Lecture 2: Atmospheric Observations

2.1 Radiative properties

This lecture gives a fairly heuristic overview of the atmosphere's structure and meridional transports. The goal is to set the stage for the upcoming, more dynamical (theoretical) lectures, but also to define and explain important terms used in the literature. Thus, this lecture should enable students to read the original literature.

Figure 2.1 (Fig. 6.2 from Peixoto and Oort (1992)) shows the *shapes* of blackbody radiation curves for temperatures of solar and terrestrial surfaces, respectively. Visible light ranges from 0.3 to 0.7 μm. All terrestrial radiation is in the infrared, with virtually no overlap between the two spectra. (Notice that the two curves are not to scale. Per unit area, the sun emits more radiation than the Earth at all wavelengths, see Fig. 3.8 in Kump et al. (1999)). The other panels show absorption as a function of wavelength. Noteworthy are

- The complete absorption of ultraviolet radiation by molecular oxygen (O_2) and ozone (O_3) above 11 km (stratosphere);
- The dominant influence of H_2O (rotation) absorption at wavelengths greater than 20 μm. Hence, H_2O is the most important "greenhouse gas".
- The CO_2 influence at wavelengths just below where H_2O dominates. CO_2 is powerful both because it operates near the maximum of terrestrial radiation and because there is a H_2O "window" (a spectral range where H_2O does not absorb).

Figure 2.2 (Fig. 1.3 from Trenberth et al., itself the Overview Chapter of the 1995 IPCC report, Houghton et al. (1996)) depicts the Earth's radiation and energy balance, and alongside indicates how the greenhouse effect works. The (globally and annually) averaged solar insolation is 342 Wm^{-2}; partly reflected by clouds, aerosols, and atmosphere, partly reflected by the surface. The total reflected portion (reflectivity) is called planetary albedo, around 0.3 for Earth. About 20% of the incoming solar (shortwave, SW) radiation is absorbed by the atmosphere; so 50% is absorbed by the surface. Sensible heat is given off (heat transfer without mass transfer), as well as latent heat (net evaporation, carrying with it the energy needed to turn water from liquid to vapour).

More infrared (longwave, LW) radiation leaves the surface than is absorbed as sunlight. The reason is that only a small portion of LW radiation makes it through the atmospheric window; the lion's share is absorbed. This energy has to go somewhere; it is emitted in equal parts upward and downward. The downward portion helps heat the ground more (see Lecture 5 on radiative-convective models for a thorough treatment).

Exercise

1. *Establish that in Figure 2.2, the fluxes add up to zero at every level that can sensibly be defined.*

2.2 Temperature structure

The radiation comes in as a function of latitude, as seen last time, and initiates atmospheric motions. Figure 2.3 (Fig. 1 from Ghil et al. (1981)) shows the basic network for measuring atmospheric state and motions. Balloons are launched and radio back temperature, humidity, and velocity information. Supplemented by airborne and spaceborne platforms, and in modern times augmented by numerical simulations that ingest the data, estimates can be formed of the state of the atmosphere (Aside: If a numerical model is used, its quality is crucial. Oort's datasets do not incorporate model information).

Figure 2.4 (Fig. 7.5 from Peixoto and Oort (1992)) gives the zonal-mean temperature of the atmosphere, for northern winter (December-January-February, DJF) and northern summer (June-July-August, JJA), and annual-mean. The familiar picture is seen of temperature dropping with height and toward the poles. It is probably less familiar that temperatures increase above a certain height (called the tropopause). This increase occurs because ozone absorbs SW radiation. The most surprising aspect perhaps is that the coldest annual-mean temperatures are found below the tropopause in the tropics. Note that the tropopause is highest in the tropics, for nontrivial reasons.

