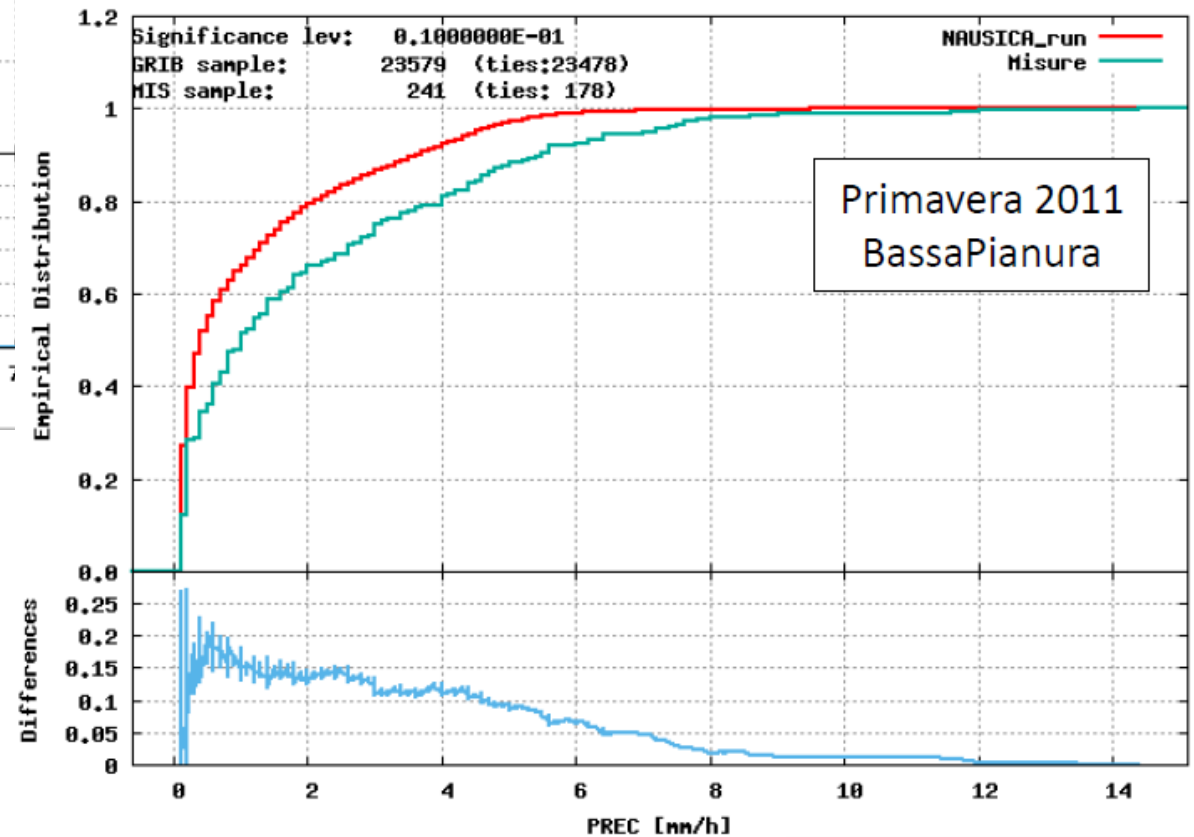


Precipitazione > 10 mm/h

Ottimo accordo su tutte le aree:
BIAS < 1 mm/h



Kolmogorov-Smirnov test (codice AREA:AB050501 - VAR: PREC - T: 20110301 20110531)



Tutte le precipitazioni orarie



Il ruolo dello strato limite atmosferico nei modelli climatici



and the effective heat capacity of the system¹¹. The effective heat capacity of the atmosphere is defined by the region of turbulent mixing through which the heat is mixed i.e. it is defined by the depth of the ABL^{5,6}. So we can define the near-surface temperature response to forcing through an energy budget model of the form:

$$\frac{d\theta}{dt} = \frac{Q}{\rho c_p h} \longleftarrow \text{Altezza strato limite} \quad (1)$$

where Q [W m^{-2}] is the heat flux divergence within the ABL, h [m] is the depth of the ABL, ρ [kg m^{-3}] is the air density, c_p [$\text{J kg}^{-1} \text{K}^{-1}$] the heat capacity at constant pressure, and θ [K] is the potential temperature. Note that this is a reasonable approximation for well-mixed layers, where the potential temperature is constant with height, but may be more complicated in stable boundary layers where the potential temperature increases with height. An assessment of the