Figure 2.5 (Fig. 7.6 from Peixoto and Oort (1992)) gives the zonal and annual-mean *potential temperature* of the atmosphere, and the *equivalent potential temperature*. The potential temperature, θ, is defined as the temperature an air or water parcel would have if it were brought, *adiabatically*, to a reference height (by default, the surface, z=0). If an air parcel was brought to the surface from high up, it would be compressed; the mechanical energy involved would go into the temperature increase. Mathematically,

$$\theta(z, z_r) \equiv T(z) + \int_{z_r}^{z} \Gamma dz', \qquad (2.1)$$

where z is height, z_r the reference height, and the adiabatic lapse rate, Γ, is defined as

$$\Gamma \equiv -\frac{\partial T}{\partial z}(z, \eta). \qquad (2.2)$$

Notice that Γ is positive if temperature drops with height. The argument, η, is entropy per unit mass and indicates that the process is adiabatic (constant entropy; the partial derivative is taken with respect to z). It follows that θ is constant under adiabatic displacements:

$$\frac{\partial}{\partial z}\theta(z, \eta) = \frac{\partial}{\partial z}T(z, \eta) + \Gamma(z, \eta) = 0, \qquad (2.3)$$

by virtue of (2.2), where Leibniz's rule has been employed in the first equality. Owing to compressibility, θ is the important quantity for static stability: When an air parcel rises, it expands and cools; the question is, does it cool by more than the surrounding air becomes colder? Thus, one should not say, "warm air rises", but "potentially warm air rises". Notice that the highest *temperature*, T, is near the surface, whereas *potential temperature*, θ, increases with height, indicating that the atmosphere is largely stably stratified. This will be crucial in a moment, when we will try to understand the energy transport by the "Hadley Circulation". The *equivalent potential temperature*, θ_e, incorporates, additionally, that water vapour may be present, which upon cooling could condense and heat the air, making the parcel more buoyant.

2.3 Atmospheric circulation

The most basic dynamical properties of the atmosphere are shown in the next two figures. Figure 2.6 (Fig. 7.15 from Peixoto and Oort (1992)) gives the zonal-mean zonal wind (positive *eastward*, which means *westerlies*), as a function of latitude and depth, for the annual mean, DJF, and JJA. We see the familiar surface pattern of easterlies (trade winds) at low latitudes, westerlies at mid-latitudes, and (barely noticeable) the polar easterlies. The strongest winds, the "jet streams", are found just below the tropopause at mid-latitudes. While the wind patterns are symmetric about the equator, on annual average, they show strong asymmetries with the seasons. Generally, the winter hemisphere shows a considerably stronger jet stream. This has important consequences, to which we will return in Lecture 12.

Figure 2.7 (Fig. 7.19 from Peixoto and Oort (1992)) shows the stream function (streamlines) of the zonally integrated circulation in the atmosphere, often called the mean meridional circulation (MMC) in atmospheric science. For an incompressible fluid (which the atmosphere is *not*, but see below), the meridional stream function is easily defined as follows. We start with the statement for mass conservation (continuity equation) in the incompressible limit,

$$\partial_x u + \partial_y v + \partial_z w = 0 , \tag{2.4}$$

where ∂_x indicates partial derivation with respect to x, and u, v, and w are the velocities in zonal, meridional, and vertical (upward) direction. Zonal integration over an entire basin, from western edge, x_w, to eastern edge, x_E, gives

$$u(x_E) - u(x_W) + \int_{x_E}^{x_W} \partial_y v dx + \int_{x_E}^{x_W} \partial_z w dx = 0 . \tag{2.5}$$

We now assume that both eastern and western coasts run exactly north-south and that the side walls are vertical. This is not necessary for the existence of a stream function, but it simplifies the algebra enormously. Notice that we use Cartesian, rather than spherical, co-ordinates for simplicity as well. If x_w and x_E are independent of y and z, we can interchange integration and differentiation, to obtain

$$u(x_E) - u(x_W) + \partial_y V + \partial_z W = 0 , \tag{2.6}$$

where V and W are the zonally integrated meridional and vertical transports, respectively,

$$V \equiv \int_{x_E}^{x_W} v dx; \quad W \equiv \int_{x_E}^{x_W} w dx . \tag{2.7}$$

For north-south coastlines, $u(x_E) = u(x_W) = 0$; alternatively, in a re-entrant channel, such as the Antarctic Circumpolar Current or the atmosphere, we have $u(x_E) = u(x_W)$. In either case, mass conservation holds in the two-dimensional form

$$\partial_y V + \partial_z W = 0 . \tag{2.8}$$

We now can define the meridional stream function, ψ, according to

$$V(y,z) \equiv -\partial_z \psi(y,z); \quad W(y,z) \equiv \partial_y \psi(y,z), \tag{2.9}$$

as cross-differentiation shows.