Strato limite e modelli climatici



BY J. TEIXEIRA, B. STEVENS, C. S. BRETHERTON, R. CEDERWALL, J. D. DOYLE, J. C. GOLAZ,
A. A. M. HOLTSLAG, S. A. KLEIN, J. K. LUNDQUIST, D. A. RANDALL, A. P. SIEBESMA, AND P. M. M. SOARES

One of the main components of the climate system is the atmospheric boundary layer, which mediates the interactions between the ocean/land surface and the free atmosphere. Several boundary layer processes are known to have a profound influence on the climate system. Relevant examples include the feedbacks between boundary layer clouds and climate. However, in weather and climate predic-

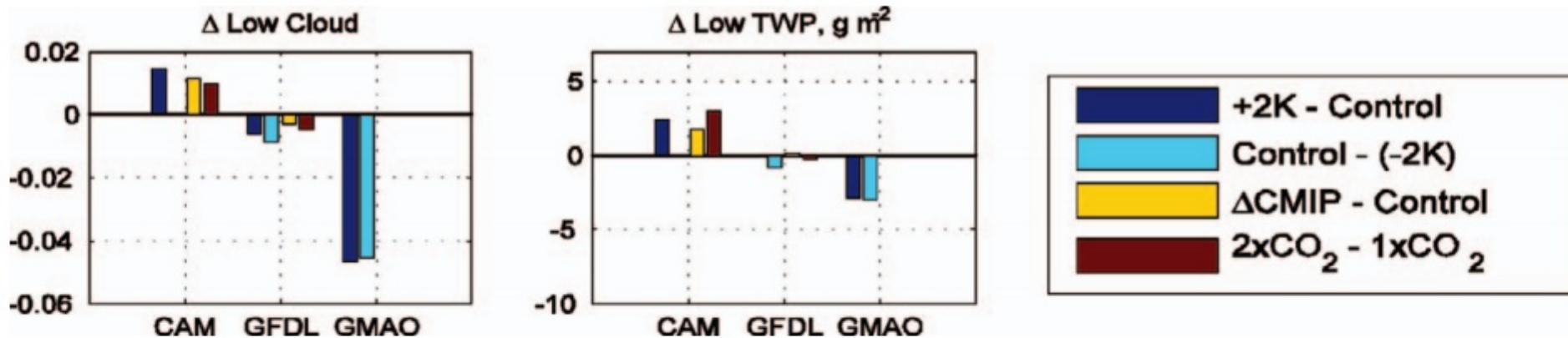
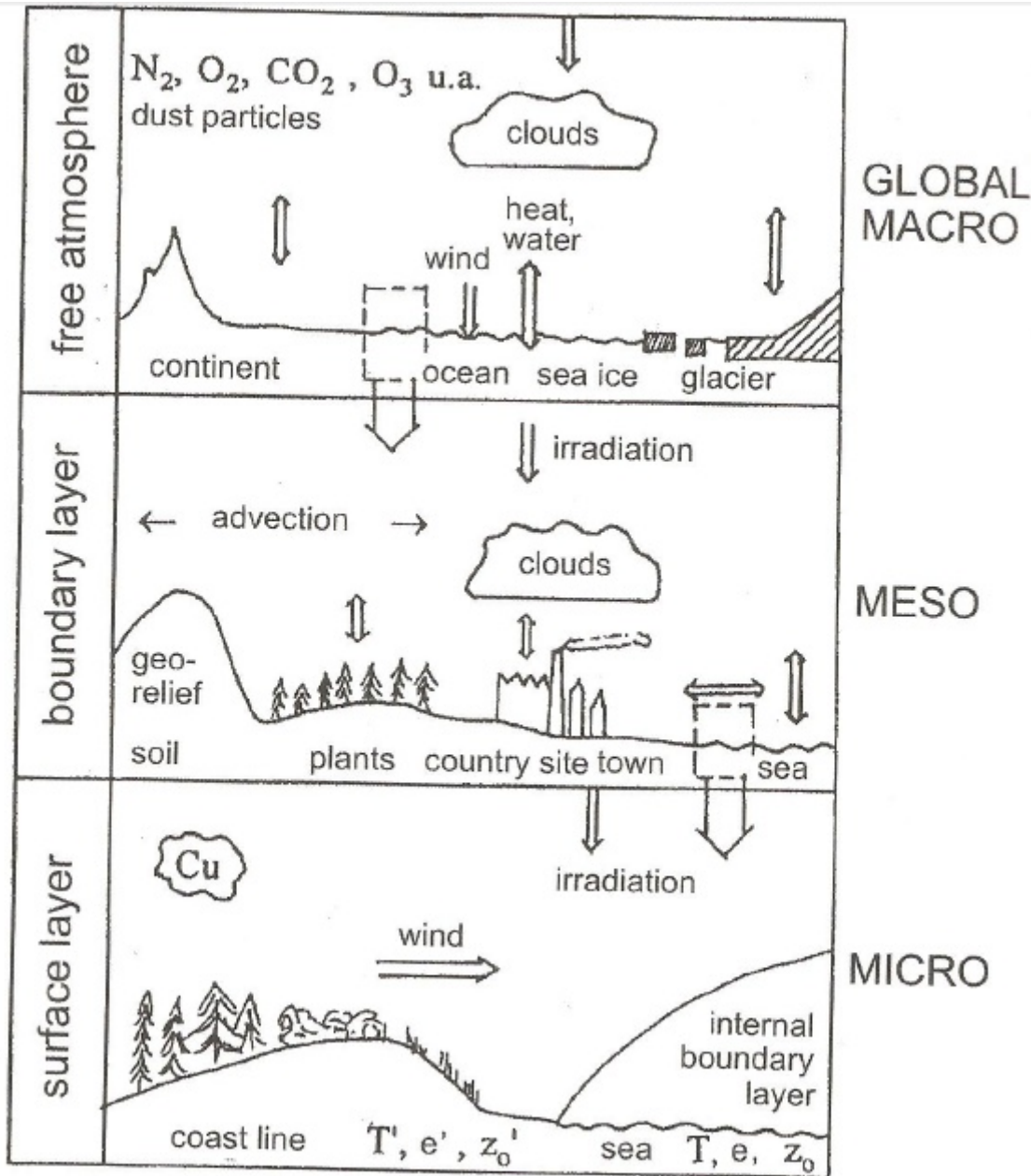


FIG. 1. Changes in low (below 700 hPa) cloud cover and low liquid + ice path (TWP) for different perturbation experiments and for three U.S. climate models (CAM, GFDL, and GMAO). Note the opposite signs, in terms of the low cloud and low TWP sensitivities, between the different models. Three SST perturbation experiments were performed with spatially uniform perturbations of +2 and -2 K and with a spatially and monthly varying SST perturbation (Δ CMIP) (adapted from Wyant et al. 2006).

Ridefinizione delle scale atmosferiche da punto di vista climatico

km	Orlanski (1975)	Böer (1959)	Kraus (1983)	Hupfer (1989)
10^4	makro- β	large climate range	macro range	global climate
10^3	meso- α	local climate range	synopt. range	zonal climate
10^2	meso- β		meso-range	landscape climate
10^1	meso- γ	micro climate range	range	plot climate
10^0	mikro- α		micro range	
10^{-1}	mikro- β		topo range	small scale climate
10^{-2}	mikro- γ			
10^{-3}				border layer climate
10^{-4}				

Ridefinizione delle scale atmosferiche da punto di vista climatico



Processi atmosferici alla scala globale rilevanti per il clima

Processi atmosferici alla mesoscala rilevanti per il clima

Processi atmosferici alla microscala rilevanti per il clima



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Atmospheric Science: An Introductory Survey, J. M. Wallace and P. V. Hobbs, 2nd Edition, 2006



Fronte generato dalla brezza