Exercises

2. *Derive the meridional stream function for slanting boundaries.*
3. *Convince yourself that, with the sign convention adopted in (2.9), the flow goes clockwise around high values.*

The atmosphere is not even approximately incompressible, but we can utilise that, if one uses pressure instead of height as the vertical co-ordinate, most atmospheric equations look exactly like their incompressible counterparts. (Deriving this is quite elaborate, and we will not attempt it here.) Returning to Fig. 2.7, we see that on annual average, the most conspicuous feature is the strong rising motion around the equator, poleward motion aloft, and subsidence in the subtropics (the very dry sinking air there creates the large desert belts on Earth). This is the Hadley circulation (or Hadley cells). At middle latitudes, there are two cells rotating counter to the Hadley cells, sometimes called Ferrel cells. However, the relatively symmetric annual-mean picture is misleading, concerning the instantaneous circulation patterns. During winter and summer, the tropical circulation is better characterised by a single Hadley cell, with rising in the summer hemisphere and subsidence in the winter hemisphere. The second tropical cell is barely noticeable; two nearly equally strong Hadley cells emerge only upon annual averaging. (N.B.: One of several errors on oceanic and atmospheric circulations in the otherwise excellent book of Kump et al. (1999) is that they sketch double Hadley cells moving around with the seasons, see their Fig. 4-16). The seasonal reversal in the Hadley circulation is, however, crucial for understanding atmospheric energy transport.

2.4 Atmospheric transports

We are interested in energy transport, both in atmosphere and ocean, because the energy transport reduces temperature gradients. But how is it defined? We argue, heuristically, that the "Bernoulli function", B, is the quantity that is being transported. It is defined as

$$B \equiv U + p\alpha + gz + \frac{1}{2}v^2 , \qquad (2.10)$$

where U is internal energy per unit mass, $\alpha \equiv 1/\rho$ is specific volume, and the last two terms are potential and kinetic energy per unit mass, respectively. The Bernoulli function is thus total energy plus the extra term, $p\alpha$. The latter reflects that if one considers the change of total energy in a control volume (say, the atmosphere north of some latitude circle), flow into that volume implies an exchange of mechanical energy (work) if there is a pressure gradient at the boundaries of that volume. This "pressure work term" can be re-written such that it is folded into the transported property, meaning that *enthalpy* per unit mass, $H \equiv U + p\alpha$, enters the calculations. In shorthand, total energy in a control volume changes because of the convergence of "Bernoulli transport" in that volume. This point is elaborated in Appendix 2.A.

In the atmosphere, there is one more contributor to energy transport. The energy that was used to transform liquid water into vapour is available upon condensation and must be taken into account. Before we get to this, let us look at the transport of water vapour in the atmosphere, differentiated both by season and by transport mechanism (Fig. 2.8, which is Fig. 12.12 from Peixoto and Oort (1992)). Alas, before we can do that, we again have to step back and develop one more tool, the breakdown of transports into various components of motion. For an arbitrary transported variable, ϕ, we can write the transport at any given latitude as

$$Q_\phi(y) = \int_z \int_{x_W}^{x_E} v\phi \, dx dz \qquad (2.11)$$

where the limits on the z-integration could apply equally well to atmosphere and ocean; again, for the atmosphere the x-integration would span the entire globe. To simplify the derivation, we assume time-independent velocity and ϕ. We define the zonal average and deviations thereof by

$$[\phi](y,z) \equiv \frac{1}{L_x(y,z)} \int_{x_W}^{x_E} \phi(x,y,z)\,dx; \qquad . \qquad (2.12)$$
$$L_x(y,z) \equiv x_E(y,z) - x_W(y,z); \quad \phi'(x,y,z) \equiv \phi(x,y,z) - [\phi](y,z)$$

By rewriting the quantities in (2.11) as the sums of zonal means and deviations thereof, we obtain

$$Q_\phi(y) = \int_z L_x [v][\phi]\,dz + \int_z L_x [v'\phi']dz\,. \qquad (2.13)$$

By definition, the mixed terms drop out. The first term in (2.13) is called the "overturning contribution" in oceanography and "mean meridional circulation" in meteorology; the second term is called, confusingly, "gyre contribution" in oceanography but "eddy contribution" in meteorology. In oceanography, the term "eddy contribution" is reserved for unresolved, "diffusive" transport. The meteorological "eddy contribution" is further broken down into time-dependent and time-mean contributions, called "transient eddies" and "standing eddies", respectively. The former are travelling highs and lows, while the latter are mainly the large latitudinal excursions of the jet stream, caused by land-sea contrasts and mountain belts.

Exercises

4. *Derive (2.13).*
5. *Derive the breakdown of the eddy contribution into transient and standing eddies.*

At last, we are equipped to analyse Fig. 2.8. The total water vapour transport is dominated by the tropical seasonal cycle, with northward transports crossing the equator in northern summer, and southward transports in northern winter. Year-round, there is the expected convergence of moisture upon the equator (more precisely, slightly north of the equator). The tropical signal is almost entirely explained by MMC contributions, in turn readily understood by looking at the seasonal fluctuations of the Hadley circulation. Low-level air contains more moisture than upper-level air; near the surface, the air moves from the winter hemisphere to the summer hemisphere, where it rises and returns.

At middle latitudes, the MMC plays virtually no role in atmospheric water vapour transport. But the combination of standing and transient eddies transports water vapour northward in the northern hemisphere, overall twice as much in winter than in summer (notice the different scales). Figure 2.9 (Fig. 12.16 of Peixoto and Oort (1992)) shows the divergence of water vapour transport, which is practically equal to the sum of evaporation minus precipitation, E-P. On annual average, we see the expected pattern of net rainfall in the tropics, net evaporation in the subtropics, and net rainfall again at middle latitudes. The seasonal cycle is as follows: More net winter rainfall in middle latitudes, more net winter evaporation in the subtropics, and more net summer precipitation in the tropics.

Before looking at the energy transport, a note of caution is in order. In (2.13), we can perform a split analogous to (2.12), but in the vertical. This would be equivalent to referencing all quantities to their section-mean, denoted by angle brackets,

$$\int_z L_x [v][\phi] dz = \int_z L_x \left[\langle v\rangle + \tilde{v}\right]\left[\langle \phi\rangle + \tilde{\phi}\right] dz = A\langle v\rangle\langle \phi\rangle + \int_z L_x [\tilde{v}]\left[\tilde{\phi}\right] dz\,, \qquad (2.14)$$

where A is the section area. We see that if there is a net flow through a section, the section-mean value of ϕ matters, which is nonsensical in the case of enthalpy transport. It would depend on the temperature scale chosen (Celsius or Kelvin?). Hence, one should always aim at analysing transports across sections with no net flow. If this is unattainable, one must remember that is the *convergence* of energy transport that gives rise to observable phenomena (heating or cooling).

Exercise

6. *Derive (2.14).*

Latent heat transport is water vapour transport, multiplied by the latent heat of vaporisation, so Fig. 2.8 contains the complete information about latent heat transport. Figure 2.11 (13.6 from Peixoto and Oort (1992)) shows *sensible heat transport* (transport of enthalpy) in the atmosphere. It is very similar to water vapour transport in low latitudes, with transport by the Hadley cell from the winter hemisphere to the summer hemisphere. In middle latitudes, again the eddies dominate, but the structures are more pronounced than for water vapour transport. Transports in winter are much stronger than in summer.

At low latitudes, we obtain the seemingly paradoxical result that the Hadley cell transports energy from the winter hemisphere – which is being cooled by radiation – to the summer hemisphere, which is being heated by radiation. However, this puzzle is resolved by looking at the transport of potential energy (Fig. 2.12, which is Fig. 13.7 from Peixoto and Oort (1992)). Here, the eddies play virtually no role, and the Hadley cell transport is from summer hemisphere to winter hemisphere. Basically, this is because the air entering the winter hemisphere is high up and has high potential energy. In appendix 2B we show that for an ideal gas, the Bernoulli function can be written as

$$B \equiv c_p \theta + \frac{1}{2} v^2 . \tag{2.15}$$

According to Fig. 2.5, potential temperature is higher in the upper branch of the Hadley cell than in the lower branch. Consequently, the energy transport is in the same direction as the upper branch, that is, from the summer to the winter hemisphere. (The transport of kinetic energy is negligible). Strictly speaking, this argument should be made for the equivalent potential temperature, taking the latent heat into account, but we will not attempt this here.

Figure 2.13 (13.11 from Peixoto and Oort (1992)) shows that indeed the transport of potential energy outweighs the sum of sensible and latent heat transport. On annual average, the total energy transport by the atmosphere is fairly antisymmetric about the equator. Winter transports are larger than summer transports, and, transport across the equator is from summer to winter hemisphere. On annual mean, the atmospheric energy transport in the tropics is smaller than that in the extratropics. At middle latitudes, the eddies combined dominate the MMC.

2.5 Oceanic energy transport

Energy (or heat) transport in the ocean is much more difficult to determine than in the atmosphere. The most energetic scales of motion in the ocean are smaller than in the atmosphere by an order of magnitude, which requires much denser spatial sampling. Moreover, determining the large-scale flow field is technically much harder in the ocean. As a consequence, evaluation of (2.11) is very hard for the oceanic case. Because of this, people have resorted to indirect methods of estimating ocean heat transport. Five methods are in use today, including the direct one.

1. *Residual calculation*: If the radiation at the top of the atmosphere and the integrated meridional transports in the atmosphere are known, the ocean-atmosphere exchange can be determined as the residual of the atmospheric energy budget (assuming no change in atmospheric heat content). If one furthermore assumes steady state in the ocean, the inferred surface exchange can be integrated to yield ocean heat transport. This is the method used by VonderHaar and Oort (1973) and more recently by Trenberth and Solomon (1994) and Keith (1995).
2. *Bulk formulas*: From ship observations, surface properties of ocean and atmosphere are known. From this, the ocean-atmosphere exchange can be inferred (see Lecture 9), and from this the ocean heat transport. A recent application is Josey et al. (1999).
3. *Direct method*: From hydrographic data and geostrophy, plus information about boundary current velocities and by demanding overall mass conservation, the integral (2.11) can be evaluated. Harry Bryden (SOC) pioneered this approach (Bryden and Hall (1980), Hall and Bryden (1982)).
4. *Inverse method*: Related to the direct method but employing more constraints than overall mass conservation alone. Pioneered by Carl Wunsch (MIT; Roemmich (1980), Roemmich and Wunsch (1985)). Recent applications are by Macdonald and Wunsch (1996) and Ganachaud and Wunsch (2000).
5. *Numerical modelling*: An ocean model driven with atmospheric conditions that are presumed to be known, can be analysed for its simulated heat transport. This approach has, however, not yet reached maturity – if a model disagrees with one of methods 1. – 4., the reason is usually ascribed to poor model performance. A collection of recent results is found in DYNAMO Group (1997).

Exercise

7. *Discuss the strengths and shortcomings of the various approaches.* Hint*: You will have to read the cited literature.*

Figure 2.13 (Fig. 3 of Ganachaud and Wunsch (2000)) shows the most recent and probably best current estimate of global heat transport, based on approach 4. There is a clear indication of an asymmetry with respect to the equator; SH transport is considerably weaker than NH transport. But Fig. 2.13 also gives an impression of the still considerable discrepancy between different estimates. Table 2.1 (Table 5 from Hall and Bryden (1982)) shows, somewhat indirectly, that in the Atlantic at 25ºN the "mean meridional circulation" dominates heat transport. (N.B.: In oceanography, this is usually called the "meridional overturning circulation", MOC, which is preferable because it makes no assumption about it being a time-average.) While total heat transport is about 1.2×10^{15} W northward, the "gyre" contribution is actually southward (mid-ocean temperatures largely higher than those in the Florida Strait). In contrast, the total transport in the upper layers is northward, so warm water moves northward and cold, deep water southward.

Appendix 2.A

Still to be written. However, Chapter 4 of Gill (1982) contains a thorough derivation of energy conservation equations in oceanic and atmospheric dynamics. The interpretation of the $p\alpha$ term is from Warren (1999).

Appendix 2.B

We want to show that, for an ideal gas,

$$B = U + p\alpha + gz + \frac{1}{2}v^2 = c_p\theta + \frac{1}{2}v^2 \qquad (2.16)$$

or, equivalently,

$$U + p\alpha + gz = c_p\theta \ , \qquad (2.17)$$

that is, the sum of enthalpy and potential energy is heat capacity at constant pressure, multiplied by potential temperature. To this end, we first derive some general thermodynamic expressions and then consider the case of an ideal gas.

The first law of thermodynamics states for the heating rate, Q,

$$Q = \frac{dU}{dt}(T,\alpha) + p\frac{d\alpha}{dt} = \frac{\partial U}{\partial T}(T,\alpha)\frac{dT}{dt} + \left(\frac{\partial U}{\partial \alpha}(T,\alpha) + p\right)\frac{d\alpha}{dt}$$
$$= c_v \frac{dT}{dt} + \left(\frac{\partial U}{\partial \alpha}(T,\alpha) + p\right)\frac{d\alpha}{dt} \quad . \tag{2.18}$$

The specific heat at constant volume (that is, at constant α), is

$$c_v \equiv \frac{\partial U}{\partial T}(T,\alpha). \tag{2.19}$$

The specific heat at constant pressure, c_p, is obtained by writing α as a function of temperature and pressure, leading to

$$Q = \left[c_v + \left(\frac{\partial U}{\partial \alpha}(T,\alpha) + p\right)\frac{\partial \alpha}{\partial T}(T,p)\right]\frac{dT}{dt} + \left(\frac{\partial U}{\partial \alpha}(T,\alpha) + p\right)\frac{\partial \alpha}{\partial p}(T,p)\frac{dp}{dt}, \tag{2.20}$$

so that

$$c_p \equiv c_v + \left(\frac{\partial U}{\partial \alpha}(T,\alpha) + p\right)\frac{\partial \alpha}{\partial T}(T,p). \tag{2.21}$$

It is time now to introduce the simplifications offered by the ideal gas law, to prevent the equations from getting more monstrous. First of all, internal energy, U, depends only on temperature and not on volume, so we can write for the internal energy

$$U = c_v T, \tag{2.22}$$

(notice that T denotes absolute temperature here). Second, the ideal gas law states that

$$p\alpha = RT, \tag{2.23}$$

where R is the general gas constant. Equation (2.21) for c_p thus simplifies to

$$c_p = c_v + p\frac{\partial}{\partial T}\frac{RT}{p} = c_v + R, \tag{2.24}$$

and we obtain for the specific enthalpy, H, that

$$H = U + p\alpha = c_p T. \tag{2.25}$$

With the simplifications obtained so far, we can write the Bernoulli function as

$$B = c_p T + gz + \frac{1}{2}v^2. \tag{2.26}$$

We now invoke the first law of thermodynamics, (2.20), again, but using the simplifications arising from the ideal gas law, to obtain

$$Q = c_p \frac{dT}{dt} + p\frac{\partial}{\partial p}\left(\frac{RT}{p}\right)\frac{dp}{dt}$$
$$= c_p \frac{dT}{dt} - \frac{RT}{p}\frac{dp}{dt} \quad . \tag{2.27}$$

Under adiabatic transformations, Q=0; if the motion is assumed purely vertical, we thus have

$$\left(c_p \frac{\partial T}{\partial z}(z,\eta) - \frac{RT}{p}\frac{\partial p}{\partial z}(z,\eta) \right)\frac{dz}{dt} = 0 . \tag{2.28}$$

It follows that the term in brackets is zero; with hydrostatic balance and the ideal gas law we have

$$-\frac{RT}{p}\frac{\partial p}{\partial z}(z,\eta) = -\frac{RT}{p}\left(-\frac{g}{\alpha} \right) = g , \tag{2.29}$$

so that with the definition (2.2) for the adiabatic lapse rate, Γ, we obtain from (2.28),

$$c_p \Gamma = g . \tag{2.30}$$

This means that we can rewrite the Bernoulli function

$$B = c_p T + gz + \frac{1}{2}v^2 = c_p\left(T + \Gamma z\right) + \frac{1}{2}v^2 = c_p\left(T + \int_0^z \Gamma dz'\right) + \frac{1}{2}v^2 , \tag{2.31}$$

so that

$$B = c_p \theta + \frac{1}{2}v^2 , \tag{2.32}$$

which is what we wanted to show. The oceanic case is considerably more difficult, but Warren (1999) showed that $c_p\theta$ approximates the sum of enthalpy and potential energy better than it does enthalpy alone. While Warren (1999) does not appeal to the atmospheric case to make this finding plausible, it does show that $c_p\theta$ is generally an excellent – and eminently simple – expression to use in energy transport calculations.

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